Rocks and Structure of the North-Central Part of the Anaconda Range, Deer Lodge and Granite Counties, Montana

U.S. GEOLOGICAL SURVEY BULLETIN 1993

Rocks and Structure of the North-Central Part of the Anaconda Range, Deer Lodge and Granite Counties, Montana

By DAVID J. LIDKE and CHESTER A. WALLACE

A description of rocks and an analysis
of polyphase deformation in thrust sheets
of the Middle Proterozoic Belt Supergroup
and Paleozoic rocks

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PLATE

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FIGURES

Rocks and Structure of the North-Central Part of the Anaconda Range, Deer Lodge and Granite Counties, Montana

By David J. Lidke *and* Chester A. Wallace

Abstract

The north-central part of the Anaconda Range consists principally of faulted and folded Middle Proterozoic and Paleozoic sedimentary rocks, Late Cretaceous to early Tertiary monzogranite and granodiorite stocks, and thin deposits of Tertiary(?) gravel and Pleistocene till and outwash. Middle Proterozoic rocks in this area include the middle and upper parts of the Belt Supergroup, which is a thick sequence of mainly clastic and some carbonate-rich to carbonate-bearing rocks. Paleozoic rocks in the area form a stratigraphic sequence of Middle Cambrian to Pennsylvanian rocks that contains several disconformities, and consists dominantly of thick-bedded limestone and dolomite and lesser amounts of carbonate-bearing shale and siltstone and some sandstone and quartzite. Late Cretaceous to Tertiary monzogranite and granodiorite stocks occur in the southern part of the area, where they have metamorphosed the Belt Supergroup rocks to hornfels; dikes and sills related to the stocks are concentrated in the southern part of the area but also occur locally in the central and northern parts. Small patches of Tertiary(?) gravel are present in the west. Pleistocene till is present in most drainages and mantles some ridges; outwash occurs in some drainages, and rock glaciers are present on the floors of some cirques. Holocene alluvium, colluvium, and talus form thin, widespread deposits. Small landslide deposits are present in the valley of Carpp Creek.

During Late Cretaceous time, Proterozoic and Paleozoic rocks were tectonically transported to their present locations and stacked along flat, regionally extensive thrust faults; the thrust-bounded sequences of rock are called thrust sheets in this report, and the bounding master thrusts are called sheet thrusts. Thrust sheets are regionally recognizable because each sheet has a stratigraphic identity and an internal stratigraphic continuity. The Georgetown, East Fork, and Cutaway thrusts are principal sheet thrusts in the north-central Anaconda Range. Following emplacement of thrust sheets, continued compression during Late Cretaceous time folded the thrust-sheet stack and produced late-phase imbricate and out-of-syncline thrust faults that offset the folds and folded thrust sheets. The Georgetown and East Fork thrusts are folded above the Cutaway thrust, and younger imbricate thrust faults cut tight and isoclinal folds along the western part of the study area.

After thrust faulting and folding ceased, granitic stocks were emplaced in the southern part of the area. These stocks are part of a Late Cretaceous to early Tertiary plutonic complex that forms most of the crest and southern flank of the Anaconda Range. Emplacement of the stocks, or isostatic adjustments related to emplacement of stocks, tilted the folded thrust sheets to the north on the north flank of the range, which produced the present northerly plunge of folds and northerly dip of the Cutaway thrust. During Tertiary time, high-angle faults offset the sedimentary rocks, the stocks, and the Cretaceous thrust faults and folds.

INTRODUCTION

The study area is located in the north-central part of the Anaconda Range in southwest Montana, about 20 mi (32 km) west of Anaconda and about 18 mi (29 km) south of Philipsburg (fig. 1). Much of this area is in the Anaconda-Pintlar Wilderness. Most of the northern part of the area is accessible from unpaved U.S. Forest Service roads south from Highway 38; the central and southern parts are accessible by trails into the wilderness from those unpaved roads.

Previous Studies

Most of the north-central part of the Anaconda Range had been previously mapped (Emmons and Calkins, 1913; Poulter, 1956; Flood, 1974; and Wiswall, 1976), but new studies mandated by the U.S. Congress under the Wilderness Act of September 3, 1964, provided an opportunity to remap the Middle Proterozoic succession and to integrate the

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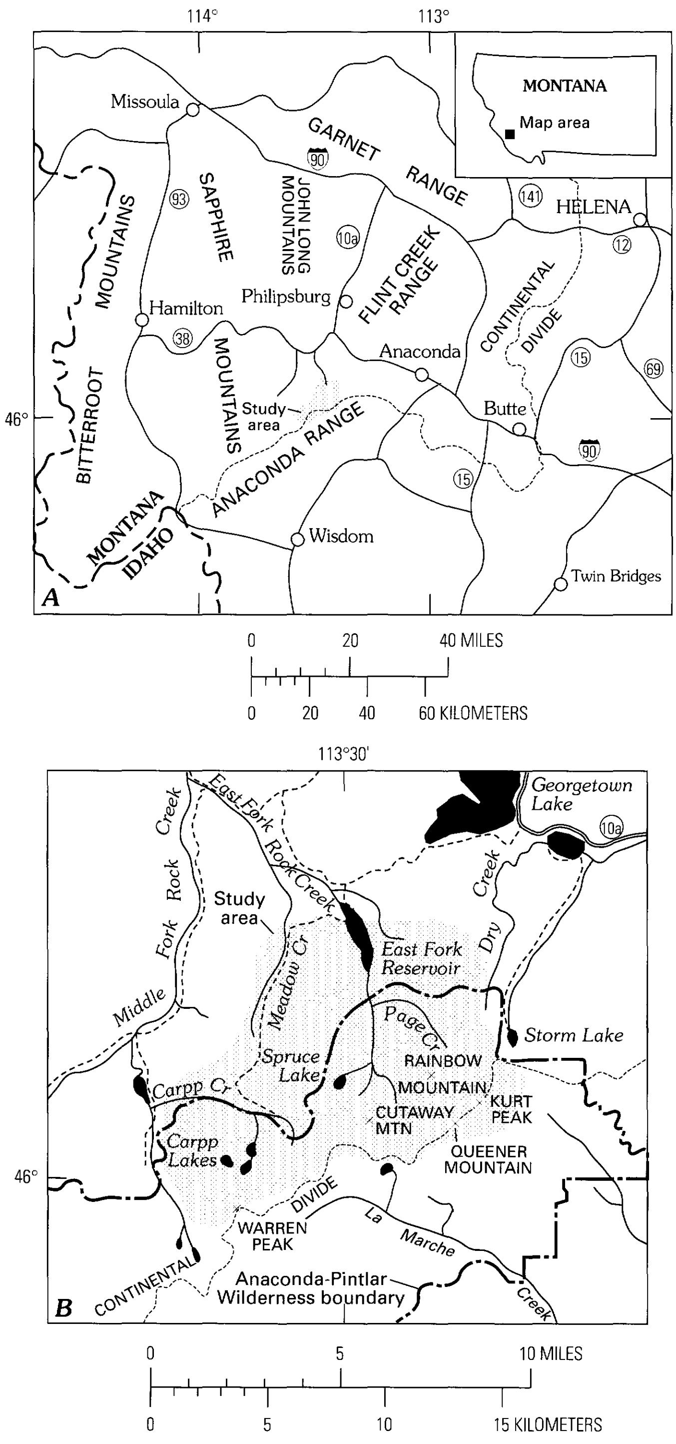


