



The structural geology and tectonic history of the northern Flint Creek Range, western Montana
by Jeffrey Frank Baken

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in
Earth Science

Montana State University

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Abstract:

The Flint Creek Range of west-central Montana represents an important and rather atypical segment of the Cordilleran fold and thrust belt. The formation of opposing structural asymmetry in the range has not been satisfactorily explained. Field mapping and structural lab work have been incorporated with modern concepts of fold-and-thrust belt geometry to resolve this question.

The range has been subdivided into three structural provinces based upon distinct structural geometries. The western province, which represents the eastern edge of the Sapphire allochthon, is characterized by "typical" eastward asymmetry; however, the eastern province displays "atypical" westward asymmetry. The northern province is represented by a change in structural trend (convex to the northeast) and opposing fold-plunges where the Flint Creek and Garnet Ranges meet across the Montana lineament.

The kinematic sequence of deformation in the Flint Creek Range incorporates regional tectonic events and features. The Deer Lodge block, which occupies the Belt embayment, moved eastward along a major decollement as the result of combined east-west compression and intrusive wedging by the Idaho batholith (100 m.y. ago). Following development of this allochthon, the Boulder batholith initiated emplacement along the decollement with the Racetrack Creek pluton and the Elkhorn Mountains Volcanics (78-72 m.y. ago) as satellite effects.

The "atypical" geometry of the eastern province was produced by the formation of a ramp interface beneath the Flint Creek Range. A crystalline buttress probably caused the ramp, which created a west-directed back-thrusting and folding event in this area. Then, final emplacement and intrusive wedging by the Idaho batholith created a volume problem in the surrounding area. This problem was solved by creation and movement of the Sapphire tectonic block. This feature thrust along west-dipping thrust faults onto the eastern province (78— 76 m.y. ago). Structure within this terrane represents the western province. The Deer Lodge block then ceased movement and was cut by the final emplacement of the Boulder batholith. (78-72 m.y. ago). The Flint Creek plutons were then emplaced along thrust faults (76.7-72 m.y. ago). Finally, Cenozoic extension produced high-angle normal faults in the region.

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A thesis submitted in partial fulfillment
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of

Master of Science

in

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

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In the summer of 1983, Don Smith was tragically killed in a hit-and-run automobile accident. His influence, advice, and presence are sorely missed.

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ABSTRACT

The Flint Creek Range of west-central Montana represents an important and rather atypical segment of the Cordilleran fold and thrust belt. The formation of opposing structural asymmetry in the range has not been satisfactorily explained. Field mapping and structural lab work have been incorporated with modern concepts of fold-and-thrust belt geometry to resolve this question.

The range has been subdivided into three structural provinces based upon distinct structural geometries. The western province, which represents the eastern edge of the Sapphire allochthon, is characterized by "typical" eastward asymmetry; however, the eastern province displays "atypical" westward asymmetry. The northern province is represented by a change in structural trend (convex to the northeast) and opposing fold-plunges where the Flint Creek and Garnet Ranges meet across the Montana lineament.

The kinematic sequence of deformation in the Flint Creek Range incorporates regional tectonic events and features. The Deer Lodge block, which occupies the Belt embayment, moved eastward along a major decollement as the result of combined east-west compression and intrusive wedging by the Idaho batholith (100 m.y. ago). Following development of this allochthon, the Boulder batholith initiated emplacement along the decollement with the Racetrack Creek pluton and the Elkhorn Mountains Volcanics (78-72 m.y. ago) as satellite effects. The "atypical" geometry of the eastern province was produced by the formation of a ramp interface beneath the Flint Creek Range. A crystalline buttress probably caused the ramp, which created a west-directed back-thrusting and folding event in this area. Then, final emplacement and intrusive wedging by the Idaho batholith created a volume problem in the surrounding area. This problem was solved by creation and movement of the Sapphire tectonic block. This feature thrust along west-dipping thrust faults onto the eastern province (78-76 m.y. ago). Structure within this terrane represents the western province. The Deer Lodge block then ceased movement and was cut by the final emplacement of the Boulder batholith (78-72 m.y. ago). The Flint Creek plutons were then emplaced along thrust faults (76.7-72 m.y. ago). Finally, Cenozoic extension produced high-angle normal faults in the region.

INTRODUCTION

The Flint Creek Range was thoroughly mapped in a series of Princeton University dissertations in the late 1950's and early 1960's. The publication of these papers marked a point at which the basic geology on the Flint Creek Range had been mapped and served to expound and enlarge upon the work of earlier geologists (see Previous Investigations).

Purpose of Investigation

The purpose of this study is to answer one of the problems raised in the Princeton dissertations. In 1958, McGill recognized a distinct "structural paradox" (p. 99). Part of the study area, located in the eastern half of the Flint Creek Range displays "atypical" asymmetry and is characterized by north-south trending structures in the form of asymmetric folds with westward vergence; high angle, east-dipping normal faults down-dropped on the east; and east-dipping thrust faults. These structures appear quite anomalous when compared with the more-typical eastward-verging structural style displayed throughout the Rocky Mountain fold and thrust belt and in the immediately adjacent, western part of the range.

This study reveals various possible tectonic influences which may have acted to produce the "atypical" structure of the eastern province. This is accomplished by a structural synthesis of the range

using up-to-date, fold and thrust belt concepts and geometry. Recently published data incorporated into the study includes new theories on the tectonics of the area, the ages and character of the various plutons surrounding the range, and stratigraphic observations comparing the eastern-province rocks with those to the west. The model presented here is new and is inconsistent with others up to this date. The problem is approached from a regional perspective in an attempt to relate the surrounding, regional tectonic features to the origin of the various structural styles displayed in the Flint Creek Range.

Field and Lab Procedures

Approximately twelve weeks were spent in the field during the summer of 1980 and the spring of 1981 measuring structural attitudes. This information was plotted on 7 1/2 and 15 minute series topographic sheets, aerial photographs (1:37,400), and a base map (1:100,000). For simplicity and clarity, the rocks were combined into Precambrian, Paleozoic, Mesozoic, and Cenozoic units on the map.

Lab work was completed during the winter and spring of 1981 and the winter of 1982. This involved the development of a composite geologic map (1:100,000 scale) utilizing previous mapping as well as the information compiled by the author (Plate 1). Several cross sections were completed and the structural style was summarized in pi diagrams representing poles-to-bedding.

Location and Geography

The Flint Creek Range is located in west-central Montana between Butte and Missoula (Fig. 1). Communities surrounding the area are Drummond on the north, Deer Lodge on the east, Anaconda on the southeast, and Philipsburg on the west (Fig. 2).

The range lies within portions of Deer Lodge, Granite, and Powell counties. The range is fifteen miles wide and extends thirty miles south of U.S. Interstate 90 from the area east of Drummond.

Peaks in the range reach the 10,164 foot-level (Mt. Powell) in the eastern half and an average height of 8,000 feet in the western half (Fig. 2). Major streams incise the range with a NW-SE orientation on the west side and a NE-SW orientation on the east side.

The north end of the Flint Creek Range is terminated by the Clark Fork River Valley, and the eastern and western margins are marked by the Deer Lodge and Flint Creek Valleys, respectively. The southern margin is marked in the west by Georgetown Lake and the Flint Creek drainage and in the east by the Warm Springs Creek drainage.

Mountain ranges surrounding the Flint Creek Range are the Sapphire and Bitterroot Ranges to the west, the Garnet Range to the north and the Anaconda and Pintlar Ranges to the south.

The land within the range is primarily public land controlled by the U.S. Forest Service and major blocks surrounding the area are controlled by the Bureau of Land Management, the state of Montana, and private individuals. Numerous small mining claims dot the area.

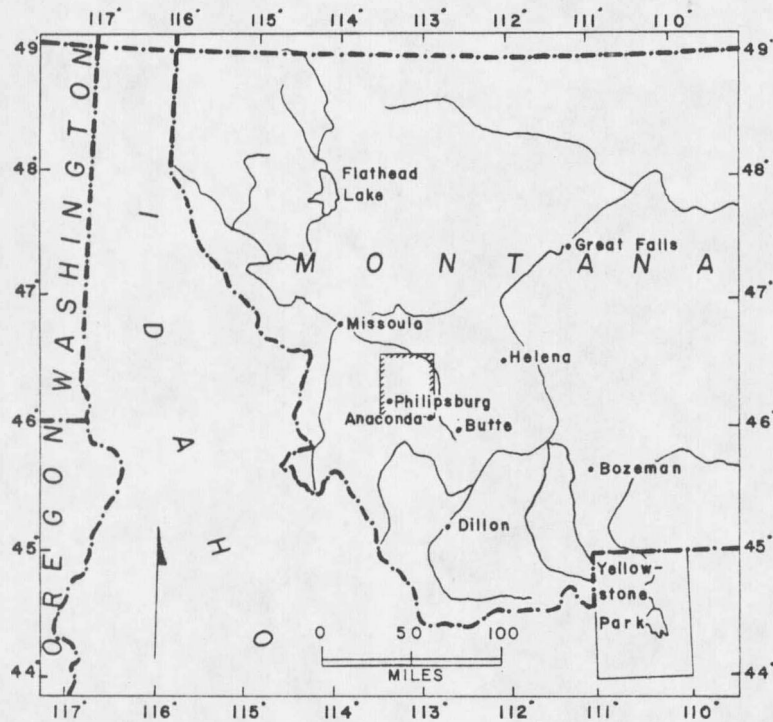


Figure 1. Location map. The study area is outlined.

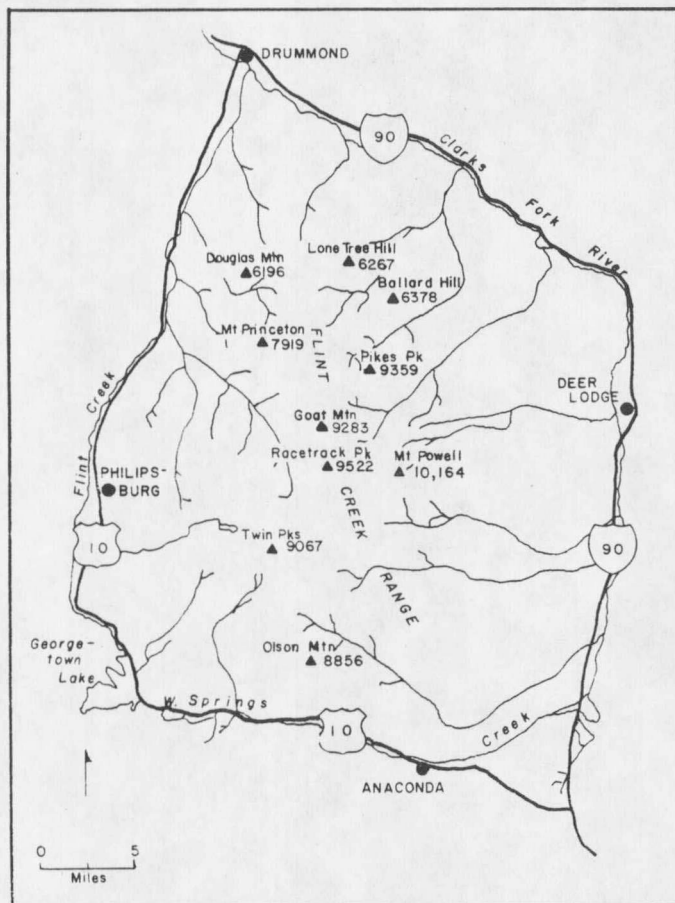


Figure 2. Highway and drainage map. The Flint Creek Range is bordered by major highways on all sides. Note the northeast-trending drainage in the east and the northwest-trending streams in the west.

Access within the study area is excellent with a road in nearly every major drainage and a network of mining and timber roads.

Previous Investigations

During the period from 1907-1915, the Flint Creek Range was mapped and described by Emmons and Calkins (Emmons, 1907; Emmons and Calkins, 1913; and Calkins and Emmons, 1915). Their 1913 publication remains an outstanding example of early American geologic literature.

The initial work by Emmons and Calkins was followed by a period of mineral exploration and several consequent publications dealing with mineral deposits and their descriptions (Pardee, 1936; Goddard, 1940; and Popoff, 1953).

The Flint Creek Range was extensively remapped from 1957 to 1963 in a series of dissertations by students from Princeton University (Poulter, 1957; McGill, 1958; Mutch, 1960; Gwinn, 1960; and Csejtey, 1963). Subsequent work by these authors produced several follow-up publications dealing with specific problems (Mutch and Gwinn, 1960) or synthesis of their work into the regional tectonic framework (Poulter, 1959; and McGill, 1965).

The 16th Annual Field Conference Guidebook published by the Billings Geological Society in 1965 dealt specifically with the geology of the Flint Creek Range (Fields and Shepard, 1965). This publication included a series of papers (including McGill, 1965) covering a wide range of topics.

In the late 1960's and early 1970's several miscellaneous studies were published which dealt with mines, ore deposits, general geology,

and igneous petrology (Allen, 1966; Wanek and Barclay, 1966; Prinz, 1967; Earll, 1972; Hyndman and others, 1972; and Hawley, 1974). The author views this period as the end of a long time interval during which the pioneering geology was completed.

The remaining investigations published on the Flint Creek Range are directed toward an analysis of the development of the range and its relationships to the surrounding regional tectonic elements (Hyndman, et al, 1975; Hyndman, Talbot, and Chase, 1975; Hyndman, et al, 1976; Hyndman, 1979 and 1980; Ruppel, et al, 1981; Baken, 1981; and Winston, et al, 1982).

STRUCTURE

Introduction

The Flint Creek Range is surrounded by several important regional tectonic features including two batholiths (Fig. 3) and is located within the central Montana salient of the Cordilleran fold and thrust belt (Fig. 4).

The range contains structural and petrologic characteristics which make it difficult to assign it to the "suprastructural" or "infrastructural" subdivisions of the Cordillera. Generally, the range is characterized by large-scale thrust faults, major decollements, minor tear faults, numerous listric normal faults, and concentric folds, all of which belong to the "suprastructure" family of structures (Dahlstrom, 1970; and Royse, et al, 1975). However, the area also contains syntectonic granitic intrusive bodies, minor contact metamorphic rocks, and fracture cleavage characteristic of the "infrastructure" or western hinterland. Furthermore, the eastern part of the Flint Creek Range displays westward vergence which is "atypical" of the Cordilleran fold and thrust belt.

Consequently, the origin of the range represents an important phase in the tectonic history of the western Montana fold and thrust belt due to its unique and anomalous structural character. The kinematic interpretation presented here hopefully will stimulate new thought on the tectonic development of the region.

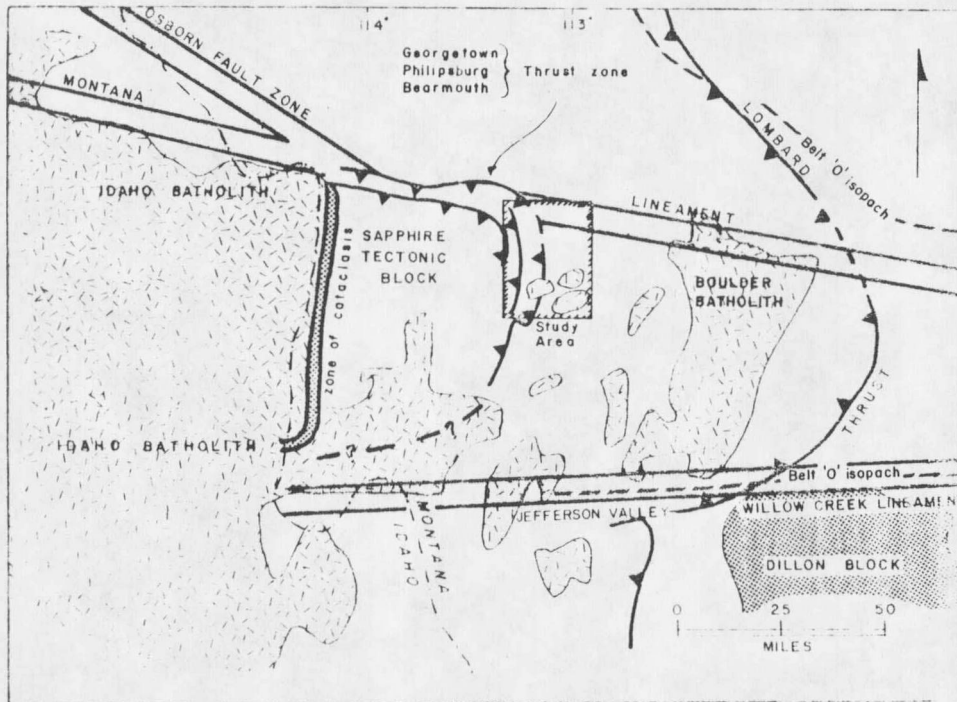


Figure 3. Regional tectonic features. The Flint Creek Range is surrounded by major tectonic features and intrusive bodies. (Modified from Csejtey, 1963).

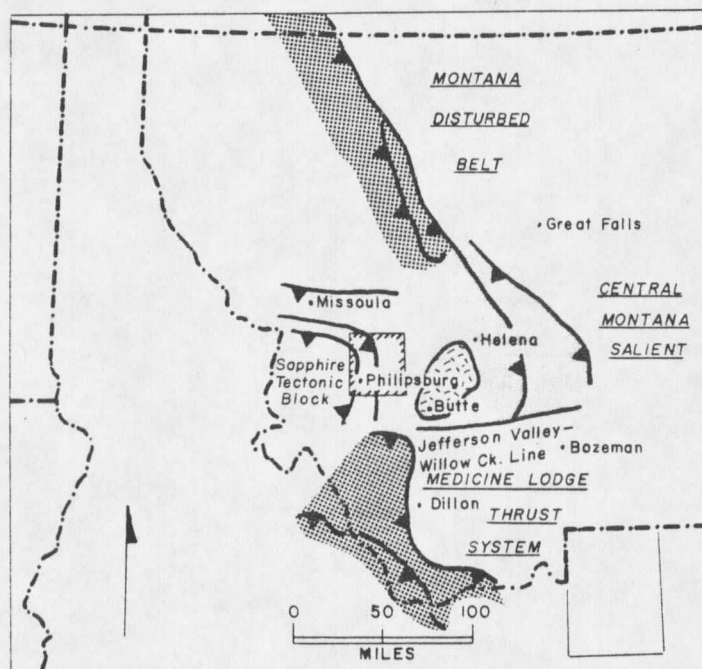


Figure 4. Segments of the Cordilleran fold and thrust belt, western Montana. The study area is outlined.

Local Structural Geology

Introduction

Mapping and gathering of structural data was completed in the northern Flint Creek Range in T. 8 and 9 N., and R. 12 and 13 W. (Fig. 5 and Plate 1). Detailed descriptions of only those features appearing in this area will be presented here. The reader is referred to Emmons and Calkins (1913), Poulter (1957), McGill (1958), Gwinn (1960), Mutch (1960), Csejtey (1963), Wanek and Barclay (1966), Earll (1972), and Wallace, et al (1978) for information on structural features in the remainder of the Flint Creek Range.

Three structural styles delineate distinct structural provinces in the range (Fig. 5). The western and eastern provinces are characterized by opposing eastward and westward asymmetry, respectively. The northern province is characterized by a sharp, structural trend change and opposing plunges where the folds of the Flint Creek and Garnet Ranges meet across the Montana lineament ["Clark Fork Sag" of Gwinn (1960)].

Western Province

Introduction. The western province is characterized by eastward asymmetry displayed in broad, relatively undeformed folds of Precambrian and Paleozoic allochthonous rocks (Figs. 6 and 7). The eastern margin is marked by the trace of the Georgetown-Princeton thrust fault (Fig. 5 and Plate 1).

Contoured pi diagrams compiled from bedding attitude data demonstrate the eastward vergence of folds. Figures 10 and 11 are

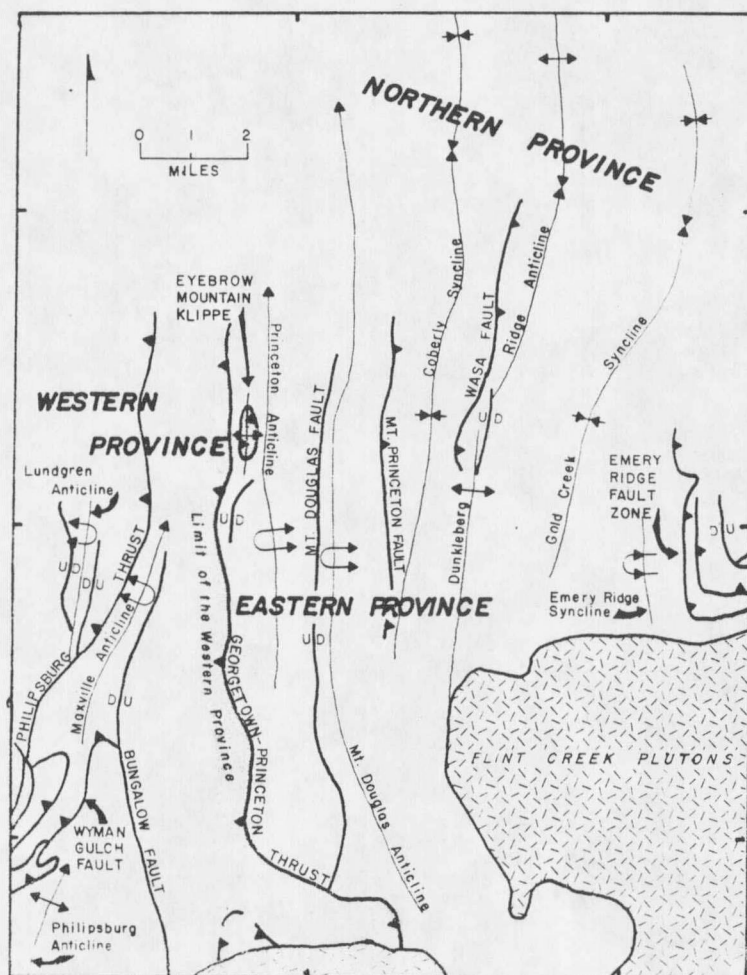


Figure 5. Index to structural provinces and features. Three distinct structural geometries delineate the three structural provinces.

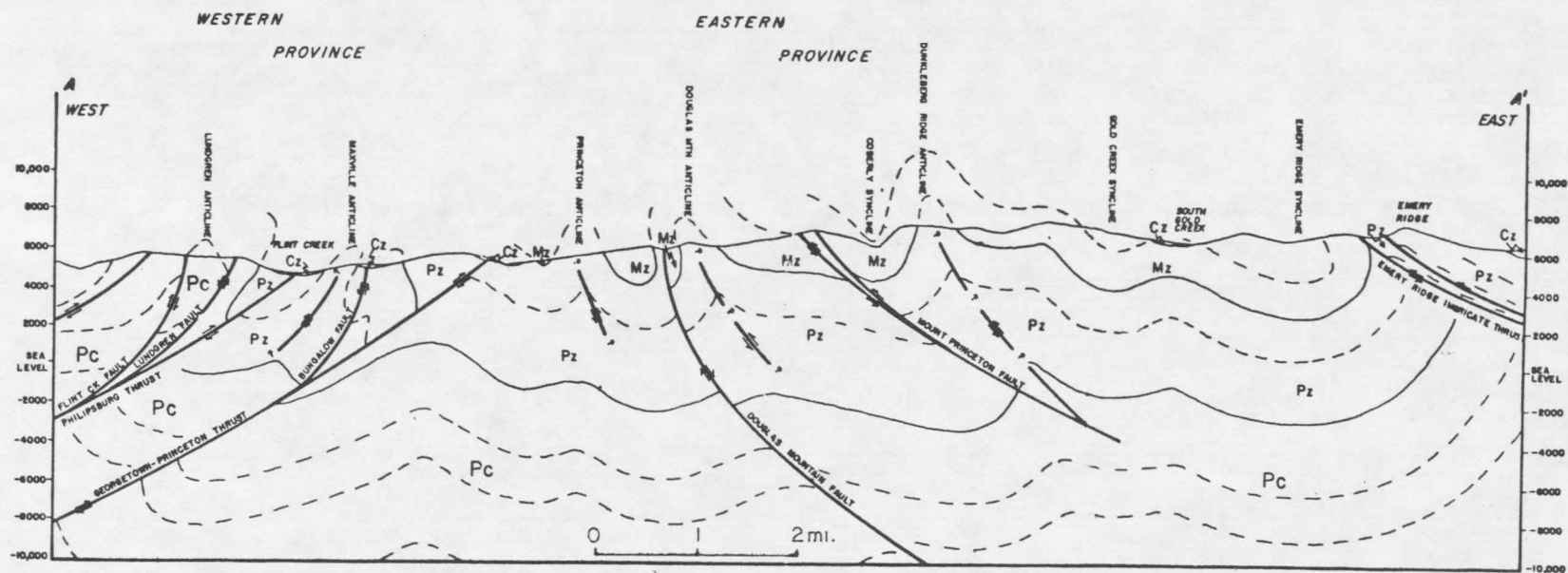


Figure 6. Cross section AA'. This east-west section shows the major structures of the western and eastern provinces. Note the opposing symmetry. The horizontal and vertical scales are the same.