Figure 1. Index maps showing location of study area. *A*, southwestern Montana region; *B*, enlarged view of study area.

stratigraphic and structural data from the Anaconda Range with new regional mapping in the Butte 1°×2° quadrangle (Wallace and others, 1987). Emmons and Calkins (1913) published the first geologic map (scale 1:125,000) that includes most of the northern part of the Anaconda Range; and this classic report first described the structures in the region, subdivided and named formations of Paleozoic age, and provided the first subdivision and correlation of the Precambrian rocks in the area. Poulter (1956) published a geologic map (scale 1:48,000) that included the north-central part of the Anaconda Range; he refined structural and stratigraphic relations of Belt Supergroup rocks, and he recognized the importance of sequential restorations to resolving problems of polyphase thrust faulting and folding in the Anaconda Range. Two theses (Flood, 1974; Wiswall, 1976) that extend into the study area have identified a large, refolded nappe and recognized polyphase folding and thrust faulting in rocks structurally beneath and directly south of the area of this report.

Present Study

The geologic data and interpretations presented here result from geologic mapping used to support a resource evaluation of the Anaconda-Pintlar Wilderness (Elliott and others, 1985). Geologists who participated in the Anaconda-Pintlar Wilderness mapping are C.A. Wallace, D.J. Lidke, J.E. Elliott, N.R. Desmarais, J.M. O'Neill, D.A. Lopez, S.E. Zarske, B.A. Heise, M.J. Blaskowski, J.S. Loen, and D.C. Ferris. The north-central part of the Anaconda Range was mapped by D.J. Lidke to obtain structural data in an area that exposes deep structures in the thrust system of the Sapphire thrust plate.

Some of our rock-unit identifications differ from those of Emmons and Calkins (1913) and from Poulter (1956); our nomenclature changes required complementary changes in previous structural interpretations. These differences from earlier work will be addressed in detail in a later section of this report.

Acknowledgments

We are indebted to the other members of the geologic mapping team of the Anaconda-Pintlar Wilderness listed above; their geologic mapping and studies in adjacent areas of the Anaconda Range contributed much to this report. A.J. Simons, P.D. Bustamante, S.E. Riley, and P.E. Thomas assisted with field work at different times during these studies. The U.S. Forest Service provided advice and logistical support for usage of a helicopter during the Wilderness studies. Insightful technical reviews and editorial comments by I.J. Witkind, E.E. Glick, K.F. Fox, E.R. Cressman, S.A. Minor, and L.M. Carter (U.S. Geological Survey) greatly improved the map and report.

STRATIGRAPHY

Middle Proterozoic (Belt Supergroup) and Paleozoic sedimentary rocks underlie most of the study area (pl. 1). The oldest rocks exposed in the north-central Anaconda Range form part of the Belt Supergroup, and they were deposited in the southern part of the Belt basin (Harrison and others, 1974). Thick deposits of clastic sediment and lesser amounts of carbonate-bearing sediment accumulated during Middle Proterozoic time to form a thick sequence of quartzite, siltite, argillite, and limestone or dolomite. The incompletely represented sequence of Belt rocks exposed in the area is about 12,000 ft (3,660 m) thick. During Paleozoic time, limestone, dolomite, and lesser amounts of sandstone, siltstone, and shale were deposited unconformably on Belt Supergroup rocks. The Paleozoic sequence is about 5,300–5,500 ft (1,615–1,675 m) thick in the north-central Anaconda Range. Permian and Mesozoic sedimentary rocks are exposed about 10 mi (16 km) northeast of the area (Wallace and others, 1987), but these rocks are absent in the study area.

Middle Proterozoic Rocks (Belt Supergroup)

Rocks of the Belt Supergroup exposed in the study area consist of, in ascending order, the Helena Formation (middle Belt carbonate), and parts of the Snowslip, Mount Shields, and Garnet Range Formations (Missoula Group). Plate 1 shows the distribution of Belt Supergroup rocks in the study area, and figure 2 shows the stratigraphic nomenclature applied to these rocks in this study and previous studies of this area and adjacent regions. Emmons and Calkins (1913) determined that "Belt Series" (Belt Supergroup) rocks in the northern Anaconda Range consisted of, from oldest to youngest, siliciclastic rocks of the Ravalli Formation, carbonate-rich rocks of the Newland Formation, and siliciclastic rocks of the Spokane Formation (the Neihart Quartzite and Prichard Formation were mapped by Emmons and Calkins south and southeast of the area). Emmons and Calkins (1913, p. 41) recognized that the carbonate-rich, Precambrian rocks in the Anaconda Range were an eastern equivalent of the Wallace Formation (middle Belt carbonate), but they mistakenly called these carbonate-rich rocks the Newland Formation, which is a carbonate unit in the lower part of the Belt Supergroup. The Helena Formation is now known to be the eastern equivalent of the Wallace Formation and these two formations are informally called the middle Belt carbonate (Harrison, 1972). Poulter (1956) continued to assign these carbonate-rich rocks to the Newland Formation, but he recognized that the overlying clastic rocks, mapped as the Spokane Formation by Emmons and Calkins (1913), were correlative with formations of the Missoula Group that is now known to overlie the middle Belt carbonate formations over most of the Belt basin (Harrison, 1972). Flood (1974) and Wiswall (1976), whose study areas slightly overlap the southwestern part of our study area, correctly identified the carbonate-rich rocks as the middle Belt carbonate, but they assigned them to the Wallace Formation. Elliott and others (1985) and Wallace and others (1987) assigned most of the carbonate-rich rocks in the Anaconda Range to the Helena Formation and they assigned the overlying clastic rocks to the Missoula Group, which they also subdivided into some of the formations that are regionally recognized within the Missoula Group. Contact-metamorphosed argillite, siltite, and quartzite, which underlie the Helena Formation near the Continental Divide, were mapped as Ravalli Group by some previous authors; however, these clastic rocks underlie a low-angle thrust fault (Cutaway thrust) below the Helena Formation, and these rocks are recognized as part of the Mount Shields Formation of the Missoula Group by Elliott and others (1985), by Wallace and others (1987), and in this report.

Helena Formation

The Helena Formation consists mainly of tan-weathering, interbedded and thinly laminated, gray argillaceous limestone and tan limy siltite, and some thin, limy quartzite and silica-cemented quartzite interbeds. The best exposures of the Helena Formation are beneath the Georgetown thrust along and north of the Continental Divide in the southern part of the study area (pl. 1). Where the Helena Formation overlies the Georgetown thrust, the formation is commonly poorly exposed and the upper part is eliminated by faulting or eroded; the lower part of the formation is eliminated by faulting along the Georgetown thrust. The Helena Formation along the Continental Divide is well exposed and is about 6,000 ft (1,830 m) thick, but most of the lower part is eliminated by faulting along the Cutaway thrust. The lowest exposed Helena Formation above the Cutaway thrust consists of about 300 ft (90 m) of thinly bedded, tan oolitic siltite; pale-green, calcareous argillite and siltite; and some beds of white, fine-grained, limy sandstone and white quartzite. These lower rocks are overlain by about 4,000 ft (1,220 m) of pale-pink- and pale-tan-weathering, thick-bedded argillaceous limestone and dolomite, which is overlain by a thinly bedded and laminated zone that consists of alternating beds and laminations of dark-gray argillaceous limestone and tan, limy siltite, and some algal laminations. The thinly laminated zone is about 1,500 ft (455 m) thick and is overlain by about 200 ft (60 m) of green, limy, thinly laminated argillite and siltite that grades upward into red and green argillite and siltite of the overlying Snowslip Formation.

Missoula Group

The Missoula Group in the north-central part of the Anaconda Range consists of the Snowslip, Mount Shields, and Garnet Range Formations, units that are composed mainly of quartzite, siltite, and argillite. None of these formations are complete in the study area, because of thrust faulting (fig. 2).

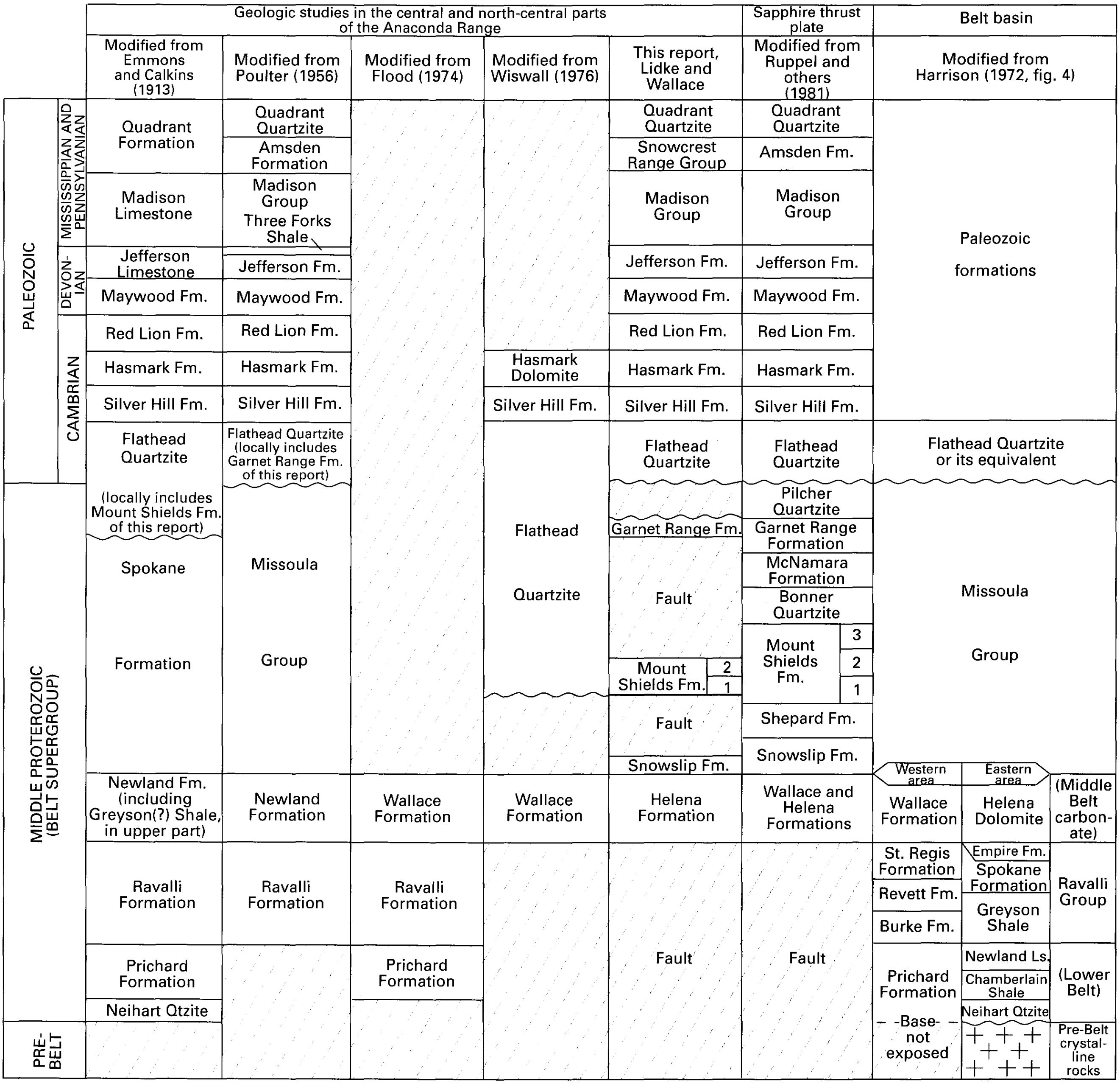


Figure 2. Comparison of Middle Proterozoic (Belt Supergroup) and Paleozoic stratigraphic sequences mapped in the north-central Anaconda Range in this study and in previous studies, with the stratigraphic sequences that characterize the Sapphire thrust plate and the Belt basin. Rock unit names in parentheses, in far right column, are used informally by Harrison (1972). Wavy line, unconformity.