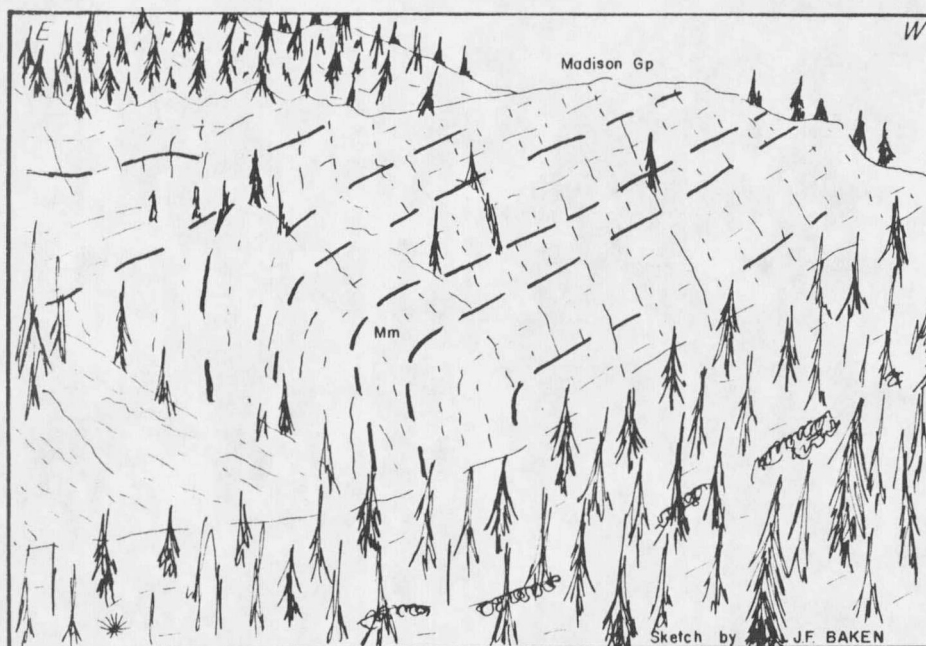


Figure 7. Fold in the Madison Formation. This south view shows the eastward vergence characteristic of the western province.

contour diagrams of randomly compiled poles to bedding for the Maxville and Lundgren anticlines, respectively. These diagrams clearly show the eastward asymmetry of these two folds.

The western province is further characterized by large-scale, north-south trending thrust faults; high-angle, north-south trending normal and reverse faults; and a klippe in which the allochthonous western province rocks overlie eastern province strata.

Faults. Georgetown-Princeton Thrust. The Georgetown-Princeton thrust fault, as mapped for this report, is a combination of Poulter's (1957) Georgetown thrust and McGill's (1958) eastern Philipsburg thrust.

McGill (1958) postulated that the Philipsburg thrust is exposed by two fault traces (represented on Plate 1 by the Philipsburg and Princeton thrusts) the result of the erosion of a single folded thrust. Figure 6 displays the interpretation presented here; that is, the presence of two distinct thrust faults. The western segment will still be referred to as the Philipsburg thrust and the eastern segment is here referred to as the Princeton thrust; that is, the northern portion of the combined Georgetown-Princeton thrust system.

The Princeton thrust lies east of the Philipsburg thrust and is parallel to it (Fig. 6). This interpretation is based on a projection of the Georgetown thrust (Poulter, 1957) northward across the Philipsburg batholith, connecting with the Princeton thrust (Fig. 5 and Plate 1). The Princeton part of the fault was the only feature mapped by the author. This segment of the Georgetown-Princeton thrust

is a north-south trending fault with a minor east-west trending segment, which is best exposed on the south side of Boulder Creek in Sec. 23, T. 8 N., R. 13 W., and at Eyebrow Mountain in Secs. 25 and 26, T. 9 N., R. 13 W. (Fig. 8). The fault which is typically covered by thick forest places Precambrian and Lower Paleozoic rocks on Upper Paleozoic and Lower Mesozoic rocks to the east and displays a moderate to subhorizontal dip to the west.

Eyebrow Mountain is a klippe which is an eastward extension of the Princeton thrust (Figs. 5 and 8; and Plate 1). This feature is an anticline involving Quadrant and Madison rocks in fault contact with an underlying syncline developed in Kootenai rocks. The fault underlying the klippe is subhorizontal.

North of the Philipsburg batholith the Princeton thrust makes a marked trend change to the west. This trend change indicates the possibility of forceful intrusion of the Philipsburg batholith after thrusting. The outcrop pattern of the fault may be due to northward tilting of the rocks on the northern flank of the batholith. This is supported by the exposure of underlying Precambrian rocks which contain a north-dipping diabase sill which is parallel to bedding in the Precambrian strata.

The Princeton fault trends northward beneath Cenozoic strata of Flint Creek and the Clark Fork River. The fault is assumed to change trend to the northwest joining east-west trending strike-slip faults along the Montana lineament [which mark the northern boundary of the Sapphire tectonic block (Fig. 3)]. The Georgetown-Princeton thrust is important in a regional sense because it represents the leading edge

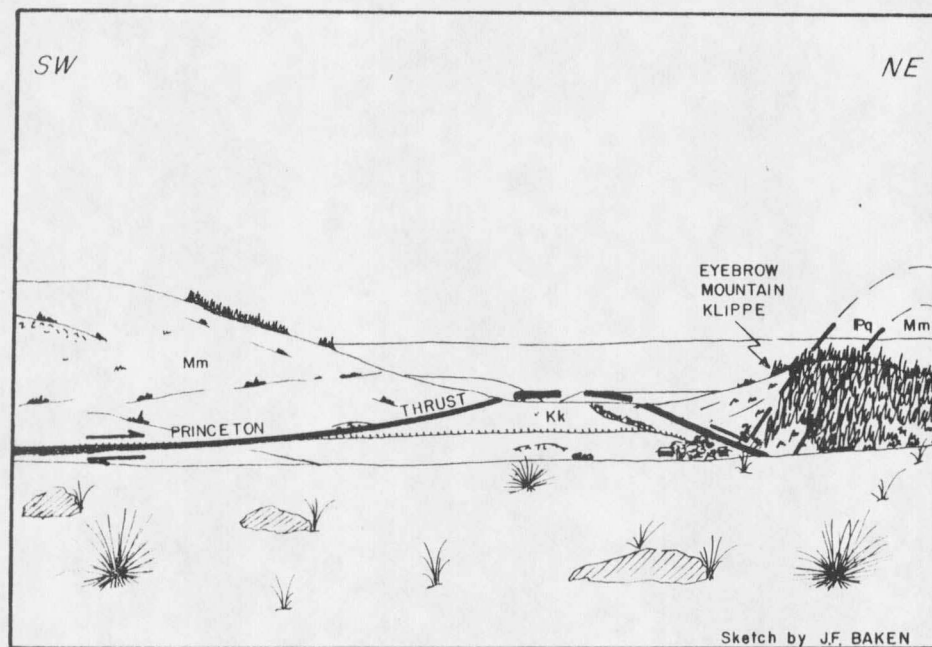


Figure 8. Princeton thrust fault and the Eyebrow Mountain klippe. This portion of the thrust is subhorizontal.

or eastern "toe" of the Sapphire allochthon (Hyndman, et al, 1975, 1979).

Philipsburg Thrust. The Philipsburg thrust was first identified by Emmons and Calkins (1913). This north-trending, west-dipping fault is not well exposed in the map area and is generally depicted by the trend of the Flint Creek Valley (Figs. 5 and 9; and Plate 1). The thrust is marked by the occurrence of Precambrian and Lower Paleozoic rocks immediately adjacent to Middle Paleozoic rocks across the Flint Creek Valley in the western half of T. 8 N., R. 13 W.

The dip of the fault surface is subject to conjecture. However, the angle is probably low to moderate as evidenced by its meandering map pattern. The fault surface is not exposed in the northern Flint Creek Range. However, McGill (1958) interpreted its appearance in a complex fault zone in the southwest quarter of T. 8 N., R. 13 W. (Plate 1), where a low-angle thrust is interrupted by two high-angle reverse faults in Secs. 30 and 31. McGill defined this low-angle thrust, which places Cambrian Hasmark on Mississippian Madison, as a segment of the Philipsburg thrust. Hyndman, et al (1975, 1979) believed movement on the thrust occurred over a distance of 60 to 65 kilometers.

The attitude, location, and trend of the thrust is unknown to the south and north of the Flint Creek Range (Fig. 9). However, Poulter (1957) and Flood (1974) believed that the fault dies out to the south in a zone of ductile deformation and broad, open folds in Precambrian strata. Hyndman, et al (1975) interpreted that the trend of the

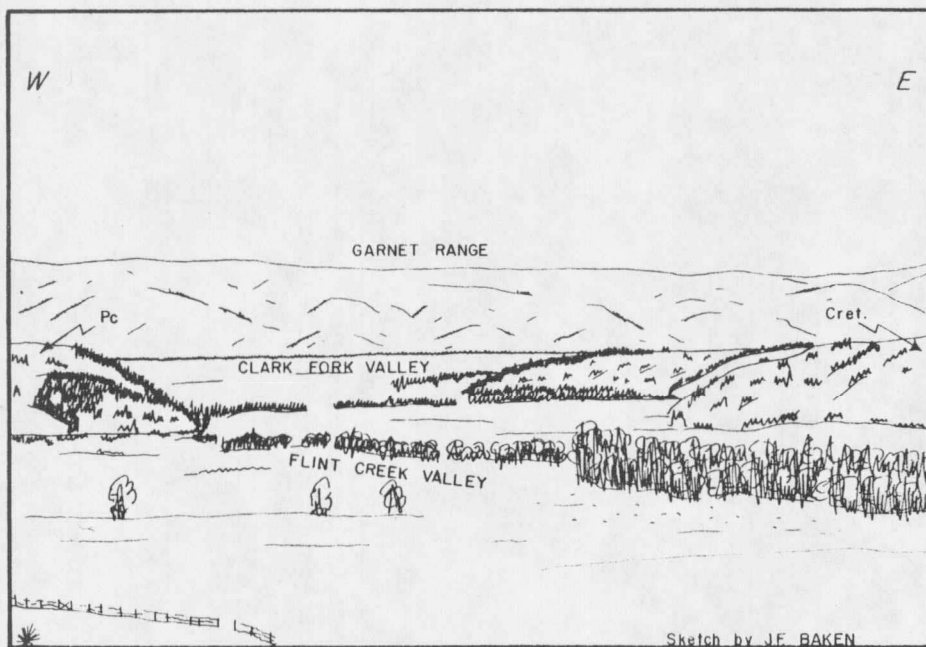


Figure 9. Northern Flint Creek Valley. The Flint Creek sediments cover the Philipsburg thrust which places Precambrian rocks (on the left) adjacent to Cretaceous rocks (on the right). The thrust probably changes trend here to the northwest behind the Precambrian hill on the left.

strike-slip faults at the northern boundary of the Sapphire tectonic block (Fig. 3) changes eastward to a south trend, eventually connecting with the Philipsburg and Georgetown-Princeton thrusts (Figs. 3 and 5).

Wyman Gulch Thrust. The Wyman Gulch thrust is best expressed in the southwest quarter of T. 8 N., R. 13 W., where it trends in a southwest direction into the northwest quarter of T. 7 N., R. 13 W. (Fig. 5 and Plate 1). This low-angle fault is not well exposed but appears to place the Madison Formation over the Cretaceous Kootenai Formation. Stratigraphic throw is estimated to be 1,300 feet.

The Wyman Gulch fault is on the east limb of the north-south trending, overturned Maxville anticline, and bifurcates to the southwest into a complex pattern of faults and folds on the west limb of the Philipsburg anticline. The fault is terminated to the north by the northwest-trending Bungalow fault (Fig. 5 and Plate 1).

The Wyman Gulch thrust is apparently an imbricate splay of the Philipsburg thrust in the southwest quarter of T. 8 N., R. 13 W., described previously.

Bungalow Fault. The Bungalow Fault is a high-angle, subvertical normal fault which trends north-south through the middle of T. 8 N., R. 13 W. (Fig. 5 and Plate 1). This fault is well exposed and is marked by a soil color change from gray and tan of the down-dropped Madison Formation and Mesozoic rocks on the west, to red Missoula Group rocks on the east. Although exposures are relatively good, no fault plane was observed.

Cross section AA' (Fig. 6) shows the Bungalow fault as a listric normal fault "soling out" at depth into the Georgetown-Princeton thrust. To the south, the fault displaces the eastern limb of the Philipsburg anticline which, in this area (north half of T. 7 N., R. 13 W.), has apparently been deformed by forceful intrusion of the Philipsburg batholith. This relationship dates the latest movement of the Bungalow fault as post-folding. Hyndman, et al (1972) also believed the batholith was post-compressional.

Folds. Maxville Anticline. The Maxville anticline is an overturned fold in the western half of T. 8 N., R. 13 W. (Fig. 5 and Plate 1). This indistinct structure trends an average 011' (Fig. 10) and crops out as a high, tree-covered ridge of Madison Limestone.

The contoured pi diagram of poles to bedding (Fig. 10) indicates a fold-axis plunge of 20' toward 011' and shows the eastward asymmetric character of the western province. That is, the majority of the poles to bedding occur in the southeastern quadrant of the diagram as evidence of consistent, west-dipping beds.

The Maxville anticline is bounded on the west and south by the Philipsburg and Wyman Gulch thrusts and in the east by the Bungalow fault (Figs. 5 and 6; and Plate 1).

Although the Maxville anticline is poorly exposed, McGill (1958) mapped a small nose of chert beds in Secs. 19, 20, and 30, T. 8 N., R. 13 W. (Plate 1) indicating overturning of the east limb. In addition, published thicknesses of the Madison (see Appendix) indicate at least a doubling of the formation in this area. The author does not rule

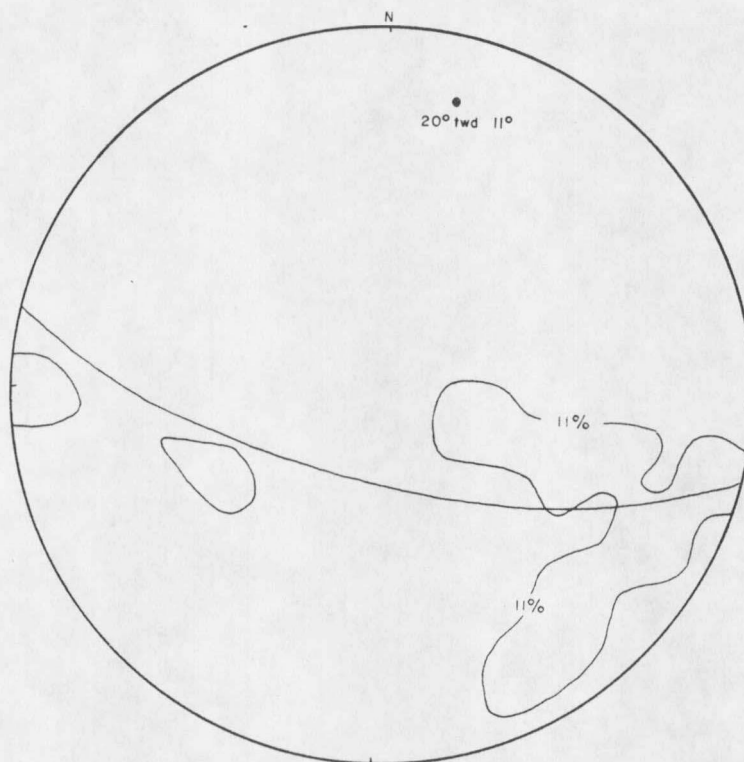


Figure 10. Contoured pi diagram of poles to bedding for the Maxville anticline. The concentration of poles in the southeast quadrant demonstrates the eastward vergence. ("N" is north.)

out the possibility of minor reverse faulting in the core of the fold to assist in creating this thickened section.

Lundgren Anticline. The Lundgren anticline is located in the east half of Secs. 5 and 8, T. 8 N., R. 13 W., and displays a partially covered nose in Sec. 32, T. 9 N., R. 13 W. (Fig. 5 and Plate 1). The fold is overturned to the east involving moderately resistant, grass-covered ridges of Precambrian Belt and basal Cambrian rocks.

Figure 11 is the contoured pi diagram of poles-to-bedding for the Lundgren anticline. The diagram indicates a fold axis plunge of 25' toward 024' with strong eastward vergence and local overturning.

The Lundgren anticline is similar to the Maxville anticline in two respects: 1) the fold is fault-bounded on the east and west by two north-south trending faults (Figs. 5 and 6; and Plate 1), and 2) the anticlinal nature of the fold is difficult to determine. This latter problem is reduced by the presence of an anticlinal nose in the southeast quarter of Sec. 32, T. 9 N., R. 13 W. (Plate 1). This nose is evidenced by a change in strike which trends beneath adjacent glacial moraine and reemerges to connect with the opposite limb in Sec. 5, T. 8 N., R. 13 W.

Philipsburg Anticline. The Philipsburg anticline is on the western edge of T. 7 N., R. 13 W. (Fig. 5 and Plate 1). This fold is extensively tree-covered, however, a prominent ridge of Quadrant Quartzite in Secs. 7, 8, and 9 provides a very clear anticlinal nose.

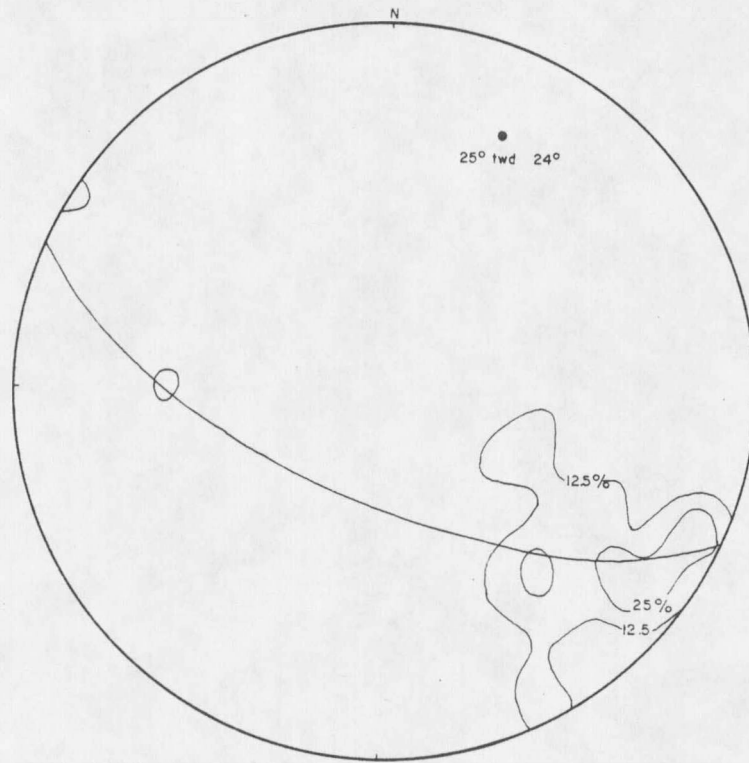


Figure 11. Contoured pi diagram of poles to bedding for the Lundgren anticline. Eastward vergence is demonstrated. ("N" is north.)

Rocks expressed in the core of the anticline include the Precambrian Belt to the Cretaceous Colorado Group.

All of the mines in the Philipsburg mining district occur on the crest of the Philipsburg anticline in the Cambrian Hasmark Formation. Earl (1972, p. 11), referring to the southern Flint Creek Range, noted that:

"There is some spatial evidence to suggest that folding has exercised some subtle control over mineralization, although localization is probably determined to at least as great a degree by favorable lithology."

The contoured pi diagram of the Philipsburg anticline (Fig. 12) shows a fold axis plunge of 20' toward 013'. This diagram does not display the typical western-province distribution of poles to bedding concentrated in the southeast quadrant. This is possibly the effect of forceful intrusion of the adjacent Philipsburg batholith immediately to the east (Fig. 13). Additional evidence of this deformation can be seen in the northern half of the township (Plate 1), where the southwestern trend of the eastern limb has been diverted to an east and northeast direction.

The Philipsburg anticline, like the Maxville and Lundgren anticlines, is fault-bounded on the east and west sides (Fig. 5). However, the Philipsburg anticline is also bounded on the southeast and south by the Philipsburg batholith.

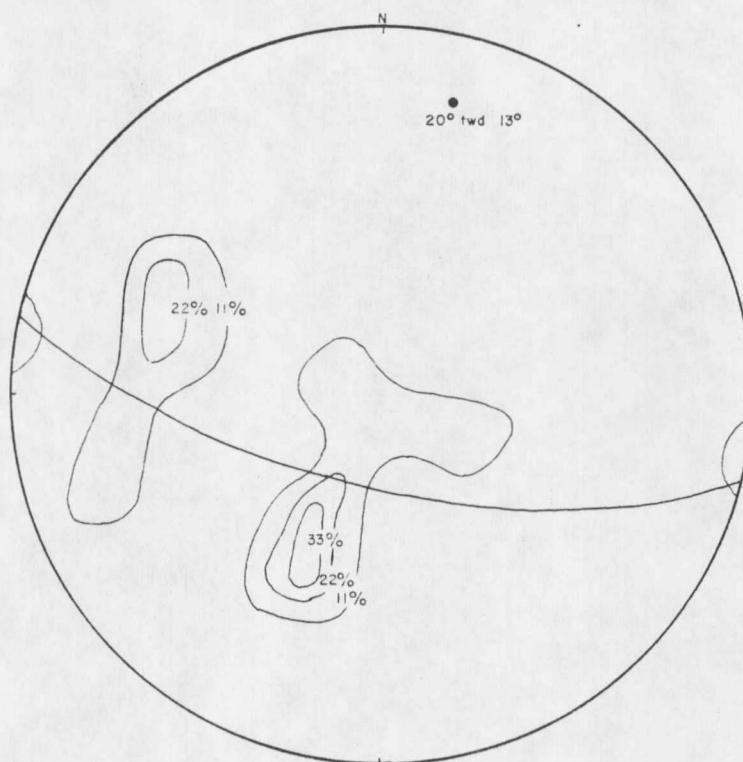


Figure 12. Contoured pi diagram of poles to bedding for the Philipsburg anticline. The fold displays slight westward asymmetry. Probably forceful intrusion of the Philipsburg batholith changed the symmetry of this feature (See Fig. 13). ("N" is north.)

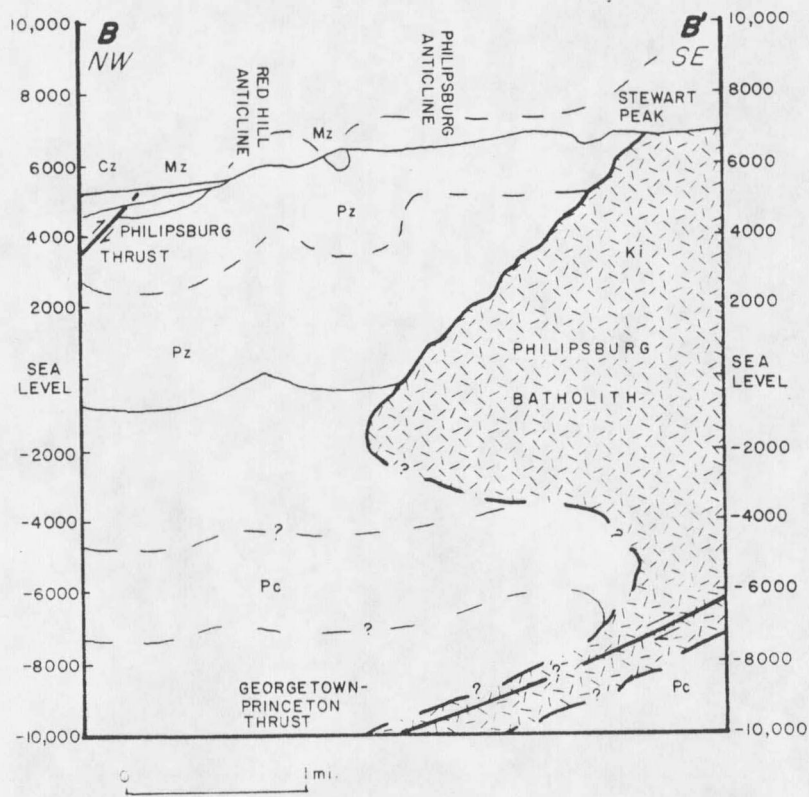


Figure 13. Cross section BB'. The Philipsburg batholith probably intruded along the Georgetown-Princeton thrust slightly deforming the preexisting folds.

Eastern Province

Introduction. The eastern structural province occupies the north and northeastern parts of the map area. This province lies south of the Clark Fork Valley and northeast of the Flint Creek plutons (Fig. 5 and Plate 1). Furthermore, the eastern province is bound on the east and north by Cenozoic strata of the Deer Lodge and Clark Fork Valleys, respectively. The western margin is represented by the trace of the Georgetown-Princeton thrust, and the southern boundary is marked by the Flint Creek plutons.

The structure of the eastern province is characterized by Paleozoic and Mesozoic autochthonous rocks (relative to the western province) containing tightly-folded, north-south trending folds that display east-dipping axial planes and westward overturning (Fig. 6). The faults are north-south trending thrusts and moderately high-angle reverse and normal faults that all dip east (Fig. 17), and several high-angle, northeast-trending faults which are covered or obscure (Fig. 5 and Plate 1). These structural features combine to produce an overall westward-verging character to the province.

Contoured pi diagrams clearly show the dominance of east-dipping limbs and westward vergence (Figs. 22 and 24).

The eastern structural province represents the area of "atypical" structure mentioned in the Introduction. It is common knowledge that "typical" asymmetry in the Rocky Mountain fold and thrust belt is eastward, in the direction of overall tectonic transport. Therefore, the eastern province represents an interesting and important

structural anomaly in this region because it suggests east-to-west tectonic transport.

Faults. Douglas Mountain Fault. The Douglas Mountain fault is exposed in the canyon walls of Douglas Creek in Sec. 31, T. 9 N., R. 12 W. (Fig. 14). This north-south trending structure crops out on the west side of the Douglas Mountain anticline involving resistant ridges of Quadrant Quartzite adjacent to nonresistant Permian rocks.