The Shepard Formation, Bonner Quartzite, and McNamara Formation, which are part of the Missoula Group, occur within 5 mi (8 km) north, northwest, or southwest of the study area, but these formations are absent in the study area itself, having been eliminated by thrust faults (fig. 2). The Pilcher Quartzite is also absent in the study area: it occurs about 38 mi (61 km) farther north but does not extend as far south as the study area in the Belt basin (Wallace and others, 1987; Wallace and others, 1989). Because of thrust faulting in formations of the Missoula Group, complete thicknesses of these formations were not determined. The maximum exposed thicknesses of the Snowslip, Mount Shields, and Garnet Range Formations are about 650 ft (200 m), 5,500 ft (1,675 m), and 400 ft (120 m), respectively. Where the Missoula Group is exposed between the Helena Formation (Middle Proterozoic) and the Flathead Quartzite (Middle Cambrian), it ranges from about 2,500 ft (760 m) to 900 ft (275 m) thick. About 15 mi (24 km) to the west in the southern Sapphire Mountains, however, the Missoula Group has a minimum thickness of about 18,000 ft (5,500 m) (Wallace and others, 1989). Emmons and Calkins (1913) recognized this abrupt local change in thickness of rocks now assigned to the Missoula Group, as did Poulter (1956); their structure sections suggest that they attributed the change in thickness to overthrusting upper plate rocks of the Georgetown thrust, which contained a thick sequence of rocks now included in the Missoula Group, onto rocks in the lower plate, which contained a thin sequence of Missoula Group that had been erosionally thinned beneath the unconformity at the base of the Flathead Quartzite.

Snowslip Formation

Overlying the Helena Formation is the Snowslip Formation, which is best exposed on the north side of Rainbow Mountain in the southeastern part of the study area; only about the lower 650 ft (200 m) of the formation is present. This lower part consists mainly of alternating zones of green and red thinly laminated argillite and siltite and some thin beds and lenticular beds of medium- and coarse-grained, well-rounded and well-sorted white quartzite. Argillaceous laminated siltite and argillaceous, fine-grained quartzite beds are about 0.4–6 in. (1–15 cm) thick, and these beds are irregularly distributed among zones of thinly laminated argillite and siltite that range in thickness from 0.4 in. to 3 ft (1 cm to 1 m). The nonrhythmic bedding and the lenticular beds of white, coarse-grained, well-sorted quartzite are characteristic features of the Snowslip Formation, but the well-sorted quartzite lenses are most common in the lower part. Water-expulsion structures, mud cracks, mud chips, and ripple cross-lamination are common; oolites, glauconite, and armored mudballs occur near the base of the formation. The upper part of the Snowslip Formation, which contains more and thicker argillaceous siltstone and fine-grained quartzite beds than the lower part, is absent in the study area.

Mount Shields Formation

Where the Mount Shields Formation is present in normal stratigraphic position, outside the study area, it overlies the Shepard Formation, which in turn overlies the Snowslip Formation; and, in normal stratigraphic position the Mount Shields Formation underlies the Bonner Quartzite (fig. 2). In the central part of the study area, the Mount Shields Formation overlies the Snowslip Formation; it overlies the Helena Formation in the western part of the area, and it underlies the Helena Formation in the southeastern part of the area. All those contacts are thrust faults. Although three members of the Mount Shields Formation are mapped in the southern part of the Belt basin (Harrison and others, 1986; Wallace and others, 1987), only the upper part of member one, the basal unit, and the lower part of member two are exposed in the study area.

Member one is characterized by zones of feldspathic and nonfeldspathic quartzite that alternate with zones dominated by interbedded red argillite and buff siltite. The quartzite zones are composed of pink, white, and buff, moderately sorted to well-sorted, planar-laminated and crossbedded strata; contacts at the base of quartzite bedding units generally are planar. The argillite and siltite zones are rhythmically interlayered and contain some beds and laminations of fine-grained quartzite. Individual zones of quartzite and zones of interbedded argillite and siltite range in thickness from about 25 to 200 ft (8 to 60 m). The quartzite zones in member one increase in thickness upward, whereas the rhythmically interbedded argillite and siltite zones decrease in thickness upward. In the western part of the area, the quartzite zones contain coarse-grained feldspathic quartzite, some channel conglomerate, and, about 100–300 ft (30–90 m) below the contact with member two, matrix-supported conglomerate. The lower part of member one, which is composed mostly of argillite and siltite zones, is absent in the study area. The contact with the overlying member two is gradational.

Member two consists mainly of thickly bedded pink and white, fine- to medium-grained feldspathic quartzite, arkose, and nonfeldspathic quartzite beds that have planar basal contacts. Argillite and siltite are rare in member two, except as partings between 0.5–3 ft (0.15–1 m) thick quartzite beds. A zone of coarse-grained quartzite and some thickly bedded channel conglomerate occurs about 200–300 ft (60–90 m) above the base of member two. The upper part of member two is absent in the study area; about 15 mi (24 km) farther west, it is present and contains characteristic thin beds of pebble conglomerate (Wallace and others, 1989).

Members one and two of the Mount Shields Formation exhibit three slightly different lithofacies in the north-central part of the Anaconda Range. Although the general lithologic description above is applicable to all three lithofacies, persistent lithologic characteristics distinguish each lithofacies, and they are informally called the western, central, and southern lithofacies in this report.

The western lithofacies is recognized only along the westernmost part of the study area, west of Meadow Creek and west of the Carpp Lakes, and is confined to the thrust sheet above the Georgetown thrust. In the upper part of member one and the lower part of member two, the western lithofacies is identified by its coarser grains and a more arkosic and conglomeratic aspect as compared to the central and southern lithofacies. The western lithofacies of member one contains channel-fill deposits of feldspathic, matrix-supported pebble and cobble conglomerate that occur about 100–200 ft (30–60 m) below the top of member one. The western lithofacies of member two includes at least two zones of feldspathic, matrix-supported conglomerate, containing cobbles as large as 3 in. (7.6 cm) in diameter, that occur in member two about 300 ft (90 m) above the contact with member one.

The central lithofacies is recognized in the cores and along the flanks of Rock Creek anticline and Spruce Creek and Porter Ridge synclines. Central lithofacies rocks of members one and two of the Mount Shields Formation underlie the East Fork thrust and overlie the East Fork branch thrust, respectively. Coarse-grained quartzite and conglomerate characterize this lithofacies in the upper part of member one and in the lower part of member two at approximately the same stratigraphic positions as in the western lithofacies; however, the coarse zone in the central lithofacies of member one consists mainly of thin beds and lenses of pebble-size conglomerate and coarse-grained quartzite interbedded with argillite, siltite, and fine-grained quartzite. The zone containing these coarse-grained and mildly conglomeratic rocks is as thick as about 300 ft (90 m) and underlies about 300 ft (90 m) of the uppermost beds of member one. The central lithofacies of member two also contains coarse zones that include some interbeds of medium- to coarse-grained quartzite and conglomerate that is characterized by matrix-supported quartzite clasts as large as 2 in. (5 cm) in diameter. These coarse zones are as thick as about 50 ft (15 m), they occur in the lower few hundred feet of member two, and the matrix of the conglomerates and the quartzite interbeds are finer grained and markedly less arkosic than are the coarse zones in the western lithofacies of member two of the Mount Shields Formation.

The southern lithofacies is recognized only along and south of the Continental Divide beneath the Cutaway thrust in the southern part of the area; these clastic rocks were previously mapped as the "Ravalli Formation" by Emmons and Calkins (1913), but they are correlated with the Mount Shields Formation by Elliott and others (1985) and Wallace and others (1987). In general, the southern lithofacies is finer grained and better sorted than the western and central lithofacies. In member one, quartzite zones in the southern lithofacies range from 25 to 50 ft (7.5 to 15 m) thick, are consistently fine to very fine grained, and contain only a few percent of feldspar; bedding surfaces are flat, and planar laminations and ripple cross-laminations are common. Zones of rhythmically interbedded argillite and siltite contain more silt-sized detritus and less sand- and clay-sized detritus than occur in the central and western lithofacies. Quartzite beds that compose member two are similar to quartzite beds of the southern lithofacies in member one; and near the base of member two, a conspicuous zone of convolute bedding, which was not identified in the western and central lithofacies, locally identifies the southern lithofacies. No beds or lenses of conglomerate or coarse-grained quartzite are known from either member in the southern lithofacies. Contact metamorphism has altered the argillaceous rocks to shades of gray and black and the quartzite to buff, gray, and white. Some calc-silicate minerals occur mainly in the argillaceous beds and as coatings along fractures. The calc-silicate minerals may represent primary or secondary carbonate in the protolith or carbonate that was mobilized and added to the Mount Shields during deformation and metamorphism.

Bedding structures, bedding characteristics, and textures common to the Mount Shields Formation elsewhere in the Belt basin identify rocks of the three lithofacies in the study area as Mount Shields, even where they have been metamorphosed. Mount Shields quartzite beds typically are normally graded and have planar basal contacts. From bottom to top, a zone of planar lamination or large-scale ripple cross-lamination (current ripples), a zone of small-scale ripple cross-lamination (wave ripples), a zone of planar lamination, and nonlaminated or microlaminated siltite characterize single quartzite beds that are as much as 3 ft (1 m) thick. Commonly, isolated well-rounded medium or coarse quartz grains are widely dispersed in a feldspathic or quartzose matrix of subangular fine-grained and silty quartzite. Bedding in the Mount Shields Formation is characteristically rhythmic, and consists of alternations of siltite or quartzite beds with argillite beds. Most contacts between argillite layers and overlying siltite or quartzite beds are planar. Although the rhythmic character of the bedding is most conspicuous in argillite zones of member one, rhythmic bedding is also common in quartzite zones of member one and in the thick quartzite beds of member two where thin layers of argillite or argillaceous siltite overlie and alternate with thick beds of quartzite. In both members of the Mount Shields Formation, feldspathic beds are interlayered with quartzose beds that contain only a trace of feldspar, and in coarser zones, the lower part of bedding units is commonly coarse grained and feldspar rich and grades upward into medium- or fine-grained, feldspar-poor layers.

Garnet Range Formation

The Garnet Range Formation is thrust over the Mount Shields Formation (member two, in areas marked by the central lithofacies) in the western part of the study area, and over the Snowslip Formation in the eastern part. The Flathead Quartzite unconformably overlies the Garnet Range Formation in the study area. Only about 300–400 ft (90–120 m) of the upper, quartzitic part of the Garnet Range Formation is present, and these rocks are similar in composition, bedding character, and mottling to the upper part of the Garnet Range Formation, as exposed near Porters Corner about 10 mi (16 km) north of the study area (Winston and Wallace, 1983, p. 79–80). In the study area, the upper part of the Garnet Range Formation consists mainly of blocky-weathering, grayish-red, reddish-brown, light-green, grayish-green, and white quartzite. Downward in the sequence, dull-red and brownish-red argillaceous and silty quartzite beds are separated by interbeds of grayish-red, silty and sandy argillite. Alternation of red or green argillaceous quartzite beds, which vary from 2 in. to 1 ft (5 to 30 cm) thick, and grayish-red, silty and sandy argillite beds, which vary from 1 to

3 in. (2.5 to 7.5 cm) thick, is the typical bedding characteristic of most of the Garnet Range Formation below the upper quartzite zone. Irregular white and pale-green mottles are common in the quartzite beds. Planar lamination, ripple cross-lamination, parting lineation, and small-scale planar and trough crossbeds are the most common sedimentary structures in the Garnet Range Formation. The uppermost quartzites of the Garnet Range Formation are similar in composition, texture, and bedding character to the overlying Flathead Quartzite, and the exact position of the disconformity separating the two units is difficult to identify in most exposures (Emmons and Calkins, 1913, p. 51; Winston and Wallace, 1983).

Paleozoic Rocks

The Paleozoic rocks in the study area are represented by the Flathead Quartzite, the Silver Hill, Hasmark, Red Lion, Maywood, and Jefferson Formations, the Madison and Snowcrest Range Groups, and the Quadrant Quartzite (pl. 1). Paleozoic formations are those defined and mapped by Emmons and Calkins (1913), except that the Quadrant Formation of Emmons and Calkins (1913), mapped as the Quadrant and Amsden Formations by Poulter (1956), is divided into the Snowcrest Range Group at the base (B.R. Wardlaw, U.S. Geological Survey, written commun., 1987) and the Quadrant Quartzite at the top. The Paleozoic rocks are about 5,300–5,500 ft (1,615–1,675 m) thick and comprise mainly dolomite and limestone, and lesser amounts of shale, siltstone, and sandstone.

Cambrian System

Cambrian rock units in the study area consist of the Flathead Quartzite and Silver Hill Formation (Middle Cambrian), and the Hasmark and Red Lion Formations (Upper Cambrian). The Flathead Quartzite is the lower part of the Flathead Formation as described by Peale (1893) from exposures near Three Forks, Mont., whereas the Silver Hill, Hasmark, and Red Lion Formations were described and named by Emmons and Calkins (1913); and the type localities of these formations are all within 15 mi (24 km) northeast of the study area.

Flathead Quartzite

The Flathead Quartzite has a maximum thickness of about 100 ft (30 m), but is locally faulted out. The Flathead, which disconformably overlies the Garnet Range Formation (Middle Proterozoic), consists mainly of white to tan, fine- to medium-grained, rounded to well-rounded, well-sorted quartzite and some thin green shale laminations.