This fault is economically important in the Douglas Creek area where the Permian Phosphoria Formation is exposed on the upthrown, western side of the fault providing easy access to the phosphate beds. An adit clearly marks the location of the phosphatic beds in the north wall of Douglas Creek Canyon.

The Douglas Mountain fault was interpreted by McGill (1958 and 1959) as a reverse fault. He suggested the fault is a west-dipping, back-limb thrust on the Douglas Mountain anticline genetically related to thrusting from the west. McGill proposed that after faulting, both the west-dipping, western limb of the fold and the fault had been rotated to an east dip by continued east-directed stress at depth. Later, McGill (1965) expressed doubt concerning this interpretation due to evidence provided by Gwinn (1961) and Mutch (1961). These authors demonstrated the existence of other normal faults and noted that the Douglas Mountain fault cuts the upright, west-dipping, western limb of the anticline to the north (Fig. 6 and Plate 1).

The Douglas Mountain fault plane is not well exposed in the map area. However, Gwinn (1961) suggested the dip of the fault plane is

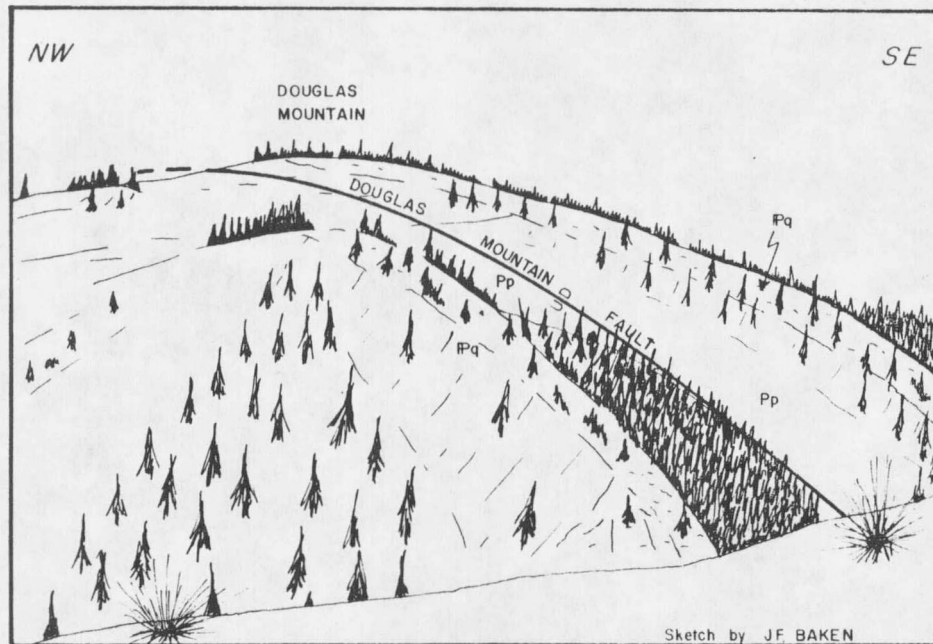


Figure 14. The Douglas Mountain fault on Douglas Mountain. This north-south trending normal fault was probably created by extension after folding and thrusting.

in the range of 55-85° east, but this is not confirmed by the author. Stratigraphic throw on the fault in the Douglas Creek area is estimated to be 900 feet.

The Douglas Mountain fault is one of two high-angle, north-south trending faults which are down-dropped to the east. Therefore, the fault formed after the suggested east-to-west tectonic transport in this province. Additionally, the fault cuts the Douglas Mountain anticline which was created during west-directed transport.

Mount Princeton Fault. The Mount Princeton fault is actually a zone of several north-south trending, east-dipping, small-scale thrust and moderately high-angle reverse faults that repeat the Taft Hill and Flood Members of the Cretaceous Blackleaf Formation of the Colorado Group and the "Gastropod" Member of the Kootenai Formation. This fault zone is best exposed in the eastern half of Sec. 17, T. 9 N., R. 12 W. (Fig. 5 and Plate 1), where the zone involves several repeated outcrops of the Taft Hill and Flood Members. Though the fault plane is not well exposed, there are two poor outcrops in Secs. 32 and 33, T. 9 N., R. 12 W., which display microstructures within the Taft Hill Member (Figs. 15 and 16).

McGill (1958) mapped the Mount Princeton fault as a normal fault, down-dropped on the east. However, more detailed mapping by Gwinn (1960) reinterpreted it as an east-dipping thrust zone with relative movement from east to west.

The Mount Princeton fault cuts across different stratigraphic horizons along its length. In Sec. 17, the fault is mapped by a repetition of the Taft Hill Member in fault contact with the Flood.

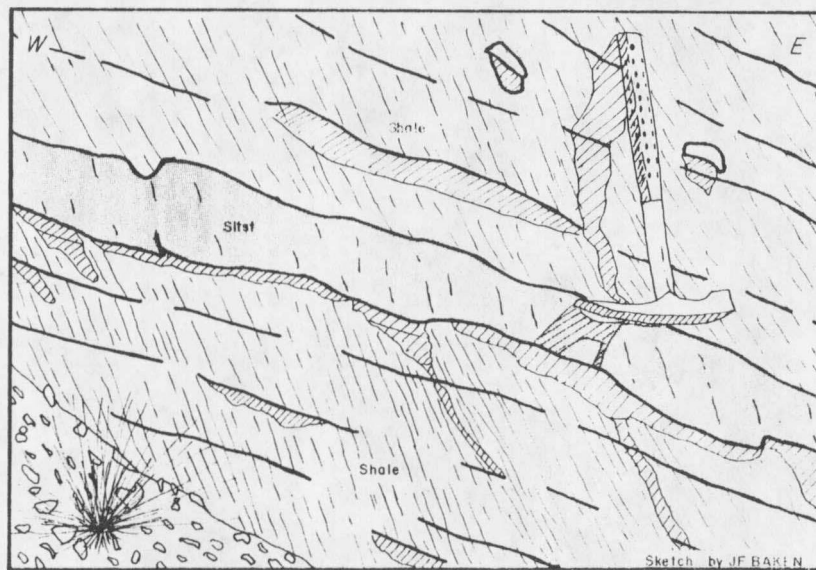


Figure 15. Microcleavage in the Taft Hill Member near the Mount Princeton fault. The cleavage has formed in shale--the result of a probable force couple produced by westward movement on the fault (See Figures 16 and 17).

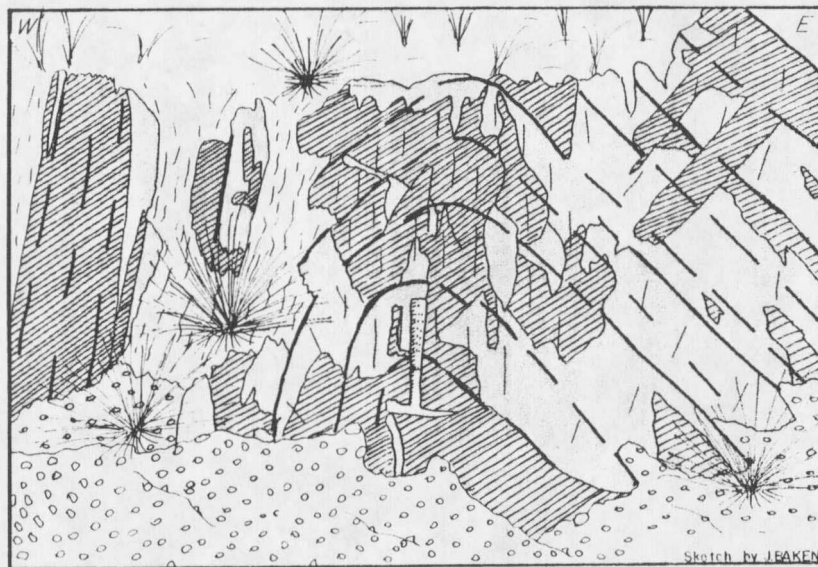


Figure 16. Small fold in the Taft Hill Member near the Mount Princeton fault. The westward asymmetric fold was probably caused by westward movement along the fault (See Figures 15 and 17).

In Sec. 32, the fault places the "Gastropod" Member of the Kootenai Formation on the Flood Member. Estimated stratigraphic throw is 100-300 feet.

The approximate fault-plane dip was measured to be 70' east on the Middle Fork of Douglas Creek; however, the dip along the northwest part of the fault appears to be much more gentle.

Figure 15 shows microstructures in the Taft Hill Member. The attitude of the fracture cleavage is 345° , 55-60' N.E., and may have formed due to the proximity of a fault splay. It is interesting to note that the fracture cleavage is prominently developed in the shale, while the siltstone appears to be relatively unaffected. Figure 17 places Fig. 15 into perspective with an overall view of the outcrop. Here, a minor thrust fault splay displaces rock from east-to-west over the siltstone-shale sequence seen in Fig. 15. The author proposes that the fault created a force couple with the overlying rock which is evidenced by the fracture cleavage.

The fold illustrated on the left side of Fig. 17 is seen in detail in Fig. 16. This minor feature also appears to be genetically related to the small fault splay just discussed. The fold displays obvious westward asymmetry as well as minor fracture cleavage with an orientation of 355° , 70-80° N.E. The fracture cleavage seems to have been caused by interbed slippage during folding.

The microstructures displayed in Figs. 15, 16, and 17 suggest a genetic relationship with west-directed movement in the eastern province. These features are suggested as field evidence for proposed

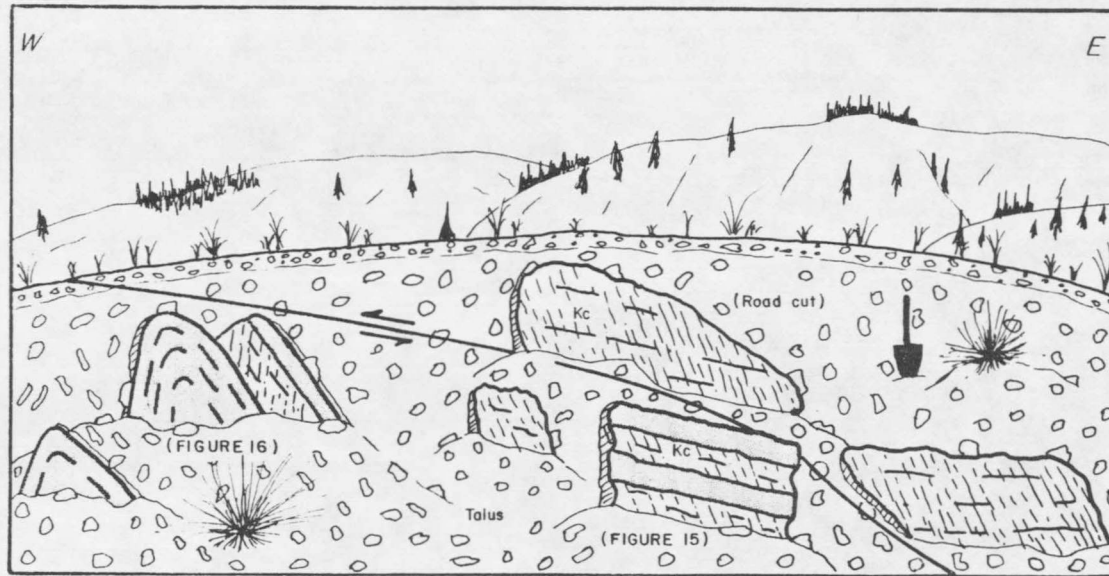


Figure 17. A Mount Princeton fault splay. The locations of Figures 15 and 16 are marked. This fault splay has apparently experienced east-west movement which may have created a force-couple and the consequent microstructures in the preceding two figures.

east-to-west tectonic transport. Thus, the Mount Princeton fault appears to be a backlimb thrust to the Douglas Mountain anticline.

Wasa-Dunkleberg Fault System. The Wasa-Dunkleberg fault system is not clearly exposed anywhere in the map area. It is inferred to crop out in Secs. 22 and 27, T. 9 N., R. 12 W., in a heavily forested area (Fig. 5 and Plate 1).

This area contains the abandoned Wasa and Forest Rose mines which exploited mineralization in the "Gastropod" Member of the Kootenai Formation. As in the Philipsburg area, these mines occur on the crest of a fold, in this case, the Dunkleberg Ridge anticline.

The author was unable to clearly map the faults, but their presence is inferred primarily from the occurrence of several repeated outcrops of the "Gastropod" Member. The faults are down on the east and place resistant ridges of the "Gastropod" Limestone adjacent to the dark-colored soil representing the Flood Member of the Blackleaf Formation.

Due to the heavy forest cover the attitudes of the faults were not established. However, Gwinn (1961) noted that the pitch of ore deposits which occur along the fault planes dips from 75° east to vertical. Stratigraphic throw on the Wasa-Dunkleberg fault system is estimated to be 100-300 feet.

Gwinn (1960) inferred the presence of a fault extending north along Dunkleberg Creek in the northeastern corner of T. 9 N., R. 12 W. (Plate 1). This fault represents a northern extension of the Wasa-Dunkleberg fault system. Gwinn (1961) observed that the fault offsets

the Cabbage Patch Beds (see Appendix), thus dating it as Late Oligocene. Mutch and Gwinn (1960) mapped an extension of the fault system in Secs. 14 and 15, T. 8 N., R. 12 W., where Allen (1966) observed offset of the Royal stock on both sides of the fault zone. Consequently, the Late Oligocene fault system has been dated as post east-to-west tectonic transport and pre-emplacement of the stock. Therefore, these faults are probably related to the same event which produced the Douglas Mountain fault.

Emery Ridge Imbricate Faults. The Emery Ridge imbricate faults crop out in a prominent ridge of Quadrant Quartzite overlying talus slopes, swales, and fault gouge in the Kootenai Formation. The outcrop trace is an arcuate pattern, concave to the northeast, occurring in Secs. 4, 5, 8, and 9, T. 8 N., R. 11 W., and Sec. 32, T. 9 N., R. 11 W. (Fig. 5 and Plate 1). A good exposure is found in the south wall of the Gold Creek Valley in Sec. 32, T. 9 N., R. 11 W. (Fig. 18), where a road cut exposes a zone of fault gouge in the Kootenai Formation (Fig. 19).

Mutch (1960; p. 1) described the Emery Ridge fault as an:

"anticline, overturned to the west, (which) is paralleled on the western side by multiple reverse faults and on its eastern side by normal faults."

The author mapped the area east of the faults as an allochthonous homocline bounded on the west and south by the Emery Ridge imbricate thrust faults and on the east and north by Cenozoic strata. A normal fault parallels the trend of the Emery Ridge imbricate faults and may be genetically related to them.

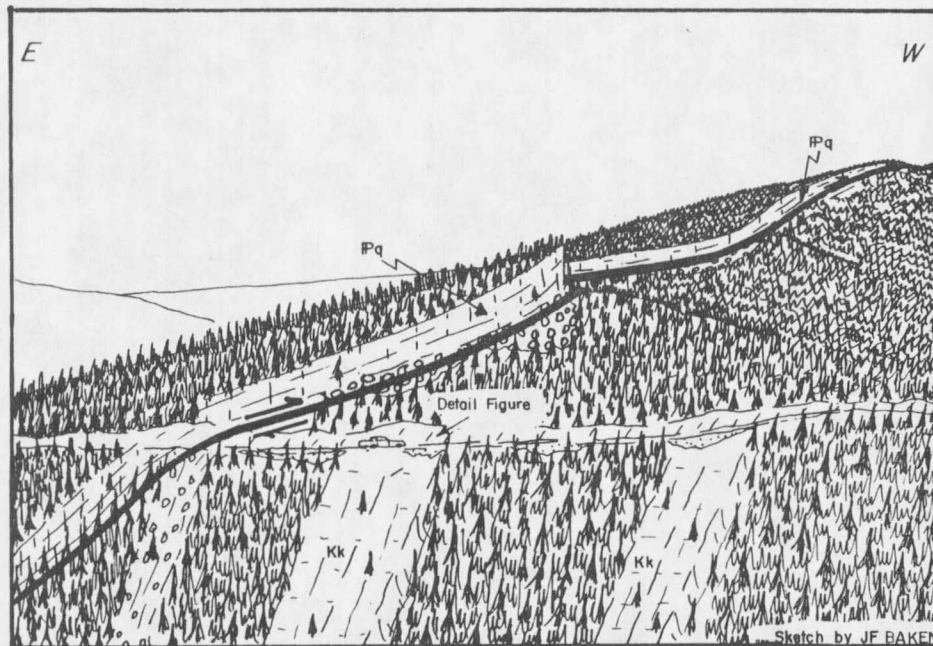


Figure 18. The Emery Ridge thrust on the south wall of Gold Creek. An imbricate of the fault is located to the south beyond the ridge. Quadrant is placed on overturned Kootenai. Figure 19 details fault gouge in the Kootenai.

The westernmost imbricate fault of the Emery Ridge thrust zone has an attitude of approximately 345° , 35° N.E. in Sec. 32, T. 9 N., R. 11 W., around the northeastern corner of the Royal stock. Apparently, forceful emplacement of this igneous body uplifted this segment of the allochthon and consequent erosion has produced the map pattern.

Immediately to the west and beneath the fault the Kootenai Formation is pulverized and weathered and is presumed to be fault gouge. Figure 19 shows an outcrop of this gouge in over- and underlying zones of relatively competent and undeformed rock separated by an incompetent, intensely deformed, middle zone. These upper and lower zones define the approximate attitude of the Kootenai in an overturned position ($340-345^{\circ}$, $45-55^{\circ}$ N.E.). The deformed middle zone displays varying attitudes which are indicated in the figure. These attitudes may indicate a force-couple caused by westward movement on the Emery Ridge thrust. Figure 6 shows the overturned Kootenai Formation beneath the Emery Ridge fault and Fig. 18 shows the same relationship in a detailed view.

Northeast-Trending Faults. The northeast flank of the Flint Creek Range is marked by several high-angle, northeast-trending faults which are indicated by stream drainage patterns. Figure 2 shows an obvious northeast trend of streams on this side of the range. Mutch (1960 and 1961) mapped these faults, but they were not directly observed by the author.

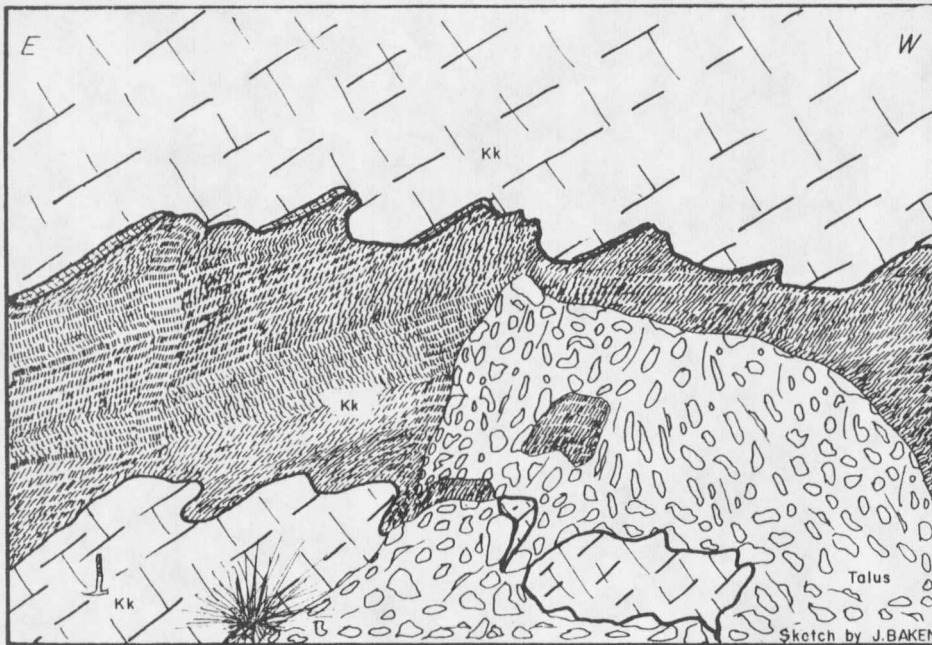


Figure 19. Fault gouge in the Kootenai Formation on Emery Ridge. The incompetent middle zone displays extreme granulation--probably the result of east-west movement of the overlying Emery Ridge thrust (See Fig. 18).

The majority of northeast-trending faults in the eastern province appear to cut pre-existing structure produced by east-to-west transport. In addition, Mutch (1961) observed that these faults offset the Royal stock. Therefore, these faults post-date west-directed transport and the intrusion of the Royal stock.

Folds. Princeton Anticline. The Princeton anticline is a well-exposed, overturned anticline. In Sec. 25, T. 9 N., R. 13 W. (Fig. 5 and Plate 1), Jurassic rocks form a nose with the Kootenai Formation (Fig. 20). The moderate east-dipping, east limb (335° , 25° N.E.) can be traced to the northwest where it changes trend and attitude to 080° , 43° N.W., in the crest of the fold, and finally to an overturned attitude of 015° , 40° S.E. in the western limb. In addition to this excellent exposure of the nose, the limbs of the Princeton anticline can be traced on the Quadrant Quartzite (Fig. 21) to the southeast quarter of T. 8 N., R. 13 W. (Fig. 6).

The Princeton anticline plunges 20° toward 012° (Fig. 22). The contoured pi diagram also shows the westward-verging character of the fold, diagnostic of the eastern structural province.

The Princeton fold displays an interesting characteristic which is shared by the other major folds of the eastern province. In the northern parts of these folds, Mesozoic rocks have been deformed into tight, asymmetric folds with westward vergence as mentioned earlier (Plate 1). However, to the south, up-plunge, these folds involve the more-competent Paleozoic rocks and therefore coalesce into fewer, less-asymmetrical folds (McGill, 1958).

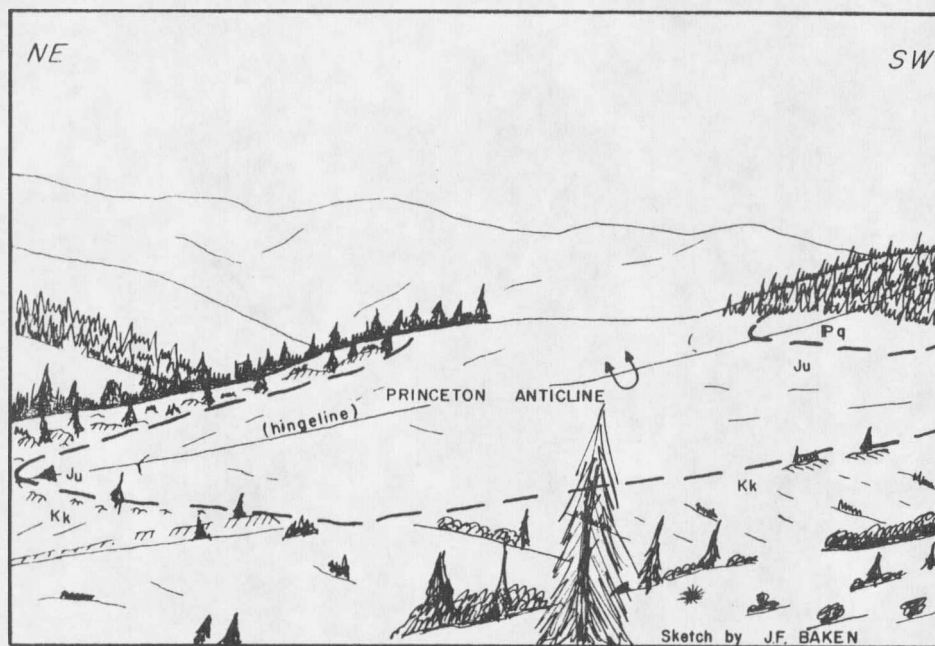


Figure 20. Anticlinal nose of the Princeton anticline. Jurassic and Cretaceous Kootenai rocks form the nose. The western limb in the foreground is overturned. The fold displays westward vergence (See Fig. 22).

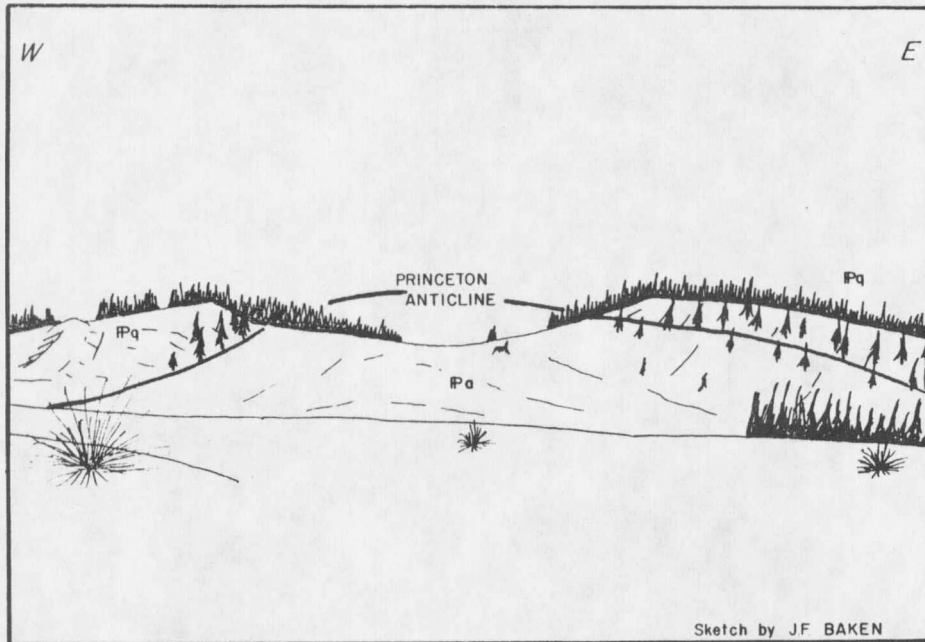


Figure 21. Quadrant ridges mark the Princeton anticline. The quadrant is an excellent marker bed throughout the area.

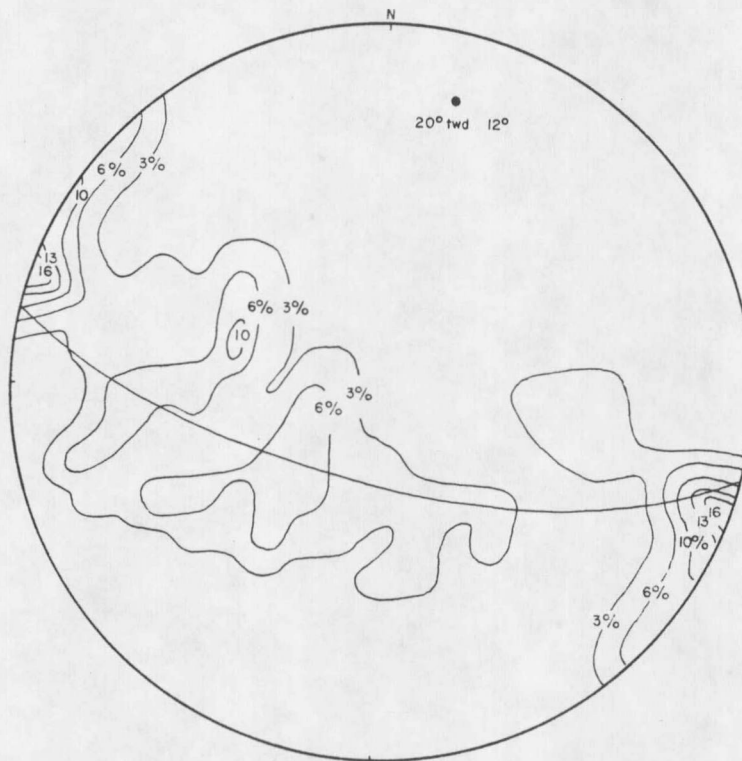


Figure 22. Contoured pi diagram of poles to bedding for the Princeton anticline. Westward vergence is apparent (See Fig. 20). ("N" is north.)

Douglas Mountain Anticline. Perhaps the most spectacular geologic structure in the northern Flint Creek Range is the Douglas Mountain anticline (Figs. 5, 6, and 23; and Plate 1). The best exposure of the Douglas Mountain anticline is on the north wall of Douglas Creek in Sec. 31, T. 9 N., R. 12 W. Resistant ridges of Pennsylvanian Quadrant stand out relative to the more-easily weathered Permian and Mississippian strata. This area is economically important as evidenced by the presence of numerous shafts, adits, strip mines and loading chutes from phosphate mining in the Permian rocks.

Due to recent oil and gas leasing activity in this area, it is interesting to note that the oil-producing Madison Limestone is breached in the core of the fold in Sec. 31. Presumably, any hydrocarbons which may have been present in this structure have long since vanished.

Poles to bedding for the Douglas Mountain anticline show a fold axis plunge of 15° toward 005° with the predictable westward-verging character (Fig. 24).

The western limb of the anticline is cut along much of its length by the previously described Douglas Mountain fault. Again, it is important to note that the east-dipping fault cuts both east-dipping, overturned, and west-dipping, upright portions of the western limb, therefore, faulting was post-folding.

Coberly Syncline. The Coberly syncline is a relatively obscure feature lying between the Mount Princeton and Dunkleberg Ridge anticlines in the center of T. 9 N., R. 12 W. (Fig. 5 and Plate 1).

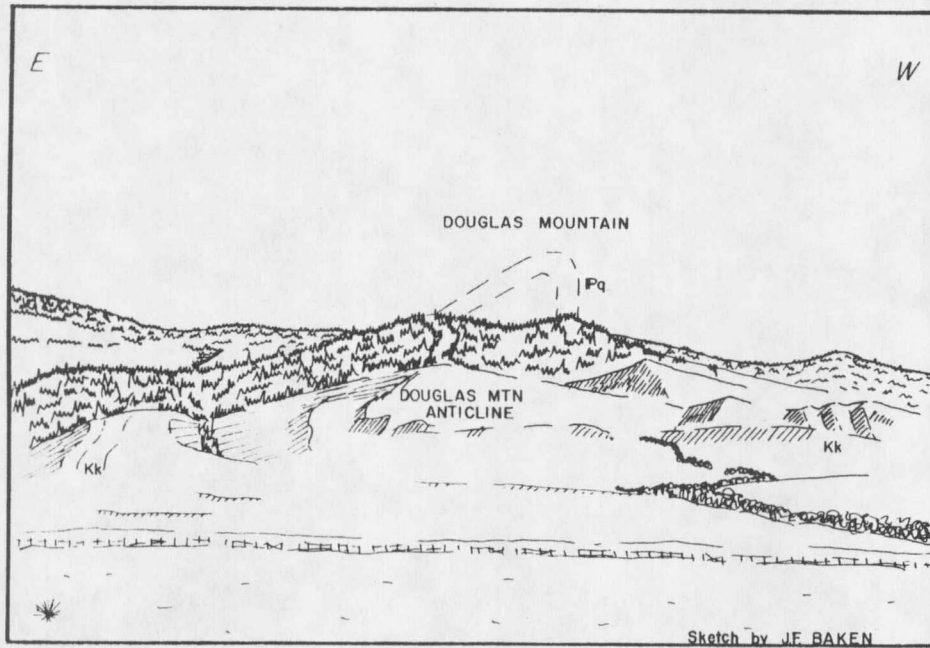


Figure 23. Douglas Mountain anticline on Douglas Mountain. This feature displays dramatic westward vergence (See Fig. 24).

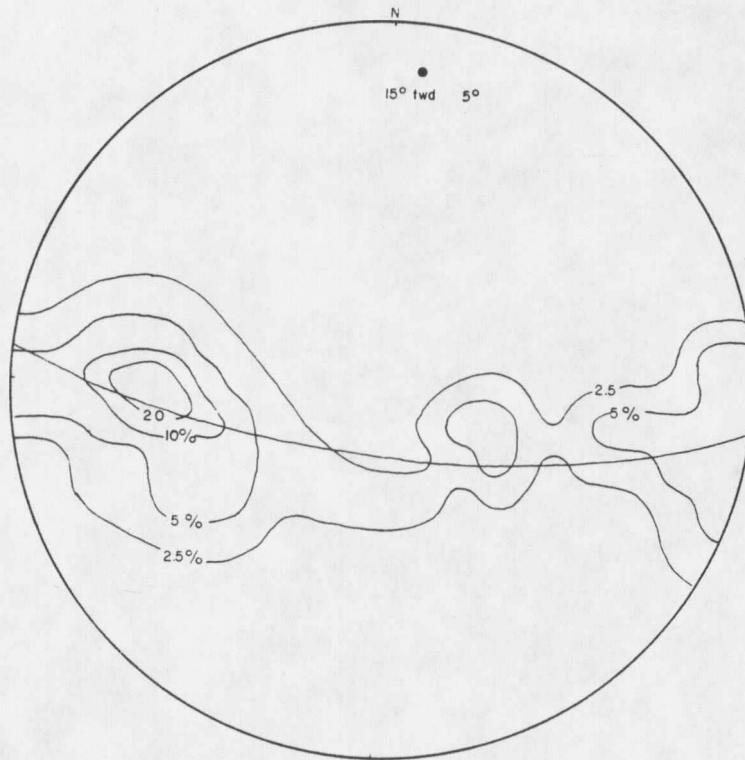


Figure 24. Contoured pi diagram of poles to bedding for the Douglas Mountain anticline. Westward vergence is demonstrated (See Fig. 23). ("N" is north.)

This structure can be traced from the southern end of the Garnet Range in the northern part of T. 10 N., R. 12 W., to the middle of T. 8 N., R. 12 W., where it is lost in the coalescing folds of the Northern Flint Creek Range.

The best exposures of the Coberly syncline occur where Cretaceous diorite sills have been folded to form structural noses in T. 10 N. and T. 9 N., R. 12 W.

The plunge of the Coberly syncline is 10° toward 009° in Secs. 9 and 10, T. 9 N., R. 12 W. (Fig. 25). In contrast, the plunge of the fold in Sec. 22, T. 10 N., R. 12 W., is 15° toward 163° (Fig. 26). Fig. 27 shows this relationship in a north-south cross section displaying Gwinn's (1960) "Clark Fork Sag".

The Coberly syncline lacks the westward asymmetry of other eastern-province folds (Fig. 25). This lack is shared by the Dunkleberg Ridge anticline and the Gold Creek syncline (Figs. 29 and 30). In all three folds Cretaceous diorite sills have been involved in folding. Gwinn (1961) believed that the sills imparted rigidity which retarded the production of westward asymmetry. Comparison of fold geometry in T. 9 N., R. 12 W., upholds this theory (Plate 1).

Royal Gold Creek Anticline. The Royal Gold Creek anticline crops out in Sec. 28, T. 8 N., R. 12 W. (Fig. 5 and Plate 1). This feature is, like most folds in this area, marked by the prominent Pennsylvanian Quadrant Quartzite.

The contoured pi diagram of poles to bedding (Fig. 28) for the Royal Gold Creek anticline shows a plunge of 10° toward 355° .

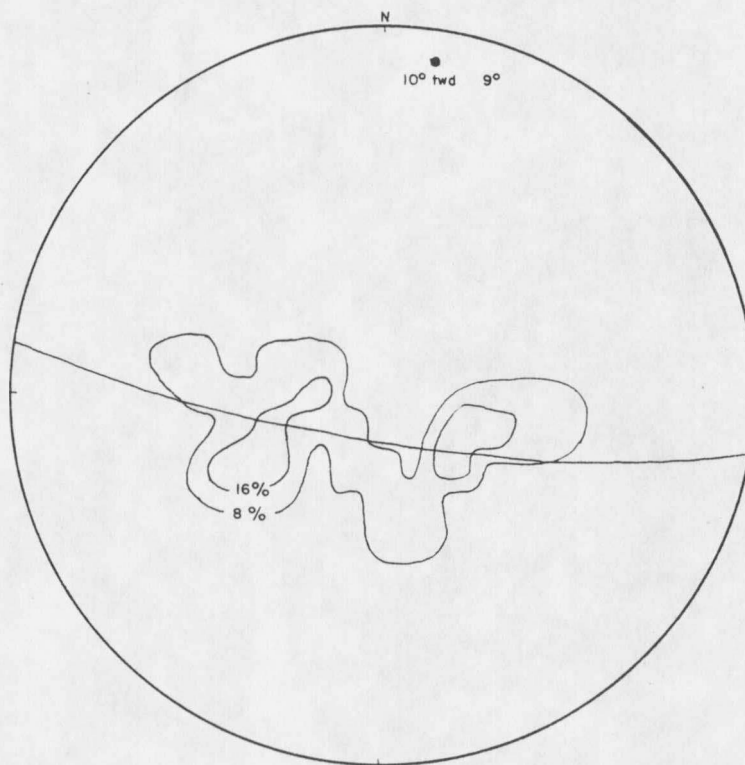


Figure 25. Contoured pi diagram of poles to bedding for the southern portion of the Coberly syncline. Good symmetry and a north plunge are displayed (See Figures 26 and 27). ("N" is north.)

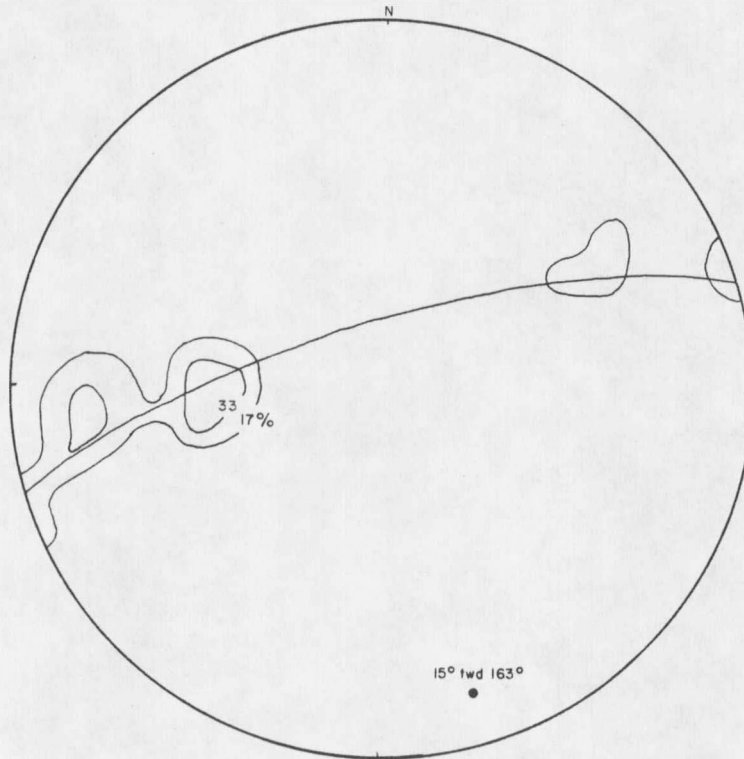


Figure 26. Contoured pi diagram of poles to bedding for the northern portion of the Coberly syncline. A comparison of fold-plunge attitudes between the northern and southern portions of the syncline reveals the Clark Fork "Sag" (Gwinn, 1960) (See Figures 25 and 27). ("N" is north.)

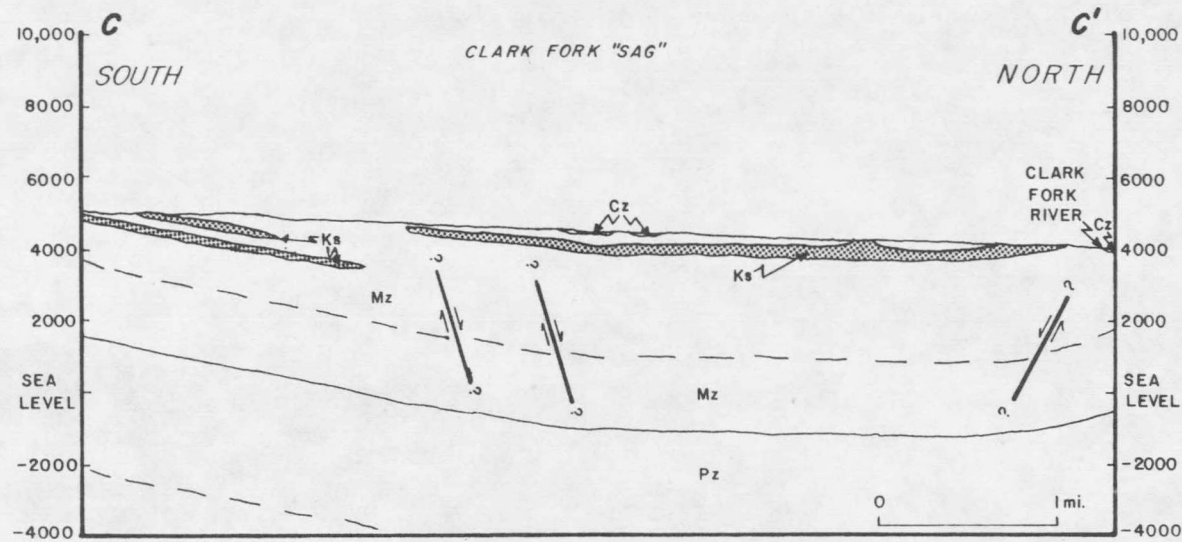


Figure 27. Cross section CC'. This westward view demonstrates the Clark Fork "Sag" in the Coberly syncline (Figures 25 and 26) (Gwinn, 1960). Mutch and Gwinn (1960) suggested the presence of covered, east-west trending faults in the Clark Fork Valley. These faults may be responsible, in part, for the structural lowland. [This view is along a portion of the Montana lineament. (See Fig. 32)]

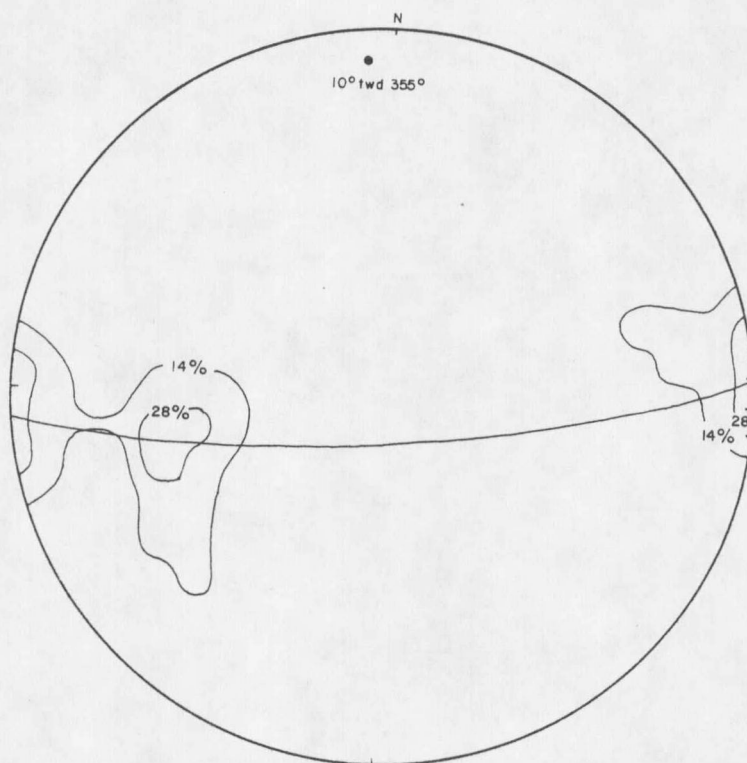


Figure 28. Contoured pi diagram of poles to bedding for the Royal Gold Creek anticline. This fold was probably formed by forceful intrusion of the Royal stock to the east (Plate 1). ("N" is north.)

Forceful emplacement of the Royal stock apparently caused the change in trend of the Emery Ridge thrusts. In the author's opinion the same event may have been responsible for the formation of the Royal Gold Creek anticline. A cursory look at Plate 1 reveals three facts supporting this view: 1) the Royal Gold Creek anticline remains tight and asymmetrical (Fig. 28) and is flanked by two tight synclines, while other folds in the vicinity coalesce into broad, open folds (Plate 1), 2) the fold is localized and occurs only adjacent to the Royal stock, and 3) the fold axis parallels the flank of the stock.

Dunkleberg Ridge Anticline. The Dunkleberg Ridge anticline crops out in the northeast quarter of T. 9 N., R. 12 W. in Secs. 1, 2, 11, 12, and 14 (Fig. 5 and Plate 1). This fold is not easily observable in the field due to heavy forest cover, but detailed mapping reveals anticlinal noses involving Cretaceous diorite sills (in the locations mentioned above) and Paleozoic rocks (Fig. 6) in the northern part of Secs. 3 and 4, T. 8 N., R. 12 W.

The contoured pi diagram shows a plunge of 15° toward 012° and substantiates the upright, symmetrical character of the fold seen in the map pattern (Fig. 29). The Dunkleberg Ridge anticline does not display westward asymmetry.

The anticline is cut by the flank of the Royal stock in the south half of T. 8 N., R. 12 W. Forceful emplacement of this body was apparently responsible for the termination of the fold.

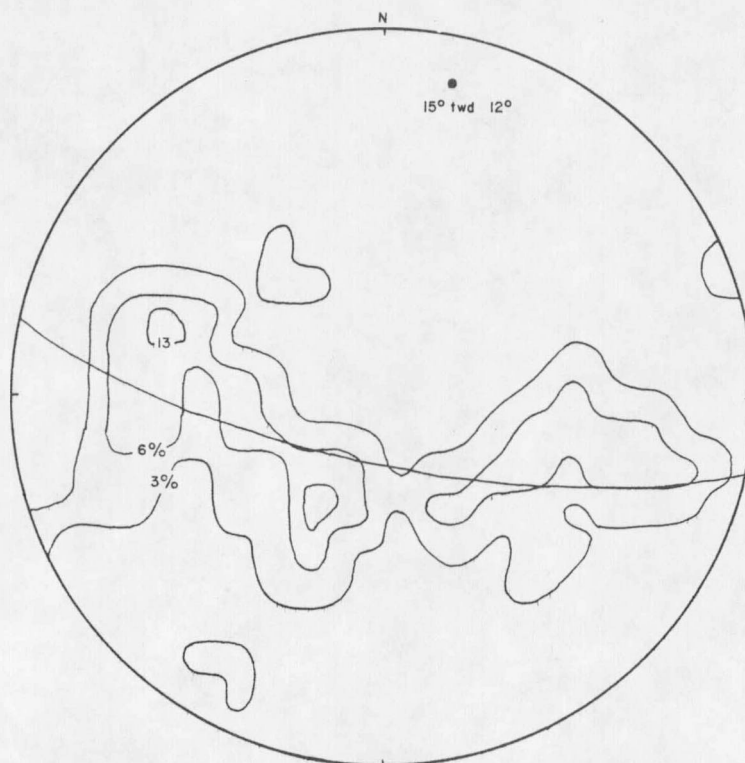


Figure 29. Contoured pi diagram of poles to bedding for the Dunkleberg Ridge anticline. Note the symmetry. ("N" is north.)

A small east-west trending syncline crops out on the northwest flank of the Royal stock as a high, broken mass of contact-metamorphosed Jurassic, Permian, and Upper Pennsylvanian strata (Plate 1). Attitudes of the beds are difficult to measure, but the map pattern indicates a synclinal form. Paleozoic rock crops out adjacent to, and dips away from, the north flank of the Royal stock in Secs. 12, 13, 17, and 18, T. 8 N., R. 11 W. This area has apparently been exposed by uplift along the flank of the stock during the same event that produced the small Jurassic syncline.

The crest and flanks of the Dunkleberg Ridge anticline have been cut by the Wasa-Dunkleberg fault system, as mentioned earlier. Gwinn (1961) observed that the crest had been fractured and hydrothermally altered. The Wasa and Forest Rose mines have exploited these elements (Silverman, 1965).

Gold Creek Syncline. The Gold Creek syncline is a broad, open, north to northeasterly trending fold that crops out in the south half of T. 10 N., R. 11 W.; the northwest quarter of T. 9 N., R. 11 W.; the southeast quarter of T. 9 N., R. 12 W.; and the northeast quarter of T. 8 N., R. 12 W. The fold involves Mesozoic and Cretaceous sills, and the hinge line forms a slightly sinuous pattern terminating near the north flank of the Royal stock (Fig. 5 and Plate 1).

The contoured pi diagram for the Gold Creek syncline shows a plunge of 10° toward 018° (Fig. 30).

Mutch (1961) mapped a change in the trend of the hingeline of the Gold Creek syncline to a southeasterly direction in the northeast

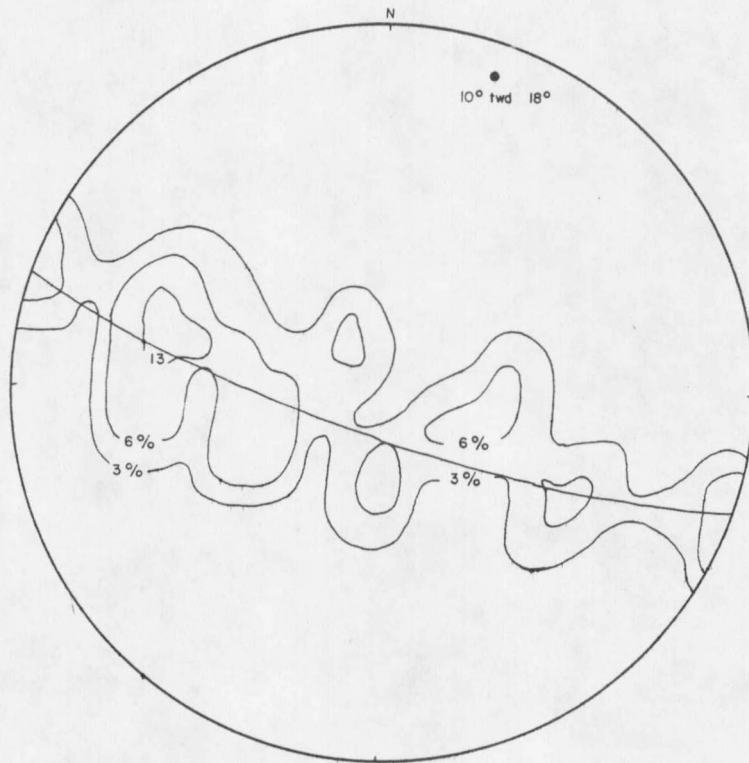


Figure 30. Contoured pi diagram of poles to bedding for the Gold Creek syncline. Note the symmetry. ("N" is north.)

quarter of T. 8 N., R. 12 W., terminating in Sec. 7, T. 8 N., R. 11 W., near the juncture of the Royal stock and the Emery ridge thrusts (Plate 1). The author's mapping does not support this interpretation. Cross section AA' (Fig. 6) reveals an overturned syncline in the Mesozoic rocks immediately to the west of the Emery Ridge imbricate thrust faults. The north-south hingeline of this small fold is placed in Secs. 6 and 7, T. 8 N., R. 11 W. This fold, in Kootenai strata, was apparently created by movement along the Emery Ridge thrusts. In addition, mapping of the diorite sill in Sec. 2, T. 8 N., R. 12 W., indicates that the hingeline of the Gold Creek syncline is in that location (Plate 1).

Northern Province

The northern province is structurally indistinct and is recognized primarily by relatively flat-lying Tertiary strata in the Clark Fork River Valley that unconformably overlie the pre-Tertiary structures of the Flint Creek and Garnet Ranges (Figs. 5 and 31; and Plate 1). This northwest-southeast trending zone may represent a segment of the Montana lineament (Fig. 3) (Weidman, 1965).