Silver Hill Formation

The Silver Hill Formation conformably overlies the Flathead, although locally it is faulted over the Garnet Range Formation. The Silver Hill consists of about 50 ft (15 m) of olive-green and greenish-black, flaser-bedded shale and interbedded argillaceous siltstone and sandstone that grade upward into about 100–150 ft (30–45 m) of light-gray limestone intercalated with thin beds and laminations of yellow to red, unevenly bedded, dolomitic and calcareous mudstone. The mudstone beds and laminations stand out in relief on weathered surfaces. The distinctive green shale upper member, described by Emmons and Calkins (1913) at the type section about 5 mi (8 km) to the northeast, is not present in the study area.

Hasmark Formation

The Hasmark Formation conformably overlies the Silver Hill Formation and is about 1,000 ft (305 m) thick in the study area. The lower part of the Hasmark consists of about 650–700 ft (200–215 m) of massive-weathering bluish-gray dolomite that locally contains an oolite and pisolite zone near the base. Weathered surfaces of the lower dolomite are mottled light and dark gray. The upper part of the Hasmark Formation consists of about 330 ft (100 m) of blocky-weathering and cream-weathering, light-gray, faintly laminated dolomite. Some zones of dark-brown shale, about 1–5 ft (0.3–1.5 m) thick, occur mainly in the middle and lower parts of the formation. The formation locally contains oolites and pisolites near the top.

Red Lion Formation

The Red Lion Formation conformably overlies the Hasmark and is about 350 ft (105 m) thick in the study area. This formation consists mainly of gray limestone that contains yellow to red, dolomitic and siliceous, unevenly bedded, internally laminated, wavy, thin interbeds and laminations; and this banded limestone is similar in appearance to some banded limestone in the middle part of the Silver Hill Formation. Near the base of the Red Lion Formation, the siliceous interbeds and laminations become more closely spaced as the limestone grades downward into a basal clastic zone that consists of about 15 ft (5 m) of red sandy mudstone and shale and thin, irregular limestone laminations and nodules. A tan-weathering, discontinuous sedimentary breccia, as much as 10 ft (3 m) thick, occurs near the top of the Red Lion Formation just north of Carpp Creek.

Devonian System

The Maywood Formation (Upper Devonian) was described and named by Emmons and Calkins (1913) from exposures on Maywood Ridge, 25 mi (40 km) north-northeast of the study area. The Jefferson Formation (Upper Devonian) was described and named by Peale (1893) from exposures near Three Forks, Mont., about 80 mi (130 km) to the east-southeast.

Maywood Formation

The Maywood Formation is about 350 ft (105 m) thick and disconformably overlies the Red Lion; the contact appears transitional across a few tens of feet of section, but Silurian and Ordovician strata are absent. The Maywood Formation consists mainly of gray, dull-green and dull-red, thin-bedded, red-, yellow-, and pink-weathering shale, calcareous shale, and siltstone; some thin interbeds of gray limestone; and, in the upper part, some beds of massive-weathering black dolomite. Locally, the Maywood contains a 15- to 30-ft (5- to 9-m)-thick feldspathic sandstone in the upper part and pink, dolomitic sandstone and sandy, oolitic dolomite in the basal part.

Jefferson Formation

The Jefferson Formation is about 800–850 ft (245–260 m) thick and conformably overlies the Maywood Formation. The Jefferson consists mainly of dark-gray to black, fine-grained dolomite interbedded with some light-gray limestone that gives the formation a striped appearance. The upper part of the formation contains more limestone than the lower part, and locally the upper part contains a pale-bluish-gray sedimentary breccia as much as 5 ft (1.5 m) thick. The dominantly clastic rocks of the Maywood Formation grade upward into dolomite and limestone of the Jefferson Formation across an interval of about 20–50 ft (6–15 m). For mapping purposes the contact was placed where the dark-gray to black dolomite and light-gray limestone become dominant over the pale-red and pale-yellow clastic rocks of the Maywood Formation.

Mississippian and Pennsylvanian Systems

Madison Group

The Lower and Upper Mississippian rocks of the Madison Group, exposed in the north-central and western parts of the study area, probably correlate with the Lodgepole and Mission Canyon Limestones elsewhere in Montana. The lower part of the Madison sequence consists mainly of about 400–450 ft (120–135 m) of dark-gray, chert-bearing, shaly, micritic, fossiliferous limestone that may be correlative with the Lodgepole Limestone. Based on biostratigraphic data, Emmons and Calkins (1913) assigned these shaly limestones to the lower part of their Madison Limestone. Poulter (1956) assigned these beds to the Upper Devonian Three Forks Formation, based on the occurrence of a pelecypod of the species *Grammysia*. Faunal data presented by Emmons and Calkins (1913, p. 69) and by Poulter (1956) conflict and do not establish either a Devonian or a Mississippian age. However, these rocks are much more similar to chert-bearing limestone and shaly, micritic limestone of the Lodgepole Limestone than they are to brown shale and interbedded limestone of the Three Forks Formation. The upper part of the Madison Group is about 1,500 ft (460 m) thick, and is mainly massive weathering, bluish-gray crinoidal limestone that is probably correlative with the Mission Canyon Limestone as suggested by Poulter (1956).

Snowcrest Range Group

Upper Mississippian(?) and Lower Pennsylvanian rocks of the Snowcrest Range Group (Wardlaw and Pecora, 1985) are exposed only in the northeastern part of the study area. These rocks were previously assigned to the lower part of the Quadrant Formation (Emmons and Calkins, 1913). Poulter (1956) and Lidke (1985) assigned these rocks to the Amsden Formation, and they were reassigned to the Snowcrest Range Group by Wallace and others (1987), based on recent work by B.R. Wardlaw (U.S. Geological Survey, written commun., 1987). From bottom to top the Snowcrest Range Group consists of the Kibbey Sandstone, the Lombard Limestone, and the Conover Ranch Formation (Wardlaw and Pecora, 1985). In the study area, the formations are represented but are too thin and too internally faulted and folded to show separately on plate 1. In the Snowcrest Range this group is mainly of Upper Mississippian age, but biostratigraphic data (Emmons and Calkins, 1913; Poulter, 1956) suggest that in the Anaconda Range it is mainly Early Pennsylvanian in age, although the lower part could be Late Mississippian. This age assignment agrees with the interpretation of Wardlaw and Pecora (1985, p. B4) that the Snowcrest Range Group becomes younger toward the northeast and that near the study area the group may contain the Mississippian-Pennsylvanian boundary. The entire Snowcrest Range Group is only about 330 ft (100 m) thick in the study area and consists mainly of interbedded red to maroon calcareous shale and siltstone, olive-green calcareous shale, and discontinuous, white to gray limestone lenses and nodules. The lower 100 ft (30 m) contains a few lenticular beds, as much as 2 ft (0.6 m) thick, of fine-grained, dolomite-cemented sandstone.