The northern province was termed the "Clark Fork Sag" by Gwinn (1960 and 1961) and is distinguished by the structural trend change across the valley with the consequent map pattern convex to the northeast and the opposing plunges where the axial traces of the folds from the Garnet and Flint Creek Ranges meet (Fig. 32).

Gwinn (1961; p. 1) observed that the youngest faults seen in this region are east-west normal faults which are:

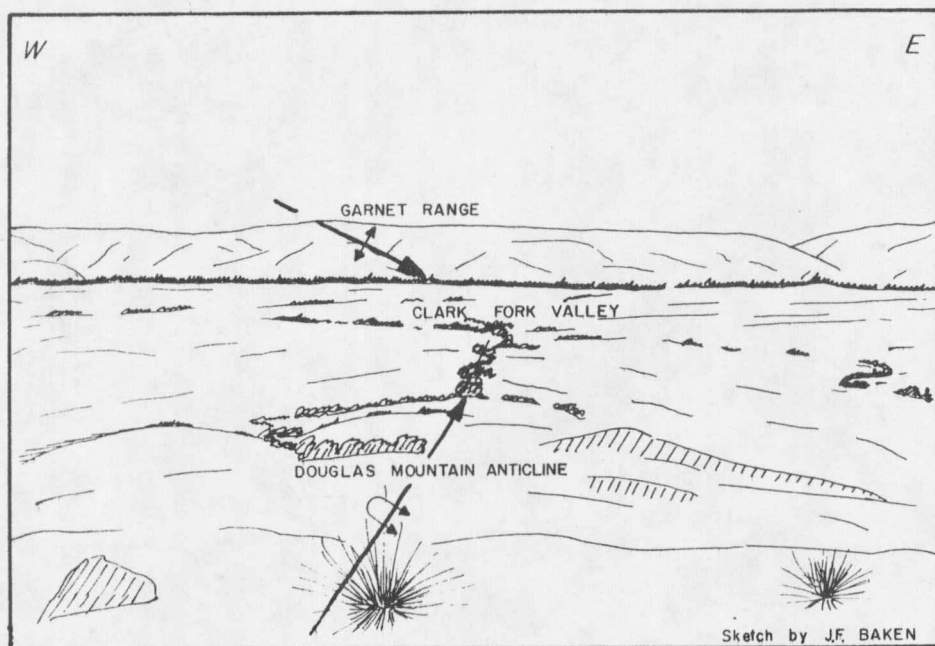


Figure 31. The Clark Fork Valley from Douglas Mountain. This view of the northern province shows the characteristic, flat-lying Tertiary strata which covers the Clark Fork "Sag" (Gwinn, 1960) (Fig. 27). The structural trend change and opposing plunges are dramatized by the hingeline traces (See Fig. 32).

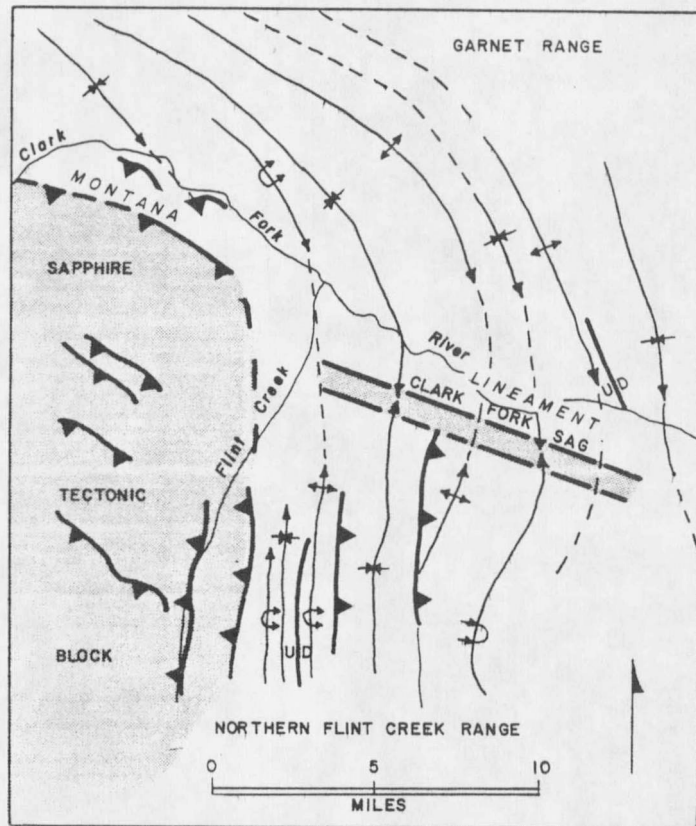


Figure 32. The northern structural province. The Clark Fork "Sag" (Gwinn, 1960) and a structural trend change characterize this province (See Figures 27 and 31). (Modified from Weidman, 1965).

"down-dropped to the north in the Flint Creek Range and down-dropped to the south in the Garnet Range" (Fig. 27).

These faults apparently post-date all other faults and folds in the ranges. The author did not observe these east-west faults and Gwinn (1961) did not map them. However, Gwinn states that their presence is inferred by stratigraphic relationships in the Tertiary strata of the area. These stratigraphic relationships are: 1) the presence of successively younger (Oligocene, Miocene, and Pliocene) strata toward the center of the valley and lower elevations, and 2) the younger, lower Cabbage Patch Beds (Pliocene) dip steeply toward the basin while the Oligocene and Miocene strata maintain relatively small dips in the same direction (Gwinn, 1961). These observations infer the effect of these faults on the deposition of the Tertiary strata. However, more concrete evidence is desirable to support that movement of these faults uplifted the Garnet and Flint Creek Ranges resulting in the opposing fold plunges (Fig. 27).

An obvious mechanism for the creation of the structural trend change in the northern province is left-lateral, strike-slip movement along the Montana lineament. The absence of any northeast-southwest trending extensional features in either the Tertiary or pre-Tertiary rocks lends evidence for movement along this lineament concurrent with folding of the strata in the Garnet and Flint Creek Ranges (Fig. 32).

Regional Tectonic Features

Introduction

Five regional tectonic features surrounding the Flint Creek Range appear to have effected the origin of the structure in the study area. These elements are the Sapphire tectonic block and Idaho batholith on the west, the Boulder batholith on the east, the Montana lineament (Lewis and Clark line) on the north, and the Jefferson Valley or Willow Creek lineament (Perry line) on the south (Fig. 3). The Belt embayment, a broader phenomenon, is the fifth structural feature and differs from the first four in that it is delineated primarily by the Proterozoic stratigraphic succession; however, its effect on the regional structure of Western Montana has been profound (Fig. 33).

Sapphire Tectonic Block

The Sapphire tectonic block is an allochthon measuring 100 kilometers (east-west) by 75 kilometers (north-south) and is 15-21 kilometers thick. It is defined laterally by the Idaho batholith and a zone of mylonitic foliation in the west; large-scale, west-dipping thrust faults in the east (the Philipsburg and Georgetown-Princeton thrusts); east-trending, high-angle, south-dipping strike-slip faults on the north; and the inferred projection of the Jefferson Valley lineament in the south (Hyndman, et al, 1975 and 1979) (Fig. 3). The structure within the allochthon is characterized by limited deformation in the west with large-scale thrust faulting in the east (Hyndman, et al, 1975).

The zone of mylonitic foliation which marks the western border of the Sapphire tectonic block is a zone of:

"penetrative mineral and shear foliation and east-southeast trending lineation" (Hyndman, 1979, p.6).

Hyndman (1975) also noted that the mylonitic fabric gradually steepens from the horizontal in the west (in contact with the Idaho batholith) to 20-25° east dip in the east. The zone is characterized by: 1) penetrative foliation which covers the entire dome of the batholith, but is much stronger on the east flank, 2) a penetrative and generally unidirectional mineral streaking and slickenside lineation within the foliation, which is much stronger on the east flank, and 3) a constant, unidirectional, east-southeast orientation of the lineation which is not radial to the dome even where the mylonitic zone wraps around the southeastern part of the batholith (Hyndman, 1980). In addition, the north-south dimensions of the mylonitic zone are closely matched by the north-south dimensions of the Sapphire tectonic block immediately to the east (Hyndman, et al, 1974) (Fig. 3).

The eastern border of the Sapphire tectonic block is marked by the Philipsburg and Georgetown-Princeton thrust faults described in the previous section.

The northern border of the Sapphire allochthon is marked by a zone of faulting which is composed of strike-slip faults which change eastward into south-dipping thrust faults with a large component of left-lateral, strike-slip movement (Nelson and Dobell, 1961; and Smedes, 1973). These thrust faults change trend to the southeast and

south, probably connecting with the Philipsburg and Georgetown-Princeton thrust faults (Fig. 32). The thrust faults apparently die out to the south in a zone of ductile deformation as mentioned previously (Local Structural Geology).

Boulder Batholith

The Boulder batholith (78-72 m.y.b.p.) intruded the Belt Supergroup, Paleozoic and Mesozoic rocks, and approximately 10,000 feet of Upper Cretaceous Elkhorn Mountains Volcanics (Robinson, et al, 1968; and Hyndman, 1979). The batholith is a hypabyssal, tabular body 50 kilometers (east-west) by 100 kilometers (north-south) and is 5-15 kilometers thick (Klepper, et al, 1971, 1974; and Hamilton and Meyers, 1974a, 1974b) (Fig.3). The body is composed of several plutons with an average composition of quartz monzonite to granodiorite (Robinson, et al, 1968).

The batholith appears to lie in a structural basin of regional extent as evidenced by the gentle dip of the country rocks surrounding the pluton (Hamilton and Meyers, 1974a, 1974b; and Klepper, et al, 1971, 1974). Hyndman (1979) states that the late Precambrian Belt and perhaps the Paleozoic and Mesozoic sections at the north end of the batholith probably dip under the body. Bouguer anomalies over the batholith suggest that the depth to the pre-Belt basement allows room for buried supracrustal rocks beneath the batholith (Biehler and Bonini, 1969; and Hyndman, 1979). Hamilton and Meyers (1974a) observed that emplacement of the Boulder batholith evidently depressed rocks near the surface to the depths at the base of the batholith.

East and northeast of the Boulder batholith the structure is characterized by broad, open, upright folds which become tighter and less symmetrical to the east with eastward vergence (Klepper, et al, 1974; Mertie, et al, 1951; and Robinson, et al, 1968). Thrust faulting is evident in the east with a zone of thrust faults trending south-southeast from the Montana disturbed belt into the area. This zone is approximately 24 kilometers wide and displays westward dip (Robinson, et al, 1968).

West of the batholith the structure is characterized by:

"folds in the Elkhorn Mountains Volcanics; joint sets cutting the Elkhorn Mountains Volcanics and batholithic rocks; northwest and northeast-trending lineaments...; east trending fault zones,...; and other faults of different trends..." (Ruppel, 1963).

Most authors agree that the batholith did not shed its roof rocks tectonically. There is some evidence for the gravity sliding of minor allochthons of Elkhorn Mountains Volcanics eastward (Hamilton and Meyers, 1974a), but no evidence for major movement of pre-intrusion country rocks has been documented (Klepper, et al, 1971; Hyndman, 1979).

The Boulder batholith is separated from the Flint Creek Range by approximately 16 kilometers of the block-faulted Deer Lodge Valley.

Montana Lineament

Weidman (1965, p. 137) defined the Montana lineament as a:

"regional tectonic feature... recognized on the basis of diverse aligned geologic elements, including straight valleys and dextral (right-lateral) strike-slip faults on

the west, south-dipping thrust faults in the Missoula-Drummond area, a sharp trend change near Drummond, and a zone of en echelon southeast-plunging folds between Drummond and Helena."

In the Flink Creek Range the northern structural province represents the Montana lineament in the Drummond area. This representation takes the form of structural trend change, opposing fold plunges, and covered, east-west trending, high-angle faults (Figs. 27 and 32).

The Montana lineament is marked by the northern boundary of the Sapphire tectonic block west of Drummond which is characterized by the high-angle faults with left-lateral, strike-slip movement previously discussed (Fig. 3).

Jefferson Valley Lineament

The Jefferson Valley lineament (Fig. 3), termed the "Perry line" by Harris (1957) and the "Willow Creek fault" by Robinson (1961), is an east-northeast trending zone of crustal movement (Csejtey, 1963). Evidence for this zone was suggested by Harris (1957) and McMannis (1963) with their observation of the distribution of the Precambrian Belt Supergroup in the Bridger Range and Jefferson Canyon.

North of the Jefferson Valley lineament the Belt is represented by a wedge of coarse-grained arkoses and arkosic conglomerates, and south of the lineament the Belt Supergroup is absent. Sahinen (1939) and Mutch (1960) extended the lineament across the Highland Mountains (south of Butte) to connect with the southern boundary of the eastward-projecting, Bitterroot lobe of the Idaho batholith.

Alexander (1951) observed a series of northeast-trending folds in the Whitehall area north of the Jefferson Valley lineament which serve to delineate the feature in that area.

Belt Embayment

The Belt embayment:

"represents a slowly sinking reentrant on the North American craton that began to form about 1,500 m.y. ago and persisted for more than 600 m.y." (Harrison, et al, 1974, p. 1).

This feature (Fig. 33) was identified by Harris (1957) from isopach maps of various Precambrian Belt Formations. The southern boundary, delineated by McMannis (1963), corresponds to the trace of the Jefferson Valley lineament and projects northeast into central Montana (Figs. 3 and 33). The northeastern boundary is marked by a gradual thinning of Belt strata (McMannis, 1965). The east-west linear character of the embayment indicates that the Montana lineament had some control over its dimensions. Harris (1957) observed that the Belt embayment generally occupied the area bounded by the Jefferson Valley and Montana lineaments.

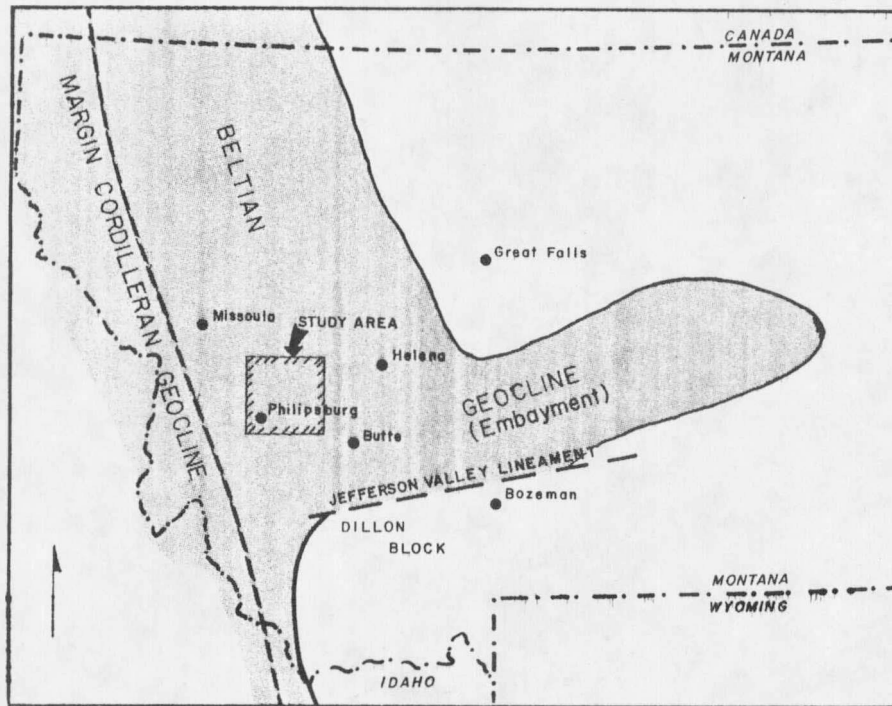


Figure 33. The Belt embayment. This feature is an east-west geocline which is filled with Belt sediments. It is bound on the south and north by the Jefferson Valley and Montana (not shown) lineaments (Modified from Harris, 1957).

THE KINEMATIC SEQUENCE OF DEFORMATION
IN THE FLINT CREEK RANGE

Introduction - Regional Tectonic History

The regional tectonic history of the last 145 m.y. will be the focus of the remainder of this paper, followed by the kinematic sequence of deformation in the Flint Creek Range and adjacent area (Table 1).

Two schools of thought currently exist concerning magmatic activity in the Cordillera during the last 210 m.y. One, is that magmatic activity was relatively continuous throughout the past 210 m.y. (Burchfiel and Davis, 1975). However, Lanphere and Reed (1973) recognized two and possibly three intrusive periods: an undefined period from 210-160 m.y., and two defined epochs: 1) 153-132 m.y. and 2) 106-79 m.y. (Table 1). The important point is that during the last 145 m.y. magmatism prevailed in the Cordillera.

The Cordillera was subjected to zones of high heat flow produced by this magmatic activity. Armstrong (1974) compared the occurrence of these zones to tectonic deformation and large-scale crustal shortening. He concluded that deformation in the Cordillera was directly associated with zones of high heat flow.

Armstrong differentiated the Sevier (145-75 m.y.) and Laramide (72-52 m.y.) orogenies (Table 1). The Sevier orogeny was characterized by thin-skinned, fold-and-thrust belt style geometry and

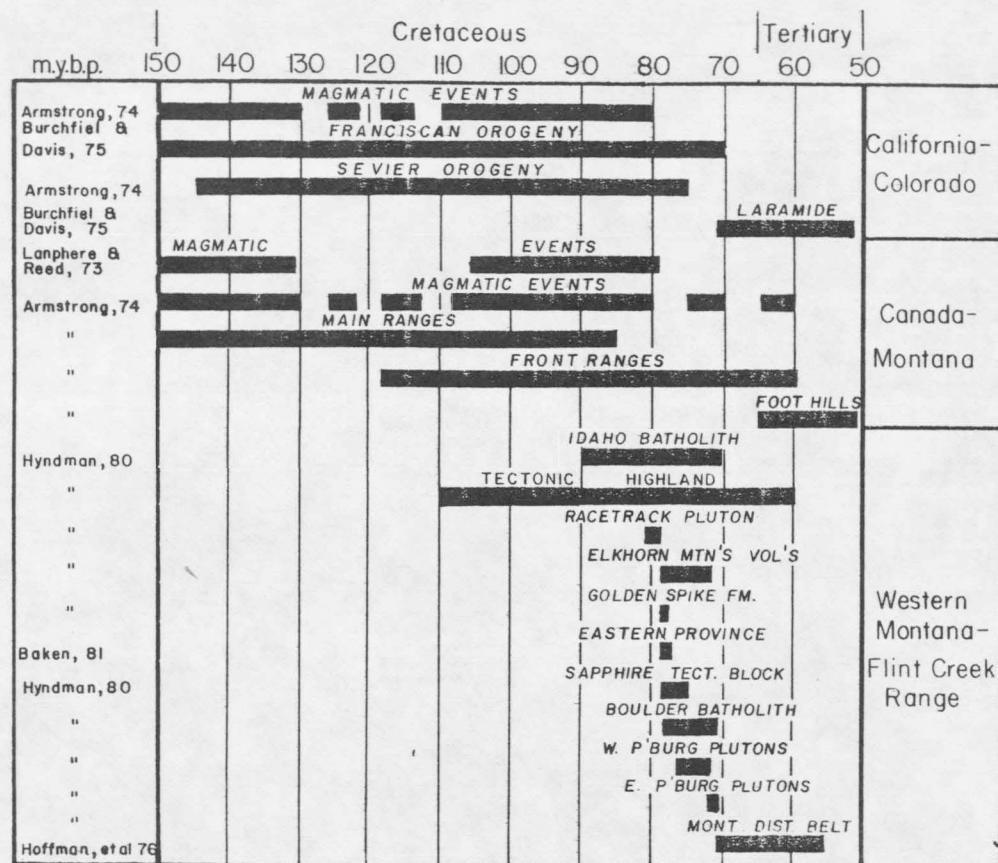


Table 1. Timing of regional tectonic events.

occurred along the entire North American Cordillera from Alaska to Mexico. Approximately 80 m.y.b.p. in a sector from central Utah to southeastern California, Sevier-style deformation ceased. North and south of this region, thin-skinned deformation continued through late Cretaceous to late Eocene (Armstrong, 1974; and Burchfiel and Davis, 1975) (Table 1). Within the central Utah-southeastern California sector Laramide-style deformation became dominant (Armstrong, 1974). This geometry was characterized by:

"differential vertical displacement of Precambrian basement rocks of up to 9 or 10 kilometers and bounding low- to high-angle reverse faults that steepen at depth" (Burchfiel and Davis, 1974, P. 380).

Armstrong (1974) suggested that approximately 80 m.y. ago magmatism ceased in the central portion of the Cordillera causing a cessation of Sevier-style deformation and a change to foreland or Laramide-style deformation. Thin-skinned deformation continued north and south of the central portion through Laramide time (72-52 m.y.).

From Middle-Late Tertiary to Holocene the Cordillera was subjected to extensional tectonics exemplified by the Basin and Range orogeny (Hamilton and Meyers, 1966). This orogenic style is characterized by large-scale, listric normal faults superimposed on the older Sevier structural geometry resulting in grabens and half-grabens filled with conglomeratic detritus (Scholten, 1960).

The present cycle of erosion and deposition is due to regional epeirogenic uplift of the Rocky Mountain region.

Deer Lodge Block Development

In the region surrounding the Flint Creek Range, the major magmatic and tectonic events took place prior to thrusting in the Montana disturbed belt 72-56 m.y. (Hoffman, et al, 1976; and Hyndman, 1980) (Table 1).

The tectonic history relevant to the Flint Creek Range began approximately 100 m.y. ago with east-west regional compression and the doming of a western tectonic highland which continued until 60 m.y. ago (Table 1) (Ryder and Scholten, 1973). The uplift was the result of the emplacement of the Idaho batholith (90-70 m.y.) (Hyndman, 1980) (Fig. 36).

Vertical and lateral mushrooming of the batholith may have created a gravity-sliding potential. Intrusive wedging combined with regional east-west compression may have caused eastward deformation along the Belt embayment and adjacent Montana and Jefferson Valley lineaments (Fig. 34).

Harris (1957) considered the effect of the Belt depositional pattern on this deformation. He concluded that the pattern of deformation in western Montana was controlled by the configuration of the east-west Belt embayment. Winston, et al (1982) concurred and suggested that the package of Belt to Mesozoic rocks within the embayment became an allochthon due to the Cretaceous and Eocene east-west compression. This allochthon, or Deer Lodge block (Winston, et al, 1982), resembles Mutch's (1960) "wedge" and is bounded on the north and south by the Montana and Jefferson Valley lineaments.

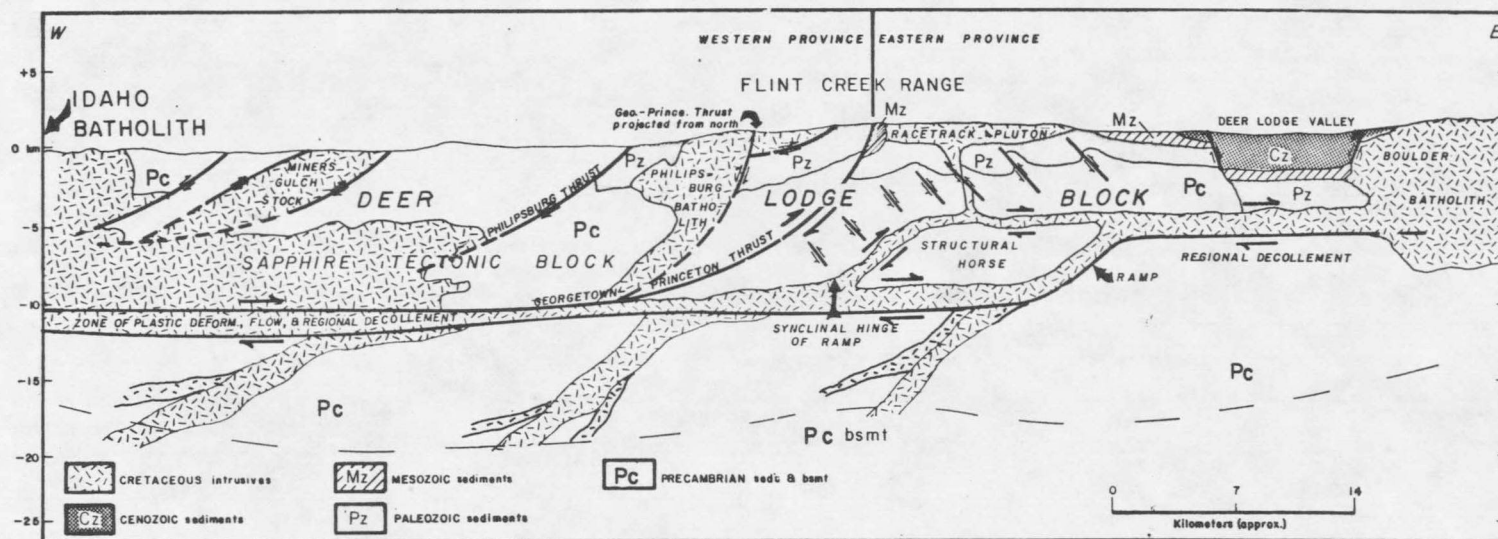


Figure 34. Regional deep cross section from the Sapphire tectonic block to the Boulder batholith. The regional decollement delineates the Deer Lodge block. The Georgetown-Princeton thrust marks the eastern limit of the Sapphire block. The Boulder batholith may have intruded along the decollement. The theorized ramp beneath the Flint Creek Range may be responsible for the development of the westward vergence in the eastern province (See Fig. 36). (Modified from Hyndman, 1979).

Inherent in this hypothesis is the presence of a regional decollement above crystalline basement (Fig. 34). This may be represented by the Lombard fault system which crops out in the east (Woodward, 1981; and Winston, et al, 1982) (Fig. 3). Schmidt and Hendrix (1981) have measured right-lateral displacement along the Jefferson Valley lineament at 16 to 30 kilometers.

Structural evidence in support of this concept is: 1) the change in structural trend from the Montana disturbed belt into the central Montana salient (Fig. 4), 2) the arcuate patterns of the Lombard and Philipsburg thrusts along the Montana and Jefferson Valley lineaments (Fig. 3), 3) the northeast-trending en echelon folds in the Whitehall area along the Jefferson Valley line, and 4) the northwest-trending folds of the Garnet Range along the Montana lineament (Fig. 32).

Igneous Activity - Phase I

Hyndman, et al (1975) and Hyndman (1979) proposed that the Boulder batholith initiated emplacement along the regional decollement following development of the Deer Lodge block (Fig. 34). The author concurs and suggests that movement along the regional decollement also provided the conduit by which the emplacement of the Racetrack Creek pluton occurred approximately 82-79 m.y. ago (Fig. 34 and Table 1). Slightly later, to the east, the Elkhorn Mountains Volcanics (78-73 m.y.) erupted above the intruding Boulder batholith. The source of these magmas may have been the still unsolidified Idaho Batholith (Hyndman, 1979) (Fig. 34).

Meanwhile, the Golden Spike Formation (78 m.y. old) was deposited at the future site of the Flint Creek Range (Table 1). The ingredients for the conglomerate were derived from the tectonic highland to the west (Fig. 36) and the Elkhorn Mountains Volcanics to the east.

Eastern Province Development

The westward-verging structural geometry of the eastern province in the Flint Creek Range presents a problem concerning its genesis. The difficulty is caused by the absence of an appropriate tectonic mechanism to produce the structure. A west-directed component of stress seems required for the solution of this problem. The following sections present two plausible tectonic mechanisms or models, which may have acted to produce this westward-verging structure.

Boulder Batholith Model

In 1960, Gwinn and Mutch stated their belief that forceful emplacement of the Boulder batholith was responsible for the eastern-province structural style. Robinson, et al (1968) presented evidence supporting this view. They summarized the age dates and time relationships of volcanism, plutonism, folding and thrusting around the batholith to indicate that the body was forcefully emplaced with a time sequence in any given locality consisting of: 1) volcanism preceding and ending prior to major folding, 2) plutonism ending after the climax of volcanism, 3) folding the country rocks preceding and accompanying plutonism near the batholith, but following plutonism at

locations away from the body, and 4) thrusting continuing some time after the climax of folding.

Contrary to Robinson, et al; Klepper, et al (1974) argued against the idea of forceful emplacement of the Boulder batholith. They argued that the emplacement of the batholith occurred after the creation of the folding in the Flint Creek Range and as evidence they maintained that the structures in the west decrease in intensity eastward (toward the batholith). Additionally, these structures occur on the flanks of the petrologically and chronologically similar Flint Creek plutons which cut them. Klepper, et al (1974, p. 1956) conclude that the:

"structures, then, are older than the Boulder batholith. Their origin is puzzling, but they cannot be ascribed to sliding off (or being deformed by) an igneous body that had not yet been emplaced."

In the author's opinion, the structures in the Flint Creek Range do not decrease in intensity toward the Boulder batholith. In fact, the eastern and northeastern borders are characterized by an increase in fault density (Plate 1). In addition, the sixteen kilometers of terrane separating the two features is buried in the block-faulted Deer Lodge Valley.

Secondly, comparison of the age dates of the Flint Creek plutons to the Boulder batholith appears to be, at best, subject to conjecture. Even comparing the absolute dates determined by Hyndman (1972 and 1980) (76.7-74 m.y. and 73.4-72 m.y. for the Flint Creek plutons, and 78-72 m.y. for the Boulder batholith) (Table 1) allows

1.3 m.y. for at least partial emplacement of the Boulder batholith prior to the first Flint Creek intrusion.

In summary, it is the author's view that forceful emplacement of the Boulder batholith as a model for the formation of the eastern province is a viable concept.

Regional Decollement-Ramp Model

Rich (1934) proposed a hypothesis for describing deformation and back thrusting in the ramp regions of the Pine Mountain overthrust in the Appalachians, which has been expanded by Serra (1977) and Morse (1977). These models of deformation can be applied to deformation in the eastern province of the Flint Creek Range.

Figure 35 shows the necessary geometry beneath the Flint Creek Range to accomplish back thrusting in the eastern province. The hypothesis assumes the presence of a major decollement and ramp interface. Another necessary ingredient is east-directed compression.

Winston, et al (1982) suggested the possibility that northwest trending reverse faults described by Schmidt and Hendrix (1981), south of the Jefferson Valley lineament, may extend northward into the Deer Lodge block. The author suggests that these extended faults may have provided a crystalline buttress over which the Deer Lodge block formed a ramp. Furthermore, Winston, et al suggested that the Deer Lodge block formed ramps over other northwest trending Precambrian lineaments or crystalline buttresses to the east.

Rich (1934) observed that thrust faults commonly develop treads or flats parallel to bedding in weak rock units with ramps cutting

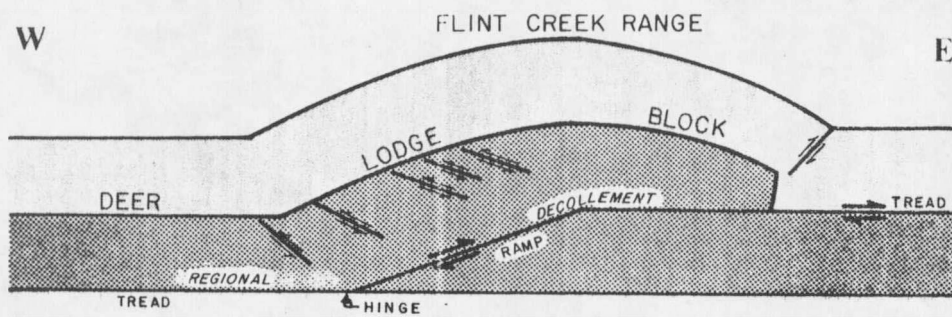


Figure 35. Regional decollement and ramp model. Deformation of the overriding block through the hinge area of the ramp may have caused backthrusting and westward vergence in the eastern province (See Fig. 35).

across more competent layers. There are two possible rock packages in which this geometry might develop beneath the Flint Creek Range (Table 2). In the lower part of the section, the Precambrian Missoula Group and Cambrian Flathead rocks provide possible competent layers with the Madison Group serving the same purpose up-section. The weak layers within which faults may have moved parallel to bedding are the Cambrian Silver Hill and the Pennsylvanian Amsden. In addition, there are several thin-bedded shales and argillites in the Missoula Group which might act as weak layers for treads.

Back thrusting may have occurred in this ramp model by the formation of reverse faults and their complimentary westward-verging folds in the synclinal hinge area of the ramp (Serra, 1977) (Fig. 35). The reverse faults are a necessary mechanism by which the upper plate deformed and migrated past the ramp region (Morse, 1977).

A possible correlation to this hypothesis was suggested by Achuff (1981) who observed a similar system of west-directed folds and thrust faults in the Northern Blacktail Range of southwestern Montana. His work showed that east-directed compression may have led to eastward thrusting and consequent back thrusting along the west side of the Blacktail-Snowcrest uplift block.

Positive aspects of this model are: 1) the location of this area in the midst of the Cordilleran fold and thrust belt where regional décollements and ramps are common (Dahlstrom, 1970), 2) the suggestion that movement of the Deer Lodge block took place along a major décollement in the Belt embayment (Winston, et al; 1982), 3) the suggested presence of a crystalline buttress beneath the Flint Creek

Range, 4) the presence of both competent and incompetent rocks capable of producing ramps and treads, and 5) the overall anticlinal nature of the eastern province which may be positioned over a ramp (Fig. 34). These arguments lend viability to the ramp-and-tread model. Therefore, the author suggests that the regional decollement-ramp model is responsible for the structural geometry of the eastern province.

Regardless of the mode by which the eastern province was formed, it experienced deformation concurrent with the deposition of the Golden Spike Formation 78 m.y. ago (Table 1). However, gentle tilting of the Golden Spike strata indicates that mild deformation continued for some time after the Golden Spike was deposited.

Western Province Development

The western province represents the eastern "toe" of the Sapphire tectonic block, as discussed previously.

Hyndman, et al, (1975) concluded that the Sapphire tectonic block became detached from the roof of the Idaho batholith and moved more than 60-65 kilometers to the east via large-scale thrust faults (Fig. 36). They suggested gravity sliding as the mechanism by which this transport occurred.

The author suggests that intrusive wedging and consequent volume problems may have been responsible for this movement. Furthermore, it is suggested that the Sapphire block is the trailing edge of the Deer Lodge block (Fig. 34). In this scenario, intrusive wedging may have caused the trailing edge of the Deer Lodge block to thrust eastward

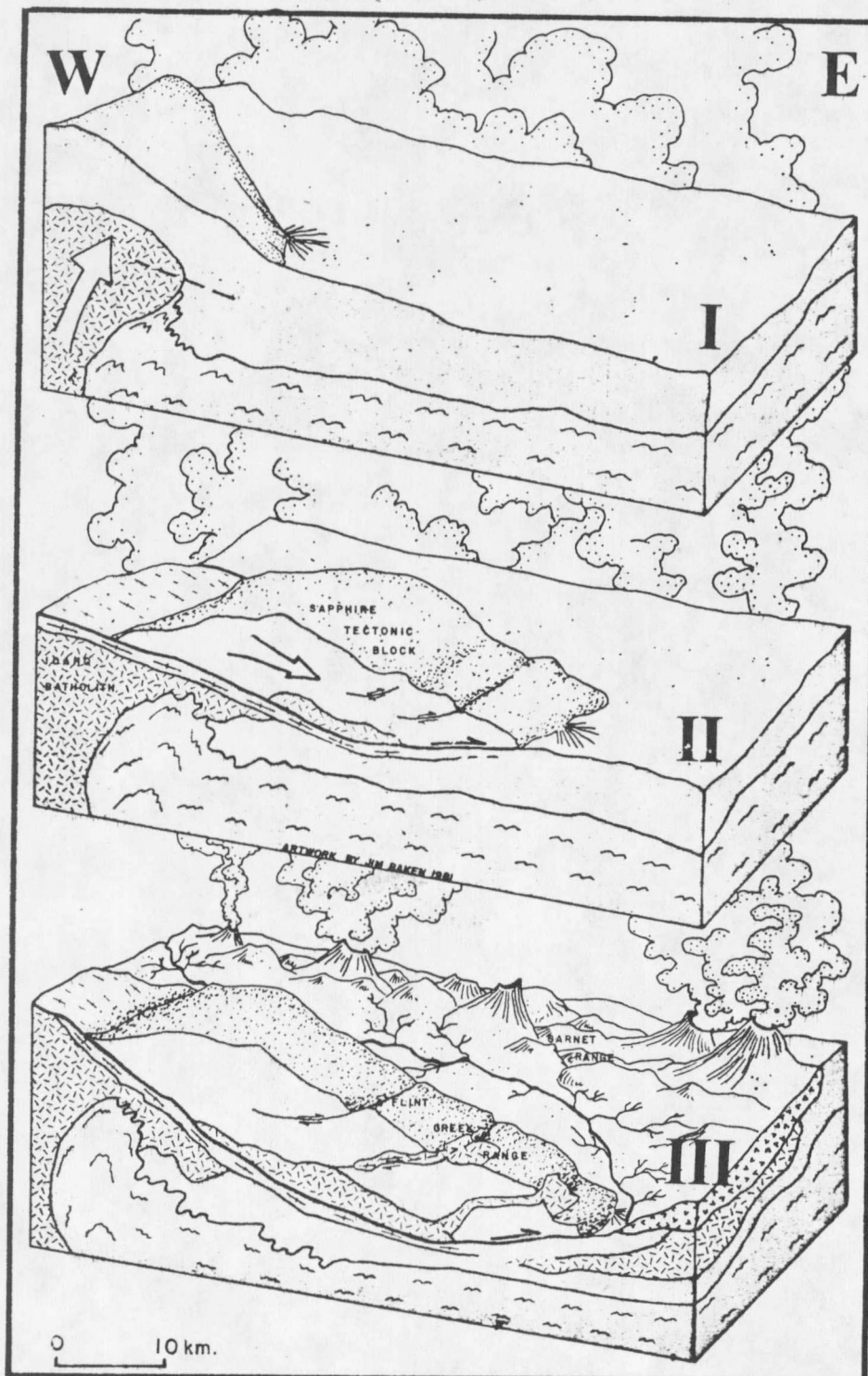


Figure 36. Development of the Sapphire tectonic block. An artist's conception of the allchthon development 78-76 m.y. ago. Intrusive wedging by the Idaho batholith may have assisted in the process (modified after Hyndman, Talbot, & Chase, 1975).

upon itself along the Philipsburg and Georgetown-Princeton thrust faults. This created the Sapphire tectonic block, a distinct, new structural terrane. The Philipsburg and Georgetown-Princeton thrusts probably "sole out" into the regional decollement (Fig. 34).

Hyndman (1980) suggested that the allochthon moved between 78-76 m.y. ago (Table 1). Several pieces of evidence determine the timing of block-movement. The Beaverhead Conglomerate in southwest Montana contains rock clasts from above the rising batholith which indicate the beginning of uplift 100 m.y. ago continuing to 60 m.y. ago, as discussed previously (Ryder and Scholten, 1973). Hyndman (1980) provided a date of emplacement for the batholith of 90-70 m.y. (Table 1). The Golden Spike Formation, which is dated by 78 m.y. old Elkhorn Mountains Volcanics clasts, contains rock allochthonous to the Sapphire block and is cut by its leading edge (Gwinn and Mutch, 1965; and Hyndman, 1979 and 1980). The western and eastern Flint Creek plutons are only slightly deformed by the leading edge of the block and are dated at 76.7-74 m.y. and 73.4-72 m.y., respectively (Hyndman, 1980). However, the earlier, undated, Racetrack Creek pluton is deformed by faulting.

Several geologists have presented evidence which contradicts Hyndman's hypothesis of gravity sliding by the Sapphire tectonic block (Garmezy and Greenwood, 1981; and Toth 1981). These workers suggested that Hyndman's mylonite zone was actually produced by the flow of magma within the chamber as a crystal-mush (in contact with the surrounding country rock). Garmezy and Greenwood presented evidence supporting this theory with a description of flow characteristics

within the mylonite zone. In addition, they described curvature of the zone around the south end of the Bitterroot Lobe. They cited this evidence to suggest a theory of progressive injection and flowage of crystal-mush to account for the development of the mylonite zone.

Hyndman (1980); however, had previously stated:

"...the mylonitic deformation appears to be an original feature which expresses the movements that created it. If mylonitization were related to doming proper, the strong fabric would presumably survive in the granites along its western flank as well as in those along its eastern flank." (p. 433).

The mylonite zone is, in fact, concentrated on the east flank. Hyndman concludes that this fabric must have formed under shallow, brittle conditions provided by the unloading of the Sapphire block.

Winston, et al, (1982) provided additional evidence of block movement with stratigraphic data. They compared measured stratigraphic sections on both sides of the leading edge of the Sapphire block and observed abrupt thickening of several thousand feet in the Lower Missoula Group and Wallace Formation. This evidence appears to substantiate that the two areas were once widely separated.

In addition, the Eyebrow Mountain klippe provides clear evidence of eastward thrusting after formation of the eastern province as discussed in the previous section.

Hyndman (1980) suggested that the Sapphire tectonic block may have "bulldozed" the uppermost Precambrian to Mesozoic rocks creating a crumpled zone (the eastern province) which parallels the eastern edge of the block. The author does not concur with "bulldozing" as

the sole force which created the structural geometry of the eastern province. Basically, it does not explain the westward asymmetry satisfactorily. However, the eastern and northern provinces do display a zone of arcuate structural trend change parallel to the leading edge of the Sapphire block. This indicates that the Sapphire allochthon may have had a minor "bulldozing" effect on the pre-existing structure.

Igneous Activity - Phase II

East of the Flint Creek Range the Deer Lodge block ceased movement along the Lombard thrust and was cut by the final emplacement of the Boulder batholith 78-72 m.y. ago (Table 1) (Hyndman, 1980; Woodard, 1981; and Winston, et al, 1982). In the Flint Creek Range the western and eastern Philipsburg plutons intruded along thrust faults 76.7-74 m.y. and 73.4-72 m.y., respectively.

Middle and Late Cenozoic History

The Late Cretaceous to Eocene compressional period was followed by Cenozoic extension.

The Cenozoic phase of extension commenced with movement along the north-south trending Bungalow, Douglas Mountain, and Wasa-Dunkleberg faults. Concurrently, the northeast trending faults in the northeastern part of the range were formed marking the end of compression within the Deer Lodge and Sapphire blocks.

The east-west, high-angle faults in the Clark Fork Valley experienced movement with a tentative scenario involving: 1) uplift

along the faults in the Oligocene and Miocene (causing the migration of the centers of deposition toward the center of the valley) and 2) acceleration of uplift and tilting during the Pliocene (creating the variation in dips in the Cabbage Patch Beds). These relationships were discussed previously and the tentative nature of the evidence causes the conclusions to be questionable. Later, the north- to northwest-trending faults of Schmidt and Hendrix (1981) may have been reactivated to produce the Deer Lodge and other north-south trending valleys to the east (Winston, et al, 1982). Winston also suggested that these features probably form listric faults at depth.

Finally, the present cycle of erosion and deposition is due to regional epeirogenic uplift of the Rocky Mountain region.

ECONOMIC GEOLOGY

Introduction

The Flint Creek Range has provided about one percent of Montana's total mineral production with extracted resources worth over one hundred million dollars (Silverman, 1965). This production represents intermittent extraction over most of this century with the most productive years during the base-metal shortages of World War I and II. Today, mining activity in the range is essentially dormant; however, recent increases in mineral prices (notably gold and silver) have spurred filing for mineral-claims.

Recent oil and gas discoveries in the fold and thrust belt of Wyoming and Utah have initiated an increase in exploration and mineral leasing in western Montana, including the Flint Creek Range. The area appears to have oil and gas potential as evidenced by recent drilling activity; however, no commercial discoveries have been made, to date.

Minerals

The most abundant economic minerals occurring in the Flint Creek Range are silver, lead, zinc, gold, iron, manganese, phosphate, and limestone. The most productive region of the range is the Philipsburg district. Primarily, fissure fillings and replacement veins take advantage of minor fault systems on the crest of the Philipsburg anticline, as discussed previously. Other productive areas in the

range include the Dunkleberg district, on the crest of the Dunkleberg anticline; and the Georgetown-Cable; Olson Gulch; Douglas Creek; Maxville; and Princeton districts. The following paragraphs briefly summarize the occurrences of these deposits.

Silver-bearing ores occur as veins in granodioritic rock and replacement veins in Paleozoic calcareous rocks. The granodioritic veins occur in groups of fractures, display fairly uniform width with conspicuous banding, and vary greatly in richness and continuity from vein to vein (Silverman, 1965). The ore is composed of pyrrargyrite, proustite, native silver, tetrahedrite-tennantite, galena, sphalerite, pyrite, arsenopyrite, quartz, rhodochrosite, and calcite (Silverman, 1965). Replacement veins have a similar mineralogy and occupy bedding-plane faults as well as fracture zones. These veins have great horizontal and vertical continuity and appear to be richer where they cut purer carbonate rocks (Silverman, 1965). In the Dunkleberg district the Forest Rose and Wasa mines have produced lead, zinc, and silver from these veins.

Gold-bearing veins in granodiorite have not been very productive; however, a small amount has been produced on Boulder Creek near Georgetown. These veins generally display a lode character and are discontinuous along strike and dip. The mineralogic assemblage includes quartz, pyrite, native gold, galena, chalcopyrite, and calcite (Silverman, 1965). Replacement veins rich in gold occur primarily in the Hasmark Formation near the contact with intrusives. Important ore and gangue minerals include pyrrhotite, magnetite,

hematite, native gold, quartz, siderite, and calcite (Silverman, 1965).

Gold placer deposits have been intermittently reworked in the past 100 years; however, the coarseness of the gravel has increased the cost of extraction to the point of unprofitability. The author encountered several independent prospectors who had filed claims on placer deposits within the area. Their operations had met limited noncommercial success.

Iron is primarily derived from magnetite in contact-metamorphic deposits in limestone. These occurrences are found near the contact with granodiorite. The ore varies in iron content from 30-65 percent Fe (Silverman, 1965).

Manganese oxide deposits in the Flint Creek Range are near associated precious-base metal ores (Silverman, 1965). Silverman notes that these deposits are the result of secondary, near-surface oxidation of large, primary masses of rhodochrosite.

The most important non-metallic mineral resource in the range is phosphate. The Phosphoria Formation provides the ore-bearing horizon. Douglas Creek and Maxville contain the most-easily accessible deposits due to breaching of the Princeton and Mount Douglas anticlines by Boulder and Douglas Creeks. High-grade phosphate beds are generally 1 to 4 feet thick and contain 32-36 percent P_2O_5 (Silverman, 1965).

Metallurgical limestone has been mined near Anaconda to be used in copper-reduction processes (Silverman, 1965). This high-quality rock is from the Mission Canyon Member of the Madison Formation.

Oil and Gas

The oil and gas potential of the area is still unknown. At the time of publication, several wells had been drilled in Sec. 9, T. 10 N., R. 13 W. by Nyvatex Oil. Information is scarce; however, it appears that the area is plagued with poor drilling conditions and a lack of shows. Amoco Oil has drilled two holes in the Deer Lodge Valley in Secs. 2 and 31, T. 7 N., R. 10 W. The company has not released any information.

In the author's opinion the area surrounding the Flint Creek Range has potential for commercial reserves of oil and gas. The rocks and structure of the range are advantageous for trapping hydrocarbons. Figure 6 and Plate 1 show many favorable structures. The Colorado, Kootenai, Jurassic, Quadrant (in places), Madison, Hasmark, and Missoula Group rocks provide possible reservoirs while the Coberly, Flood, Kootenai, Phosphoria, Madison, Jefferson, and Hasmark rocks are possible source rocks. However, within the range, metasomatic fluids and the temperatures associated with the nearby Flint Creek plutons preclude the presence of oil and gas.

SUMMARY

The Flint Creek Range, located in west-central Montana, displays the entire stratigraphic record from the Precambrian Ravalli Formation to Tertiary and Quaternary strata. In addition, Cretaceous intrusives, sills, and contact-metamorphosed rocks occur as the result of the Flint Creek plutons and their effects. In general, the western half of the range is characterized by Precambrian and Lower Paleozoic outcrops. The eastern and northeastern parts of the range primarily display Middle Paleozoic to Mesozoic strata. The south end of the range contains the Flint Creek plutons.

The Flint Creek Range lies within the Cordilleran fold and thrust belt, but cannot be easily assigned to the "infra-" or "suprastructural" subdivisions.

Locally, the range has been divided into three structural provinces. The western province represents the eastern edge of the Sapphire tectonic block. This terrane is characterized by eastward asymmetry displayed in broad, relatively undeformed folds and large-scale, west-dipping thrust faults. The eastern border is marked by the Georgetown-Princeton thrust. The folds within the province consistently display eastward asymmetry while the thrust faults reveal significant eastward transport.

The eastern province occupies the north and northeastern parts of the range. The limits of this terrane are marked by the Georgetown-Princeton thrust on the west, Cenozoic-filled valleys to the north and

east, and the Flint Creek plutons to the south. This province displays "atypical" westward asymmetry characterized by tight, commonly overturned, folds; small- and moderate-scale, east-dipping thrust faults; north-south trending, high-angle faults; and several, high-angle, northeast-trending faults. The structures within this province show clear evidence of west-directed transport and subsequent deformation.

The northern province occupies the area of the Clark Fork River Valley. This terrane is characterized by flat-lying Tertiary strata and represents a segment of the Montana lineament. The area is represented by a structural trend change and opposing fold plunges where the structures of the Flint Creek and Garnet Ranges meet across the valley.

Surrounding the Flint Creek Range are several regional tectonic features which have assisted in the production of the structural geometry displayed in the range. The Sapphire tectonic block to the west is a distinct structural terrane marked on three sides by detachment and fault zones. The structure within the allochthon is characterized by limited deformation in the west and large-scale thrusting in the east.

The Boulder batholith is a quartz monzonite to granodiorite igneous body east of the Flint Creek Range, which intruded into the Belt, Paleozoic, Mesozoic, and Elkhorn Mountains Volcanics. The batholith lies in a structural basin and is separated from the Flint Creek Range by the block-faulted Deer Lodge Valley.

The Montana lineament is a regional line recognized by diversely aligned geologic elements along its length. A portion of this feature is represented by the northern structural province.

The Jefferson Valley lineament is an east-west trending zone of crustal weakness. The lineament extends west from the Highland Mountains to the southern edge of the Flint Creek Range and is characterized by the abrupt disappearance of a wedge of coarse-grained arkoses and arkosic conglomerates in the Highlands.

The Belt embayment is a feature which is recognized by stratigraphic data. The Belt strata indicate the presence of an east-west Precambrian basin and are marked to the north and south by the Montana and Jefferson Valley lineaments.

The history of deformation in the Flint Creek Range (illustrated in Table 1) began 100 m.y. ago with the doming of a tectonic highland west of the range. This uplift was the result of the initial stages of emplacement of the Idaho batholith. Intrusive wedging by this body combined with east-west regional compression to deform the rocks occupying the Belt embayment which caused the detachment of the Deer Lodge block. This allochthon is bounded by the Montana and Jefferson Valley lineaments to the north and south, and the Lombard thrust to the east.