Quadrant Quartzite

The Quadrant Quartzite is exposed only in the northeastern part of the study area and consists of about 300 ft (90 m) of white to tan, fine-grained, massive-weathering quartzite, which is lithologically similar to the Flathead Quartzite. The Quadrant was first described by Peale (1893, p. 39–43) and later more clearly defined by Scott (1935). The Quadrant is generally considered to be of Pennsylvanian age, and at several localities in nearby areas it is disconformably overlain by Lower Permian sedimentary rocks of the Phosphoria Formation and related units; the top of the Quadrant is not exposed in the study area.

Tertiary(?) and Quaternary Deposits

Small patches of Tertiary(?) gravel are preserved in the westernmost part of the study area, and Quaternary deposits and associated erosional features occur over most of the area. Thin deposits of poorly stratified, matrix-supported Tertiary(?) gravel, present primarily in the Meadow Creek drainage and along Porter Ridge, consist mainly of rounded to subrounded pebbles and cobbles of quartzite of the Belt Supergroup and rare limestone clasts in a sandy and silty matrix. Pleistocene till and outwash occur in all valleys, and till also mantles some low ridges. The till consists of poorly sorted, poorly stratified boulder and cobble gravel in a poorly sorted matrix of sand, silt, and clay. Till and rock glaciers in the area may represent as many as four stages of Pleistocene to Holocene(?) glaciation (Poulter, 1956). Tills of two, and perhaps three different ages were identified during this study, although they were not mapped separately: (1) a younger till is confined to present valleys, commonly forms hummocky topography, and characteristically lacks plutonic clasts; (2) an older till generally mantles ridges, commonly is covered by soil, lacks a pronounced hummocky topography, and characteristically contains plutonic clasts. (3) A third and oldest till may be represented by a small erosional remnant of gravel present at an elevation of about 9,200 ft (2,800 m) in the saddle about 1 mi (1.6 km) southwest of Spruce Lake.

Angular boulders, as large as about 6 ft (2 m) in diameter, interlock in mound-shaped masses to form rock glaciers and protalus ramparts in the upper part of cirque basins; these deposits are the youngest glacial materials in the area and they may have formed partly or entirely during Holocene time. Patchy, thin deposits of Holocene alluvium, colluvium, talus, and landslide debris occur throughout the area, but with the exception of a few landslide deposits, these younger surficial deposits were not mapped or were included in larger areas of till and outwash.

INTRUSIVE ROCKS AND METAMORPHISM

Three stocks and numerous dikes and sills cut the sedimentary rocks and cut thrust faults and folds in the study area; the stocks occur in the southern part of the area near the Continental Divide. Contacts of the stocks with host rocks are generally sharp, and metamorphic grades of contact aureoles within the host rocks are hornblende hornfels to albite-epidote hornfels. Two of the stocks, which are in contact with one another in the southern and southwestern parts of the area, consist of granodiorite and monzogranite. Part of another granodiorite stock is exposed near the southeast border of the study area. The stocks are part of a plutonic suite of Late Cretaceous to early Tertiary age that occupies most of the central and southern parts of the Anaconda Range (Wallace and others, in press). Most stocks of the complex postdate thrust-and-fold-related deformation but predate movement along Tertiary high-angle faults. Isotopic age determinations were not made on the stocks in the area, but the stocks are similar in composition and texture to nearby stocks and batholiths in the Anaconda Range and southern Sapphire Mountains that have potassium-argon isotopic ages between about 78 and 43 Ma (Desmarais, 1983; Wallace and others, 1989).

The dikes and sills of the study area are more variable in composition and texture than the stocks; they range in composition from andesite to rhyolite, and they generally are porphyritic, but the groundmass shows a range of textures from phaneritic to aphanitic. Most of the dikes and sills have compositions similar to the monzogranite and granodiorite stocks in the area and to a quartz diorite stock exposed just beyond the northeast border of the study area (Elliott and others, 1985), which suggests that they are genetically related to the stocks.

The stocks metamorphosed Proterozoic and Paleozoic sedimentary rocks in the southern part of the study area. Field observations, combined with limited petrographic data, suggest that the metamorphic aureoles adjacent to the stocks range from hornblende-hornfels facies near the intrusive contacts to albite-epidote hornfels facies a few hundred to several hundred feet (100–300 m) from the contacts, and commonly to essentially nonmetamorphosed rock a few thousand feet (1 km) or more from the contacts. Locally, the Middle Proterozoic rocks have been altered to gneiss and schist near contacts with the stocks. Dikes and sills produced thin metamorphic aureoles; these generally do not exceed upper albite-epidote facies.

REGIONAL TECTONIC SETTING

The area of this study is only a small part of the Sapphire thrust plate, a regional tectonic element in south-central and southwestern Montana that occupies nearly the entire area between the Idaho and Boulder batholiths and south of the north boundary of the Lewis and Clark line; the north-central part of the Anaconda Range occupies a central position near the southernmost known extent of the plate (fig. 3). The full