Following the development of this allochthon, the Boulder batholith initiated emplacement along the Deer Lodge-block decollement. Satellitic effects of this were the intrusion of the undated Racetrack Creek pluton in the Flint Creek Range and the Elkhorn Mountains Volcanics to the east (78-72 m.y.). The probable

magma source for these rocks was the Idaho batholith. In the meantime, the Golden Spike Formation was being deposited at the future site of the Flint Creek Range (78 m.y. ago).

During this time the structural geometry displayed in the eastern province was created. Some authors have debated over the possibility that forceful emplacement of the Boulder batholith was responsible for this structure. This author suggests that the concept has merit. However, a more viable theory involves the development of a ramp along the Deer Lodge-block decollement beneath the Flint Creek Range. Movement of the Deer Lodge block along this ramp created a zone of back thrusting and west-directed transport with the resultant westward asymmetry in the eastern province.

Approximately 78-76 m.y. ago the Sapphire tectonic block formed as the result of intrusive wedging adjacent to the Idaho batholith. This allochthon (once the trailing edge of the Deer Lodge block) thrust eastward upon the eastern province of the Flint Creek Range along the Philipsburg and Georgetown-Princeton thrusts. These thrusts are believed to "sole out" into the Deer Lodge-block decollement.

Following thrusting of the Sapphire block, the Deer Lodge block ceased movement and was cut by the final intrusion of the Boulder batholith (78-72 m.y. ago). In the Flint Creek Range the western and eastern plutons intruded along preexisting thrust faults (76.7-74 m.y. and 73.4-72 m.y., respectively).

Cenozoic extension completed the deformation in the range. North-south trending, high-angle faults developed concurrent with northeast trending, high-angle faults in the northeast part of the

range. Finally, the Garnet and Flint Creek Ranges experienced uplift along tentative, covered, east-west, high-angle faults.

The present cycle of erosion and deposition is due to regional eperogenic uplift.

REFERENCES CITED

- Achuff, J.M., 1981, Analysis of faulting and folding in the northern Blacktail Range, southwestern Montana: Geological Society of America Abstracts with Programs, p. 189.
- Allen, J.C., 1966, Structure and petrology of the Royal Stock, Flint Creek Range, central-western Montana: Geological Society of America Bulletin, v. 77, p. 291-301.
- Alexander, R.G., 1951, Geology of the Whitehall area, Montana: Ph.D. dissertation, Princeton University, Princeton, New Jersey.
- Armstrong, R.L., and Suppe, J., 1973, Potassium-Argon geochronometry of Mesozoic igneous rocks in Nevada, Utah, and southern California: Geological Society of America Bulletin, v. 84, p. 1375-1392.
- _____, 1974, Magmatism, orogenic timing, and orogenic diachronism in the Cordillera from Mexico to Canada: Nature, v. 247, p. 348-351.
- Baken, J.F., 1981, Structural geology and tectonic history of the northern Flint Creek Range, west-central Montana: in Montana Geological Society Guidebook, Southwest Montana Field Conference, p. 161-166, map in pocket.
- Biehler, S., and Bonini, W.E., 1969, A regional gravity study of the Boulder batholith, Montana: in Larsen, L., and others, eds., Igneous and metamorphic geology: Geological Society of America Memoir 115, p. 401-422.
- Burchfiel, B.C., and Davis, G.A., 1968, Two-sided nature of the Cordilleran orogen and its tectonic implications: International Geological Congress, 23rd, Prague 1968, Report Section, 3, Proclamation, p. 175-184.
- _____, _____, 1972, Structural framework and evolution of the southern part of the Cordilleran orogen, western United States: American Journal of Science, v. 272, p. 97-118.
- _____, _____, 1975, Nature and controls of Cordilleran orogenesis, western United States: Extensions of an earlier synthesis: American Journal of Science, v. 275-A, p. 363-396.
- Calkins, F.C., and Emmons, W.H., 1915, Description of the Philipsburg quadrangle, Montana: United States Geological Survey Atlas, Folio 196.

- Csejtey, B., 1963, Geology of the southeast flank of the Flint Creek Range, western Montana: Ph.D. dissertation, Princeton University, Princeton, New Jersey, 175 p.
- Dahlstrom, C.D.A., 1970, Structural geology in the eastern margin of the Canadian Rocky Mountains: Bulletin Canadian Petroleum Geology, v. 18, no. 3, p. 332-406.
- Earl, F.N., 1972, Mines and mineral deposits of the southern Flint Creek Range, Montana: Montana Bureau of Mines and Geology Bulletin 84.
- Ekleman, W. R., and Kulp, J.L., 1957, Uranium-lead method of age determination, Part II, North American localities: Geological Society of America Bulletin, v. 68, p. 1117-1140.
- Emmons, W.H., 1907, The Granite-bimetallic and Cable mines, Philipsburg quadrangle, Montana: United States Geological Survey Bulletin 315-A, p. 31-55.
- _____, and Calkins, F.C., 1913, Geology and ore deposits of the Philipsburg quadrangle, Montana: United States Geological Survey Professional Paper 78, 271 p.
- Fields, R.W., and Shepard, W., eds., 1965, Geology of the Flint Creek Range, Montana: Billings Geological Society Guidebook, 16th Annual Field Conference, 188 p.
- Flood, R.E., 1974, Structural geology of the upper Fishtrap Creek Area, central Anaconda Range, Montana: unpublished M.A. thesis, University of Montana, Missoula, Montana, 71 p.
- Garmezy, L., and Greenwood, W.R., 1981, Mylonitization coincident with batholith emplacement: Bitterroot lobe, Idaho batholith: Geological Society of America Abstracts with Programs, p. 197-198.
- Goddard, E.N., 1940, Manganese deposits at Philipsburg, Granite County, Montana: United States Geological Survey Bulletin 922-G, p. 157-204.
- Gwinn, V.E., 1960, Cretaceous and Tertiary stratigraphy and structural geology of the Drummond area, western Montana: Ph.D. dissertation, Princeton University, Princeton, New Jersey.
- _____, and Mutch, T.A., 1960, Westward overturning and thrusting in the Flint Creek Range, Montana (Abstract): Geological Society of America Bulletin, v. 71, p. 1934.

- _____, _____, 1965, Intertongued Upper Cretaceous volcanic and nonvolcanic rocks, central-western Montana: Geological Society of American Bulletin, v. 76, no. 10, p. 1125-1144.
- Hamilton, W., and Meyers, W.B., 1966, Cenozoic tectonics of the western United States: Review of Geophysics and Space Geophysics, v. 4, p. 509-549.
- _____, _____, 1974a, Nature of the Boulder batholith of Montana: Geological Society of America Bulletin, v. 85, p. 365-378.
- _____, _____, 1974b, Nature of the Boulder batholith: Reply: Geological Society of America Bulletin, v. 85, p. 1958-1960.
- Harris, S.A., 1957, The tectonics of Montana as related to the Belt Series: in Billings Geological Society Guidebook, Crazy Mountain Basin, 8th Annual Field Conference, p. 22-33.
- Harrison, J.E., 1972, Precambrian belt basin of northwestern United States—Its geometry, sedimentation, and copper occurrences: Geological Society of America Bulletin, v. 83, no. 5, p. 1215-1240.
- _____, Griggs, A.B., and Wells, J.D., 1974, Tectonic features of the Precambrian Belt basin and their influence on post-Belt structures: United States Geological Survey Professional Paper 866, 15 p.
- Hawley, K., 1975, The Racetrack pluton—A newly defined Flint Creek pluton: Northwest Geology, v. 4, p. 1-8.
- Hobbs, B.E., Means, W.D., and Williams, P.F., 1976, An outline of structural geology: John Wiley and Sons, Inc., New York, New York.
- Hoffman, J., Hower, J., and Aronson, J.L., 1976, Radiometric dating of the time of thrusting in the disturbed belt of Montana: Geology, v. 4, no. 1, p. 16-20.
- Hopson, C.A., and Mattinson, J.M., 1973, Ordovician and Late Jurassic ophiolitic assemblages in the Pacific Northwest (Abstract): Geological Society of America Abstracts with Programs, v. 5, no. 1, p. 57.
- Hyndman, D.W., 1972, Petrology of igneous and metamorphic rocks: McGraw and Hill Book Company, New York, New York, 533 p.

- _____, 1977, Mylonitic detachment zone and the Sapphire tectonic block, in Mylonite detachment zone, eastern flank of the Idaho batholith: Geological Society of America Rocky Mountain Section Annual Meeting, Field Guide No. 1, p. 25-31.
- _____, 1979, Major tectonic elements and tectonic problems along the line of section from northeast Oregon to west-central Montana: Geological Society of America, MC-28C, 11 p.
- _____, 1980, Bitterroot dome-Sapphire tectonic block, an example of a plutonic-core gneiss-dome complex with its detached suprastructure; in Coney, P.J., and others, eds., Cordilleran metamorphic core complexes: Geological Society of America Memoir 153 (in press).
- _____, Obradovich, J.D., and Ehinger, R., 1972, K-Ar age determinations of the Philipsburg batholith: Geological Society of America Bulletin, v. 83, p. 473-474.
- _____, Talbot, J.L., and Chase, R.B., 1975, Boulder batholith: A result of emplacement of a block detached from the Idaho batholith infrastructure?: Geology, p. 401-404.
- _____, et al, 1976, Petrology of the Philipsburg batholith, Flint Creek Range, western Montana-Satellitic bodies of the Boulder and (?) Idaho batholiths: Montana Bureau of Mines and Geology Open File Report, 90 p.
- Klepper, M.R., Weeks, R.A., and Ruppel, E.T., 1957, Geology of the southern Elkhorn Mountains, Jefferson and Broadwater Counties, Montana: United States Geological Survey Professional Paper 292, 82 p.
- _____, Robinson, G.D., and Smedes, H.W., 1971, On the nature of the Boulder batholith of Montana: Geological Society of America Bulletin, v. 82, p. 1563-1580.
- _____, _____, _____, 1974, Nature of the Boulder batholith of Montana: discussion: Geological Society of America Bulletin, v. 85, p. 1953-1960.
- Konizeski, R.L., McMurtrey, R.G., and Brietkrietz, A., 1961, Preliminary report on the geology and groundwater resources of the northern part of the Deer Lodge Valley, Montana: Montana Bureau of Mines and Geology, Bulletin 21.
- Lanphere, M.A., and Reed, B.L., 1973, Timing of Mesozoic and Cenozoic plutonic events in circum-Pacific North America: Geological Society of America Bulletin, v. 84, p. 3773-3782.

- McGill, G.E., 1958, Geology of the northwest flank of the Flint Creek Range, western Montana: Ph.D. dissertation, Princeton University, Princeton, New Jersey.
- _____, 1959, Northwest flank of the Flint Creek Range, Montana: Montana Bureau of Mines and Geology, Geologic Map Series, no. 3.
- _____, 1965, Tectonics of the northern Flint Creek Range, Montana: in Billings Geological Society Guidebook, 16th Annual Field Conference, p. 127-136, map in pocket.
- McMannis, W.J., 1963, LaHood Formation—a coarse facies of the Belt Series in southwestern Montana: Geological Society of America Bulletin, v. 74, p. 407-436.
- _____, 1965, Resume of depositional and structural history of western Montana: American Association of Petroleum Geology Bulletin, v. 49, p. 1801-1823.
- Mertie, J.B., Jr., Fischer, R.P., and Hobbs, S.W., 1951, Geology of the Canyon Ferry quadrangle, Montana: United States Geological Survey Bulletin 972, 97 p.
- Morse, J., 1977, Deformation in ramp regions of overthrust faults: Experiments with small-scale rock models: in Wyoming Geological Association Guidebook, 29th Annual Field Conference, p. 457-470.
- Mutch, T.A., 1960, Geology of the northeast flank of the Flint Creek Range, Montana: Ph.D. dissertation, Princeton University, Princeton, New Jersey.
- _____, 1961, Geology of the northeast flank of the Flint Creek Range, western Montana: Montana Bureau of Mines and Geology, Geologic Map Series no. 5.
- Nelson, W.H., and Dobell, J.P. 1961, Geology of the Bonner quadrangle, Montana: United States Geological Survey Bulletin 1111-F, p. 189-235.
- Popoff, C.C., 1953, Lead-zinc deposits of the Dunkleberg district, Granite County, Montana: United States Bureau of Mines Report of Investigations 5014, 49 p.
- Poulter, C.J., 1957, Geology of the Georgetown thrust area southwest of Philipsburg, Montana: Ph.D. dissertation, Princeton University, Princeton, New Jersey.
- _____, 1957, Geology of the Georgetown thrust area southwest of Philipsburg, Granite, and Deer Lodge Counties, Montana: Montana Bureau of Mines and Geology, Geologic Map Series no. 1.

- _____, 1959, Structural synthesis of an area in southeastern Granite County, Montana: in Billings Geological Society Guidebook, 10th Annual Field Conference, p. 22-33.
- Price, R.A., and Mountjoy, E.W., 1970, Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca Rivers- A progress report: Geological Association Canada Special Paper, no. 6, p. 7-25.
- Prinz, W.C., 1967, Geology and ore deposits of the Philipsburg district Granite County, Montana: United States Geological Survey Bulletin 1237, p. 1-66.
- Rich, J.L., 1934, Mechanics of low-angle overthrust faulting illustrated by Cumberland thrust block, Virginia, Kentucky, and Tennessee: American Association of Petroleum Geology Bulletin, v. 18, p. 1584-1596.
- Robinson, G.D., 1959, The disturbed belt of the Sixteen mile area, Montana: in Billings Geological of Society Guidebook, 10th Annual Field Conference, p. 24-33.
- _____, 1961, Origin and development of the Three Forks Basin, Montana: Geological Society of America Bulletin, v. 72, p. 1003-1014.
- _____, 1967, Geologic map of the Toston quadrangle, southwestern Montana: United States Geological Survey Miscellaneous Geological Investigations, Map I-486, scale 1:24,000.
- _____, Klepper, M.R., and Obradovich, J.D., 1968, Overlapping plutonism, volcanism and tectonism in the Boulder batholith region, western Montana: Geological Society of America, Studies in volcanism, Memoir 116, Coats, R.R., Hay, R.L., and Anderson, G.A., eds., p. 557-576.
- Ross, C.P., 1959, Geology of Glacier National Park and the Flathead region, northwestern Montana: United States Geological Survey Professional Paper 296, 125 p.
- Royse, F., Jr., Warner, U.A., and Reese, D.L., 1975, Thrustbelt structural geometry and related stratigraphic problems, Wyoming-Idaho-Northern Utah: Rocky Mountain Association Geological Guidebook, Deep Drilling Frontiers in the Central Rocky Mountains, p. 41-54.
- Ruppel, E.T., Wallace, C.A., Schmidt, R.G., and Lopez, D.A., 1981, Preliminary interpretation of the thrust belt in southwest and west-central Montana and east-central Idaho: in Montana Geological Society Guidebook, Southwest Montana Field Conference, p. 139-159.

- Ryder, R.T., and Scholten, R., 1973, Syntectonic conglomerates in southwestern Montana: Their nature, origin and tectonic significance: Geological Society of America Bulletin, v. 84, p. 773-796.
- Sahinen, U.M., 1939, Geology and ore deposits of the Rochester and adjacent mining districts, Madison County, Montana: Montana Bureau of Mines and Geology, Memoir 19.
- Schmidt, C.J., and Hendrix, T.E., 1981, Tectonic controls for thrust belt and Rocky Mountain foreland structures in the northern Tobacco Root Mountains-Jefferson Canyon area, southwestern Montana: in Montana Geological Society Guidebook, Southwest Montana Field Conference, p. 167-180.
- Scholten, R., 1960, Sedimentation and tectonism in the thrust belt of southwestern Montana and east-central Idaho: in Wyoming Geological Association Guidebook, 15th Annual Field Conference, p. 73-83.
- _____, 1968, Model for the evolution of the Rocky Mountains east of the Idaho batholith: Tectonophysics, Elsevier Publishing Company, Amsterdam, p. 109-126.
- Serra, S., 1977, Styles of deformation in the ramp regions of overthrust faults: in Wyoming Geological Association, 29th Annual Field Conference, p. 487-498.
- Silbering, N.J., 1973, Geologic events during Permian-Triassic time along the Pacific margin of the United States: in Logan, A., and Hills, L.V., eds., The Permian and Triassic systems and their mutual boundary, Alberta Society of Petroleum Geology, Calgary, Alberta, p. 345-362.
- Silverman, A., 1965, Economic geology of the Flint Creek Range: in Billings Geological Society Guidebook, 16th Annual Field Conference, p. 108-119.
- Skipp, B., and Peterson, A.D., 1965, Geologic map of the Maudlow quadrangle, southwestern Montana: United States Geological Survey Miscellaneous Geologic Investigations Map I-452, scale 1:24,000.
- Smedes, H.W., 1973, Regional tectonic setting of the Boulder batholith (Abstract): in Guidebook for the Butte field meeting of the Society of Economic Geology, Butte, Montana, p. A.
- Stewart, J.H., 1972, Initial deposits in the Cordilleran geosyncline: Evidence of a Late Precambrian (≈ 850 m.y.) continental separation: Geological Society of America Bulletin, v. 83, p. 1345-1360.

- Toth, M.I., 1981, Petrology and structure of the east-central Bitterroot lobe of the Idaho batholith: Geological Society of America Abstracts with Programs, p. 228.
- Wallace, C.A., Klepper, M.R., French, A.B., and Scarborough, D.M., 1978, Preliminary map of the northwestern part of the Butte 1° x 2° quadrangle, Montana: United States Geological Survey open-file report no. 78-371.
- Wanek, A.A., and Barclay, C.S.V., 1966, Geology of the northwest quarter of the Anaconda quadrangle, Deer Lodge County, Montana: United States Geological Survey Bulletin 1222-B, p. B1-B28.
- Weidman, R.M., 1965, The Montana lineament: in Billings Geological Society Guidebook, 16th Annual Field Conference, p. 137-143.
- Winston, D., Jacob, P., Baldwin, D.O., and Reid, J.P., 1982, Proterozoic block faulting in the Belt basin, Montana and Idaho: Its effect on Rocky Mountain thrusting and basin and range extension (Manuscript).
- Woodward, L.A., 1981, Tectonic framework of disturbed belt of west-central Montana: American Association of Petroleum Geology Bulletin, v. 65, p. 291-302.

APPENDIX
STRATIGRAPHY

Introduction

The stratigraphic units on Table 2 are grouped by eras to simplify the structure for mapping and visual display purposes. The scope of this study did not include stratigraphic characterization; therefore, the following are in the form of brief, summary statements. The rocks are described from personal observations and those of Emmons and Calkins (1913), Poulter (1957), McGill (1959), Gwinn and Mutch (1960), and Gwinn (1961).

Proterozoic

Ravalli Formation

The Ravalli Formation crops out in the western part of the map area in T. 8 N., R. 14 W. (Plate 1). The formation characteristically forms steeply sloping, white, rounded hills typified just west of Flint Creek where U.S. Highway 10A passes through the southeastern corner of the township. These hills support sparse Ponderosa Pine. Emmons and Calkins (1913) reported a thickness for the Ravalli of 2,000 feet south of this area.

The Ravalli Formation is typically a light olive-gray, very-fine-grained quartzite (Poulter, 1957) and slate. Dark micaceous layers occur throughout the formation which define bedding (Poulter, 1957). These layers are more common near the top of the formation and are several millimeters thick while the quartzite beds are up to three feet thick.

Generalized Stratigraphic Section

ERA	PERIOD-EPOCH	TIME M.Y.	GP-FM	MBR	THICK FEET	LITHOLOGY
CENOZOIC	Recent	1.8				talus, alluv., soil, & bk. beds.
	Pleistocene					glcl. till, Buffalo(?), Bull Lake(?), & Pinedale
	Pliocene					semi-consol. cgl, ss, siltst & clay
	Miocene					fluv. silt, sd, & pbl. cgl buffa. vol.-rich siltst & f-spathic cgl & ss, minor ls & diatomite
MESOZOIC	Cretaceous	65	COLORADO GP	GOLDEN SPIKE	4000-6000	and-dac vol s & vol.-rich ss, siltst, & cgl
				CARTER OK	4500-6000	tan-gy, chrt-pbl-rich ss, siltst, & mdst
				JENS	1000-1810	dk-gy sh & siltst
				COBERLY	570-650	gy-blk, gss. ls & tan, S&P, ss
				DUNKLEBERG	1500-1700	chrt-pbl cgl & vol.-rich ss, siltst, porcellanite bds
				BLACK-LEAF	900-1000	tan to lt-gy, ss, siltst, & mdst
				TAFT HILL	1000	
				FLOOD	680-700	dk-gy to blk sh & fn-gnd ss gy, org. ls, sh & siltst
				UPPER LS	80-130	gy, med- to cse-Xln ls, Gastropod
				UPPER DET	400-500	maroon & gy mdst & siltst, S&P ss, & ls-pbl cgl
				LOWER LS	150-220	gy, fn-Xln ls, w/ siltst & mdst
				LOWER DET	80-130	gy-brn, chrt-pbl cgl & S&P ss
	Jurassic	141	ELLIS GP	MORRISON		gy-maroon siltst, sh & S&P ss
				SWIFT		blk, chrt-pbl cgl & calc, S&P ss & siltst
				RIERDON	480	gy, thn-bdd ls, sh & siltst
				SANTOOTH		dk-gy to blk, calc., siltst, sh & ls
PALEOZOIC	Permian	230		SHEDHORN		tan-wh, fn-gnd orthoqtzite
				PHOSPHORIA	175-210	bdd chrt, cgl, ss & siltst; blk phos. mdst; ool. phos.; chrt-brcc. & siltst
				PARK CITY		lt-gy dol & siltst w/ bdd chrt
	Pennsylvanian	280		QUADRANT	350	tan-wh, mass, fn-gnd orthoqtzite
				AMSDEN	298	maroon, fn-lam'd dol & sh
	Mississippian	310		MADISON GP	1000-2300	med- to dk gy, mass.-thk bdd, foss., cse-Xln ls
	Devonian	345		JEFFERSON	864	dk gy, dense ls & dol; choc.-gy, mass- to thk-bdd, cse-Xln dol c ls
				MAYWOOD	275	lt-gy, thn-intbdd dol & siltst
	Cambrian	500		SAGE LIMESTONE		fn-lam d, fn-Xln ls, thn cgl beds
				RED LION	330	dk gy sh
				DRY CREEK SHALE		
				HASMARK	1100	lt-gy to blk, fn Xln dol
				SILVER HILL	330	lt-gy, bndd ls; dk- to lt-gy & maroon, intbdd sh & siltst
				FLATHEAD	75-100	wh-pink, mass. qtzite
PRECAMBRIAN	Proterozoic	570		MISSOULA GP	10,000	lt-pink & maroon, mass. qtzite & thn-bdd sh
				NEWLAND	7000	lt-gy to buff, bndd, sil. ls & sh
				RAVALLI	2000	lt olive-gy qtzite & slate

Table 2. Generalized stratigraphic section.

Near the top of the Ravalli is a 385 foot transitional sequence of black shales and light-colored quartzites which grades upward to interbedded shales and limestones of the overlying Newland Formation.

Newland Formation

The Newland Formation is expressed in gently rounded, light-colored hills which commonly support a cover of prairie grasses. The formation crops out in the southwestern corner of T. 8 N., R. 13 W. on the east side of Flint Creek (Plate 1). Poulter (1957) reported a thickness of 7,000 feet along the continental divide to the south.

The Newland Formation is typically a light-gray to buff, banded, siliceous unit of alternating limestone and shale (Poulter, 1957). The bands of limestone and shale vary from a fraction of an inch to one foot in thickness (Poulter, 1957). The argillaceous layers have a wavy nature and stand in relief on the weathered surface. The carbonate bands are commonly finely crystalline and weather more easily providing contrast with the argillaceous layers.

Poulter (1957) observed that the Newland grades upward into a slightly calcareous argillite with shallow water and desiccation features including a rip-up-clast conglomerate. The top of this transition zone is marked by the maroon color of the overlying Missoula Group.

Missoula Group

The Missoula Group crops out extensively in the western half of the map area on the Georgetown-Princeton thrust plate (Fig. 5). The unit usually occurs as gently rounded hills which vary from heavily-

forested to open grassy slopes with a red soil. Poulter (1957) calculated that the Missoula Group is approximately 10,000 feet thick in the Flint Creek Valley near Georgetown Lake; however, McGill (1959) estimated that these rocks were about 4,400 feet thick on the northwest flank of the Flint Creek Range. Winston (personal comm.) believes that the Missoula Group thins out rapidly to the south and does not appear south of the Jefferson Valley-Willow Creek lineament (Perry line). Poulter (1957) also described this southward thinning.

The Missoula Group varies from light-pink to brick-red and maroon and consists of interbedded massive, cross-bedded quartzites and thin-bedded shales and argillites (Poulter, 1957). The quartzite layers are fine- to coarse-grained and well-sorted with occasional layers of quartz and quartzite-pebble conglomerates (Poulter, 1957). A characteristic feature of this unit is the occurrence of maroon mud-chip layers throughout the strata. The argillaceous layers display desiccation and shallow-water features. The group is commonly marked by intraformational conglomerates (Poulter, 1957).

The Missoula Group is unconformably overlain by the Flathead Formation. The contact is marked by an angular unconformity of 5-15° (Poulter, 1957) and the presence of an occasional quartzite-pebble conglomerate at the base of the Flathead.

Paleozoic

Cambrian

Flathead Quartzite. The Flathead Quartzite occurs in the western part of the map area and is typified in two locations on the

Philipsburg and Georgetown thrust sheets where it crops out as resistant ridges and steep-faced cliffs. The most revealing outcrop of this formation is in a road cut along Boulder Creek in Sec. 15, T. 8 N., R. 13 W. where it is approximately 75-100 feet thick, though Poulter (1957) measured 293 feet near Georgetown Lake (Plate 1). The Flathead thins to the north in this area.

The Flathead Formation is a white to pink quartzite which is massive to thick-bedded with occasional planar cross-bedding (Poulter, 1957). The rock is fine-space to medium-grained with an occasional coarse-grained layer. A quartzite-pebble conglomerate occurs at the base.

There is a twenty-foot transitional zone at the top of the formation where massive quartzites grade upward into the thin-bedded shales and siltstones of the Silver Hill Formation.

Silver Hill Formation. The base of the Silver Hill Formation is readily seen in Boulder Creek in Sec. 15, T. 9 N., R. 13 W. (Plate 1). Unfortunately, the formation is not seen in its entirety anywhere else in the Flint Creek Range (Poulter, 1957). Poulter measured the observable portion of the Silver Hill Formation to be 330 feet thick, which includes a lower unit 95 feet thick and an upper unit 235 feet thick. Emmons and Calkins (1913) reported an uppermost zone 112 feet thick which is covered elsewhere in the Flint Creek Range.

Poulter (1957) described the lower member as a dark greenish-gray and dark maroon (on Boulder Creek) micaceous shale with interbedded light-gray quartzose siltstone. The siltstone occurs as thin (a fraction of a millimeter to several millimeters thick), lenticular

layers interbedded with the shale. The shale occurs in some places in equal portions with the siltstone producing a laminated character; however, the shale usually dominates the unit. The upper member has a banded character consisting of alternating light-gray limestone and buff, argillaceous limestone. The argillaceous limestone bands are a fraction of a centimeter to several centimeters thick and stand in relief on a weathered surface. These bands often display a crinkled character.

Emmons and Calkins (1913) reported that the upper 112 feet of the Silver Hill is composed of banded calcareous shales; however, in this area the unit is covered. Consequently, the transition into the Hasmark Formation has not been observed.

Hasmark Formation. The Hasmark Formation occurs both on the Georgetown-Princeton thrust plate and on the underlying block (Fig. 5). It crops out as cliffs and ridges and is very well exposed north of Boulder Creek in the eastern half of Sec. 15, T. 8 N., R. 13 W. (Plate 1). There are several estimates of thickness for the Hasmark (Emmons and Calkins, 1913; Poulter, 1957; and McGill, 1959), and McGill's (1959) 1,100 feet was used because of its location within the study area.

Poulter (1957) and McGill (1959) described the Hasmark as a light- to medium-gray and black, finely crystalline dolomite. The unit is massive with faint banding. It consists of a lower section 700 feet thick which is very similar to the Jefferson Formation because of its black and medium-gray color. However, Poulter (1957)

noted the presence of light-colored worm trails, siliceous blebs, mottled dolomite and pisolitic beds in the Hasmark Formation. The upper section is 400 feet of light-gray dolomite. The Hasmark serves as a good marker in the Paleozoic because it has a consistently uniform character with a barely-perceptible reaction to cold, dilute hydrochloric acid.

The upper 15-20 feet (McGill, 1959) of the Hasmark is a brick-red, sandy dolomite and sandstone which is disconformable in places with an overlying shale (Poulter, 1957).

Red Lion Formation. The Red Lion Formation does not crop out prominently in the study area; however, McGill (1959) measured 330 feet of the unit. Poulter (1957) described two members in the Red Lion.

Dry Creek Shale Member. The Lower Dry Creek Shale Member is four feet of dark greenish-gray shale which was mentioned in the previous section as disconformably overlying the Hasmark Formation.

Sage Limestone Member. The Upper Sage Limestone Member is a wavy, laminated, finely-crystalline limestone which occurs in massive ledges. The laminated character is produced by alternating layers of buff, maroon, and purple argillaceous and siliceous limestone with light-gray, pure limestone. Thin beds of conglomerate with flat limestone pebbles and coarse calcite beds with intercalated fossil fragments are included in the member (McGill, 1959). McGill reported that these fossils are trilobites and brachiopods dated as Middle-Late Cambrian.

The contact with the overlying Maywood Formation is not well exposed but has been reported as disconformable (Poulter, 1957).

Devonian

Maywood Formation. The Maywood Formation is not well exposed in the study area and is usually represented by swale topography. McGill (1959) measured 275 feet of the unit. The best outcrop of the Maywood is north of Boulder Creek in the southwest quarter of Sec. 14, T. 8 N., R. 13 W. (Plate 1).

McGill (1959) described the Maywood as a light-gray, thinly interbedded, flaggy dolomite and dolomitic siltstone which weathers to yellowish-brown and maroon. Poulter (1957) observed a few shale beds in this unit and planar cross-bedding in the siltstone sections. He also observed that this formation is more massive near the top and bottom. McGill observed that the lower third of the section is composed of thick units of calcareous and dolomitic sandstone.

The contact with the overlying unit is covered.

Jefferson Formation. The Jefferson Formation occurs most prominently near the old mining camp of Princeton on Boulder Creek. The rocks crop out as resistant ridges and cliffs in the center of the north half of T. 8 N., R. 13 W., just north of the road on Boulder Creek (Plate 1). Poulter (1957) measured 864 feet of the Jefferson in the area near Georgetown Lake; however, he noted that the top and the base of the unit were covered.

McGill (1959) described the Jefferson Formation as a dense, dark-gray, medium- to coarse-sucrosic, dolomitic limestone. Poulter noted that the unit is cut by numerous, white calcite veinlets; contains zones of limestone breccia in its upper parts; and has abundant stromatoporids with common echinoid spines, brachiopod fragments and crinoid stems. The formation can be difficult to distinguish from the Madison Formation. However, the Jefferson has black chert bands and a consistent petroliferous odor.

The transition between the Jefferson Formation and the overlying Madison is difficult to distinguish and is consistently marked by a covered shale section.

The Three Forks Formation is not usually recognized in this part of Montana; however, Poulter (1957) measured 445 feet of the unit in the area along the East Fork of Rock Creek. He described an interbedded, grayish-brown calcareous shale and argillaceous limestone sequence. The rock is thin-bedded and has a laminated character. The limestone beds are micritic to very finely crystalline and display common bryozoan and crinoid fragments.

Mississippian

Madison Group. The Madison Group consists of the Lower Lodgepole Formation and the Upper Mission Canyon Formation, but because of poor exposure these formations are considered together as the Madison Group. Nowhere within the map area is the lower part of the Madison Group (Lodgepole Formation) seen; however, fair exposures of the upper part can be seen in the northwest quarter of Sec. 16, T. 8 N., R. 13

W. (Plate 1). This location is on the east flank of the Flint Creek Valley along U.S. Highway 10A. The Madison rocks crop out as light-gray cliffs and resistant ridges which support pine forest and open grassy parks. Poulter (1957) and McGill (1959) both noted that the unit is not well exposed and published approximate thicknesses of 1,000-1,500 feet and 2,300 feet, respectively.

The Madison Group is a medium- to dark-gray, massive to thick-bedded, coarsely-crystalline limestone which weathers to a light, chalky-gray. The rocks are highly fossiliferous with some biostromes of brachiopod and crinoid fragments with an occasional weak, petroliferous odor (Poulter, 1957).

McGill (1959) described the Madison Group in three parts. The basal unit consists of 30 feet of black, conodont-rich shale which is not seen except in mines, test pits and as float. The middle unit is a dark-colored, chert-rich section where chert nodules occur parallel to bedding or as continuous layers several inches thick. The upper 500 feet of the Madison is a light-colored, pure limestone.

The upper part of the Madison Group is not well exposed within the map area and is presumed to be disconformable with the overlying Amsden Formation (Poulter, 1957).

Pennsylvanian

Amsden Formation. The Amsden Formation is very poorly exposed in the map area. Poulter (1957) measured an approximate thickness of 288 feet with both the top and the base covered.

The formation is a maroon dolomite and dolomitic shale with occasional beds of light-gray, micritic limestone and dolomite about three feet thick. The section is characteristically flaggy and finely laminated on weathered surfaces. Scattered, light-gray blebs can be seen throughout the maroon dolomite and shale.

The contact between the Amsden Formation and the overlying Quadrant Formation is not well exposed, but McGill (1959) reported it as gradational.

Quadrant Formation. The Quadrant Formation crops out very well throughout the map area as high resistant walls which are excellent marker beds. The rocks are typified in a road and mining cut north of Douglas Creek in the southeast quarter of Sec. 31, T. 9 N., R. 12 W. where several mine tunnels take advantage of the adjacent phosphate beds (Plate 1). In this location the Quadrant occurs as nearly vertical walls. Emmons and Calkins (1913) measured 350 feet of the unit.

The Quadrant Formation is consistently a tan to light-gray and white, massive, fine-grained orthoquartzite. Porosity is very poor. Bedding is indistinct and jointing is very well developed making the exact bedding attitude difficult to measure; however, the consistency of the contact with the over- and underlying rocks provides evidence of gross attitude. Large talus slopes of blocky boulders and cobbles commonly accompany the orthoquartzite outcrops and characteristically cover the contacts of the adjacent units.

The transition between the Quadrant Formation and the overlying Permian rocks has not been documented; however, the author observed a relatively sharp contact throughout the area.

Permian

Permian System. The Permian section can be subdivided into three different formations; however, lack of good outcrop prevents this use. The Permian rocks crop out most prominently adjacent to the Quadrant Formation in the southeast quarter of Sec. 31, T. 9 N., R. 12 W. where they have been mined for phosphate (Plate 1). McGill (1950) estimated the thickness of the Permian section at approximately 175-210 feet.

McGill (1959) and Gwinn (1961) subdivided the Permian rocks in the Flint Creek Range into the basal Park City Formation, the medial Phosphoria Formation, and the upper Shedhorn Formation.

Park City Formation. The Park City Formation consists of 120 feet of light-gray, silty dolomite and dolomite siltstone with occasional red and gray chert beds.

Phosphoria Formation. The Phosphoria Formation is 64 feet thick and is composed of a lower chert breccia and siltstone; a middle black, phosphatic mudstone and oolitic and ovulitic phospharite unit; and an upper unit of bedded chert, lenticular chert conglomerate, sandstone, and siltstone.

Shedhorn Formation. The Shedhorn Formation consists of 27 feet of tan to white, fine-grained orthoquartzite. The rock commonly displays a pitted texture, limonite spots, and local cross-bedding.

The top of the Permian section is covered; however, the contact appears to be relatively sharp and unconformable (McGill, 1959; and Gwinn, 1961).

Mesozoic

Jurassic

Jurassic System. The Jurassic System is not subdivided for mapping purposes because the rock is not well represented. The rocks are not completely exposed in the map area; however, a fair exposure occurs on the nose of the Princeton Anticline in the south half of Sec. 25, T. 9 N., R. 13 W. (Fig. 5 and Plate 1). McGill (1959) estimated the incomplete Jurassic section at 480 feet thick, which compares to the 690-foot estimation by Gwinn (1961).

Gwinn (1961) subdivided the Jurassic rocks into the lower Ellis Group (Sawtooth, Rierdon, and Swift Formations) and the upper Morrison Formation. Gwinn estimated the thickness of the Ellis Group at 500 feet. McGill (1959), who did not recognize the Rierdon Formation, estimated the thickness at 320 feet. These two authors consequently estimated the thickness of the Morrison Formation at 190 feet and 160 feet respectively, although McGill stated that his Morrison may be part of the underlying Swift Formation. Gwinn (1961) and McGill (1959) described the following formations.

Sawtooth Formation. The Sawtooth Formation unconformably overlies the Permian and is composed of a black to dark-gray, calcareous siltstone and shale with thin interbeds of argillaceous, gray limestone. A sandstone bed occurs at the base.

Rierdon Formation. The middle Rierdon Formation is characterized by gray, thin-bedded, argillaceous limestone, shale, and siltstone which weathers to a tan color.

Swift Formation. The Swift Formation is composed of a black-pebble, chert conglomerate at the base which disconformably overlies the Rierdon. This basal unit lies under a sequence of calcareous, salt-and-pepper sandstone, silty sandstone, and conglomeratic sandstone which varies from thin bedded to massive and commonly displays cross-bedding.

Morrison Formation. The Morrison Formation is a gray to gray-green siltstone with some shaley and sandy beds, which displays a prominent iron oxide stain on the weathered surface. An intercalated layer of salt-and-pepper sandstone occurs near the top of the formation.

The top of the Jurassic System is unconformably truncated by the basal conglomerates of the Kootenai Formation.

Cretaceous

Kootenai Formation. The entire Kootenai Formation is not exposed in the Flint Creek Range; however, a good exposure of a large part of

the unit occurs on the nose of Douglas Mountain anticline in Secs. 29 and 30, T. 9 N., R. 12 W. (Fig. 5 and Plate 1). McGill (1959) estimated the thickness of the Kootenai at 1,255 feet while Gwinn (1961) estimated 900 to 1,100 feet.

Gwinn (1961) subdivided the Kootenai into four members:

Lower Detrital Member. The Lower Detrital Member is 80-130 feet of gray to reddish-brown, salt-and-pepper sandstone and siltstone with silica cement. The base of this member is a consistent marker bed of gray to black, chert-pebble conglomerate which is up to 35 feet thick.

Lower Limestone Member. The Lower Limestone Member is 150-220 feet of gray, finely-crystalline limestone interbedded with varicolored siltstone and mudstone.

Upper Detrital Member. The Upper Detrital Member is 400-500 feet of maroon, green, and gray mudstone and siltstone interbedded with minor gray and brown, calcareous, salt-and-pepper sandstone. One hundred feet from the top of this member is a flat-pebble limestone conglomerate.

Upper "Gastropod" Limestone Member. The upper 80-130 feet of the Kootenai is called the Upper "Gastropod" Limestone Member. This unit is characteristically gray, medium- to coarsely-crystalline limestone. This excellent marker bed is consistently very fossiliferous with recrystallized, fresh-water gastropod shells. The unit is commonly thick-bedded with interbedded pink, calcareous shale and siltstone. Near the top is a calcareous, salt-and-pepper sandstone.

The upper calcareous, detrital part of the Kootenai Formation is transitional with an overlying sandy limestone of the Colorado Group.

Colorado Group. The Colorado Group crops out throughout the northeastern part of the area and is most completely exposed in the trough of the Gold Creek Syncline in T. 9 N., R. 11 and 12 W. (Fig. 5 and Plate 1). Gwinn (1961) noted that the Colorado Group is coarser grained and much thicker (9,170-11,500 feet) in the west. He speculated that this may be due to the nearness of these sediments to the western margin of the Cretaceous seaway. The nomenclature used in the following descriptions is adapted and modified from terminology used in the Sweet Grass Arch and Elkhorn Mountains areas by Gwinn (1961).

Blackleaf Formation. Flood Member. Gwinn (1961) described the Flood Member as a lower "transition" bed 230-250 feet thick and an upper unit 450 feet thick. The lower unit is composed of intercalated gray, argillaceous limestone and tan to gray shales and siltstone. Iron oxide stains are prominent on weathered surfaces. The upper unit is characterized by dark-gray to black, non-calcareous shale with minor fine-grained, calcareous, ripple-laminated sandstone laminae with abundant burrows and trails.

Taft Hill Member. Gwinn estimated the Taft Hill Member to be 900-1000 feet thick composed of tan to light-gray, calcareous sandstone and gray to green siltstone and mudstone. The lower 300

feet of this member is cross-bedded, pyritic sandstone, while the upper 200 feet is volcanoclastic.

Dunkleberg Member. The Dunkleberg Member is 1,500-1,700 feet thick. The type section is in a gulch-network on the southwest side of Saddle Mountain in the east half of Sec. 8, T. 10 N., R. 11 W. (Plate 1) (Gwinn, 1961). Gwinn described the Dunkleberg as a basal chert-pebble conglomerate (400 feet thick) overlain with siliceous, volcanoclastic mudstone, siltstone, and sandstone beds which are calcareous and feldspar-rich. Numerous light-gray, white, and tan porcellanite beds resistant to weathering are diagnostic of this member.

The contact between the Dunkleberg Member and the overlying Coberly Formation is sharp and disconformable (Gwinn, 1961).

Coberly Formation. The Coberly Formation crops out in the trough of the Coberly syncline as relatively resistant sandstone and limestone ridges (Fig. 5). The type section is located in the north half of Sec. 21 and the east half of Sec. 8, T. 10 N., R. 11 W. (Plate 1) (Gwinn, 1961). Gwinn measured the Coberly to be 570 to 650 feet thick.

The Coberly Formation described by Gwinn (1961) and Mutch (1961) is a sequence of interbedded gastropod-rich limestones and salt-and-pepper sandstones. The limestones are gray to black with high- and low-spired snail shells and oyster fragments. These limestones are superficially similar to the Upper "Gastropod" Limestone Member; however, they differ in the variety of fossil forms. The sandstones

are tan, lenticular, calcareous beds which contain little volcanic detritus differentiating them from the sandstones above and below.

The gradational contact between the Coberly Formation and the overlying Jens Formation is marked above the highest sandstone or limestone bed, below the main body of the Jens Formation (Gwinn, 1961).

Jens Formation. The Jens Formation is exposed extensively on the west flank of the Gold Creek syncline; however, the type section is in the extreme north end of the area in the northwest quarter of Sec. 21 and the north quarter of Sec. 20, T. 10 N., R. 11 W. (Fig. 5 and Plate 1) (Gwinn, 1961). The thickness of the Jens as measured by Gwinn varies from 1,000-1,510 feet.

Gwinn (1961) described the Jens Formation as a thick succession of shales and siltstones between the Coberly and Carter Creek Formations. The rocks are dark-gray shale and siltstone. In the west, minor, thin-bedded, silty, siliceous sandstone, siltstone, and silty mudstone are in the middle of the unit.

The contact between the Jens Formation and the overlying Carter Creek Formation is mostly obscure. However, Gwinn described it as disconformable with a basal, pebble-rich sandstone in the Carter Creek.

Carter Creek Formation. The Carter Creek Formation is exposed in the Gold Creek and Carter Creek synclines with a notable exposure of the lower 778 feet in the northeast quarter of Sec. 20, T. 10 N., R. 11, W. (Fig. 5 and Plate 1) (Gwinn, 1961). The thickness of the unit

varies from 6,000 feet in the west to 4,500 feet in the east (Gwinn, 1961).

Gwinn described the Carter Creek Formation as intercalated tan to gray and greenish-gray sandstone, siltstone, and silty mudstone. A prominent, dark, marine shale in the lower 2,000 feet of the Carter Creek overlies a basal, chert-pebble-rich sandstone which is in contact with the Jens Formation.

The top of the Carter Creek Formation is marked by an angular unconformity and is overlain by the Golden Spike facies of the Elkhorn Mountains Volcanics. The contact is poorly exposed and is primarily indicated by an abrupt thinning of the formation.

Golden Spike Facies of the Elkhorn Mountains Volcanics. The Golden Spike facies does not occur in the study area, but can be seen immediately to the east in the middle of T. 9 N., R. 10 W. The rocks crop out in the walls of the Clark Fork Valley and the hills southwest of Garrison, Montana. Gwinn and Mutch (1961) estimated the thickness of the Golden Spike at 1,500-6,000 feet, although the complete section is not visible.

The Golden Spike facies as described by Gwinn (1961) and Mutch (1961) is an intertongued succession of andesitic-dacitic lavas; tuffs; and volcaniclastic siltstones, sandstones, and massive conglomerates. The volcanic tongues of the unit include light to dark, greenish-gray siltstone and purplish-gray to greenish-gray, coarse-grained sandstone. These beds display poor sorting and bedding and are volcaniclastic. The conglomeratic tongues are characterized

by drab-colored, lenticular beds of pebble- to boulder-sized quartzite, chert, and limestone fragments from Paleozoic and Precambrian rocks. These conglomerates and sandstones which contain abundant plant fragments commonly truncate the lower beds within the member.

Gwinn and Mutch interpreted the Golden Spike facies as an intertongued facies of volcanic detritus (shed from the Elkhorn Mountains Volcanics to the east) and conglomerates and sandstones (shed from an unnamed uplift to the west).

An angular unconformity above the Golden Spike facies marks the contact with Tertiary beds of Miocene (?) age.

Cenozoic

Tertiary Strata

Outcrops of the Tertiary strata are quite variable and scattered throughout the area. The previous workers in the area described isolated outcrops of Tertiary; however, Konizeski, et al (1961) covered the subject most thoroughly, Gwinn (1961) provided a description of the Cabbage Patch Beds.

Konizeski, et al (1961) described approximately 250 feet of semi-consolidated conglomerate, sandstone, siltstone, and clay which are stained with limonite. The rocks were tentatively dated as Oligocene (?). The Early Miocene is approximately 270 feet of volcanic ash and lime-cemented, fluvial siltstone, sandstone, and pebble conglomerate. Overlying these rocks is an unmeasured thickness of colluvial sand and

silt, flood-plain silt and fine sand, and channel deposits of coarse sand and gravel. Volcanic ash occurs throughout the sequence.

Cabbage Patch Beds. The Cabbage Patch beds crop out in the southwest quarter of Sec. 25, T. 10 N., R. 12 W. on the south side of the Clark Fork Valley. These rocks form light-colored, grass-covered slopes with occasional resistant beds standing out with relief (Plate 1). Gwinn (1961) described the rocks and estimated their thickness at 0-500 feet.

The Cabbage Patch beds are composed of three units. The upper sequence is poorly-bedded tuffs and volcanic-rich siltstones which overlie a micaceous, feldspathic conglomerate and sandstone sequence. This unit overlies 0-2400 feet of laminated limestone, fine-grained detrital rocks, and an impure diatomite bed. Massive bedding is a prominent characteristic of the Cabbage Patch.

Quaternary Deposits

The Quaternary deposits are composed of glacial till (from three glacial periods), talus, alluvium, soil, and lake sediments. The latter are not described here. Mutch (1961) described the three glacial tills and interpreted them to represent early, intermediate, and late glacial periods corresponding to the Buffalo (?), Bull Lake (?) and Pinedale glacial advances (Poulter, 1957).

Mutch described the early till as extensively weathered material with no morainal form. The sediment is composed of quartzite and hornfels cobbles and small boulders with a distinctive lack of granitic rocks. The intermediate till is found in extensively

weathered deposits with very poor, morainal form. This till contains igneous rock as well as the quartzite and hornfels mentioned above. The late glacial till displays good glacial and morainal topography and is composed of unweathered granite boulders.

Igneous and Metamorphic Rocks

Cretaceous Intrusives

Granitic bodies of Cretaceous age, 76.7-72 m.y. old (Hyndman, 1979), occur in the Flint Creek Range as a group of stocks which, in most cases, cut preexisting structures in the sedimentary rocks. McGill (1959) described these intrusive rocks as gray, medium-crystalline granodiorite and quartz diorite with small phenocrysts of zoned plagioclase (andesine). The margins of the stocks display faint foliation parallel to the country rock.

Cretaceous Sills

Several Cretaceous sills intrude the Colorado Group in the northeastern part of the study area in T. 9 N., R. 12 W (Plate 1). Gwinn (1961) and Mutch (1961) described these sills as equigranular, porphyritic diorite with plagioclase (labradorite-andesine) and augite phenocrysts and interstitial granophyric quartz, potash-feldspar, and oligoclase. The country rock surrounding the sills is slightly contact-metamorphosed. The sills are locally conformable with the over- and underlying rocks; however, over a distance they change stratigraphic position by up to 700 feet (Gwinn, 1961).

Tertiary Volcanics

Several exposures of Tertiary volcanics are on the northern margin of the area. Mutch (1961) and Gwinn (1961) described these as basaltic, andesitic, dacitic, and rhyolitic extrusive rocks which vary widely from horizontally-lineated lava flows to angular and unsorted breccia.

Metamorphic Rocks

The country rocks around the Flint Creek plutons display various grades of contact metamorphism. These rocks, which occur in an aureole-like pattern around the plutons, exist within one and half miles of the igneous contact (McGill, 1959). McGill defined several zones of metamorphism in the epidote-amphibolite facies. Characteristic rocks are marble, hornfels, and quartzites. These metamorphic rocks are not identified in this study; instead, they are correlated with their unmetamorphosed parents and mapped as such.

LEGEND

Sedimentary Rocks

Cz Cenozoic

Talus, Alluvium,
Glacial till, &
Fluvial & Lake
Sediments

Mz Mesozoic

Colorado Gp.,
Kootenai Fm.,
Morrison Fm.,
& Ellis Gp.

Pz Paleozoic

Shedhorn Fm., Phos-
phoria Fm., Park
City Fm., Quadrant Fm.,
Amsden Fm., Madison
Gp., Jefferson Fm., May-
wood Fm., Red Lion Fm.,
Hosmark Fm., Sil-
ver Hill Fm., &
Flathead Fm.

Pc Precambrian

Missoula Gp., New-
land Fm., & Rav-
alli Fm.

Igneous rocks

Tertiary volcanics
Andesitic composition

Cretaceous sills
Basaltic composition

Cretaceous intrusives
Granodioritic & quartz monzonitic
composition

Symbols

Faults

Thrust or high angle reverse faults - points
on surface side

Normal faults - (up, or down)
on surface side

unestablished inferred

Folds

Arrows indicate direction of plunge

upright anticlines

upright synclines

Contacts

established inferred

Attitudes

strike and dip of sedimentary beds

upright

subvertical

vertical

subhorizontal

horizontal

Scale

0 1 2

Miles

(1:50,000)

Scale

0 1 2

Miles

(1:50,000)

Scale

0 1 2

Miles

(1:50,000)

Scale

0 1 2

Miles

(1:50,000)

Scale

0 1 2

Miles

(1:50,000)

Scale

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(1:50,000)

Composite Geologic Map of the Flint Creek Range, Montana

Compiled by

Jeffrey F. Baken

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