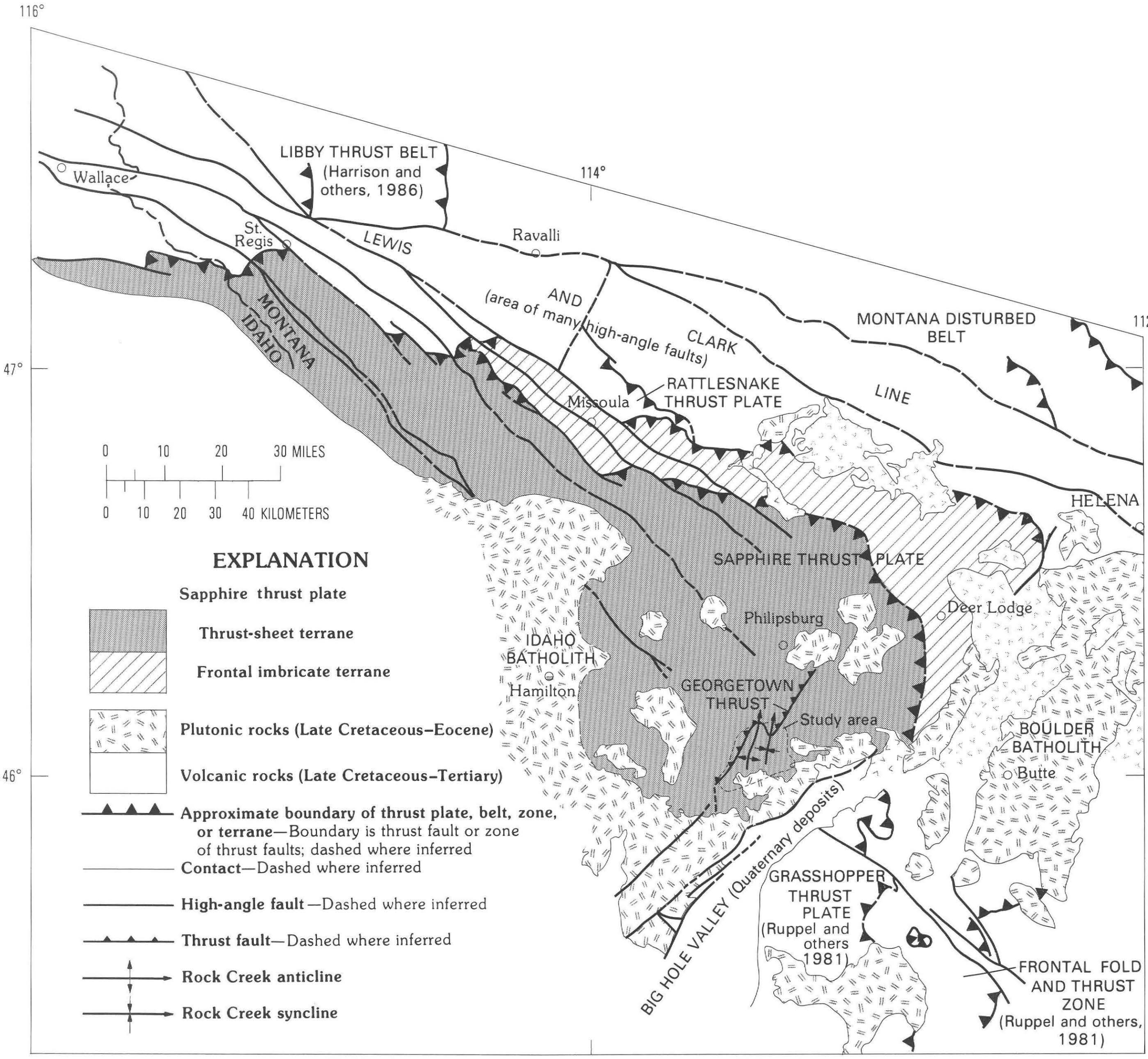


Figure 3. Tectonic sketch map of the Sapphire thrust plate, southwestern Montana.

extent of the Sapphire thrust plate, however, is not known because: (1) north of the Idaho batholith, thrust faults of the leading edge of the plate are truncated by high-angle faults of the Lewis and Clark line; (2) the east edge of the plate was intruded by the Boulder batholith; (3) the southwestern part of the plate was intruded by the Idaho batholith; and (4) relations to thrust plates south of the Sapphire plate, such as the Grasshopper plate (fig. 3), are incompletely known (Ruppel and others, 1981). Thrust faults and folds associated with the Sapphire thrust plate formed partly, or entirely, during Late Cretaceous time, because Late Cretaceous sedimentary rocks were involved in the thrust faulting and folding (Emmons and Calkins, 1913), and Late Cretaceous stocks as old as 82 Ma cut thrust faults and folds of the plate (Wallace and others, 1989). Most of the plutons in the Sapphire thrust plate postdate compressional deformation (Lidke and others, 1987; Wallace and others, 1989) and were emplaced in Late Cretaceous and early Tertiary time. However, some of the older, and commonly most mafic, plutons probably were emplaced and deformed during late stages of thrust faulting and folding (Desmarais, 1983). Tertiary high-angle faults locally offset Proterozoic and Paleozoic sedimentary rocks, the thrust faults and folds, and intrusive rocks as young as middle Eocene described by Desmarais (1983).

Two terranes in the Sapphire thrust plate, which differ in structural style, have been described by Lidke and Wallace (1988) and by Wallace and others (1989): (1) a thrust-sheet terrane; (2) a frontal imbricate terrane. The thrust-sheet terrane, located in the interior part of the Sapphire thrust plate, contains folded stacks of thrust sheets. The frontal imbricate terrane, located along the leading edge of the Sapphire thrust plate, is characterized by a wide zone of anastomosing, imbricate thrust faults and tight folds and probably represents the fragmented frontal zone of thrust sheets. In the thrust-sheet terrane, individual sheets are composed of distinct lithologic sequences, and the thrust sheets are bounded by major thrust faults that form a system of master faults that generally parallel the bedding except locally where they ramp and cut across the bedding. The bounding thrust faults, called sheet thrusts in this report, are commonly folded harmonically with the thrust sheets into broad to tight folds that are cut by late-phase thrust faults. The thickness of individual thrust sheets ranges from about 5,000 to 15,500 ft (about 1,500 to 4,700 m). The stacked thrust sheets of the thrust-sheet terrane terminate in anastomosing, imbricate thrust faults and folds of the frontal imbricate terrane. Thrust faults of the frontal imbricate terrane dip steeply to gently, cut bedding at moderate to steep angles, and commonly cut the limbs and axial regions of tight folds associated with this terrane.

Although the Georgetown thrust was considered a fundamental tectonic boundary in many previous tectonic interpretations of this region (Emmons and Calkins, 1913; Poulter, 1959; Hyndman, 1980; Ruppel and others, 1981), the thrust sheets and associated structures cross the Georgetown thrust rather than terminate against it. In general, the previous tectonic interpretations considered the Georgetown thrust and the Philipsburg thrust about 10 mi (16 km) to the north to approximate a major lithologic and tectonic boundary that separates an allochthonous sequence of Middle Proterozoic sedimentary rocks west of the boundary from a parautochthonous sequence of Middle Proterozoic to Cretaceous sedimentary rocks east of the boundary. However, more recent mapping (Wallace and others, 1987; Wallace and others, 1989) has shown that: (1) the mountainous region west of the Georgetown and Philipsburg thrusts is composed of at least three regionally extensive thrust sheets of Middle Proterozoic rocks; (2) the mountainous region east of the Georgetown and Philipsburg thrusts is composed of at least four regionally extensive thrust sheets, three of which consist of Middle Proterozoic rocks; and (3) thrust sheets in both regions terminate in the frontal imbricate zone of the Sapphire thrust plate. These relations of thrust sheets indicate that the Georgetown thrust is only one of several master faults that characterize the thrust-sheet style of deformation in the interior of the Sapphire thrust plate, and suggest that it probably is inappropriate to consider rocks in the footwall and east of the Georgetown thrust as parautochthonous.

STRUCTURE

The area of this report is in the south-central part of the thrust-sheet terrane (fig. 3) where some of the deepest thrust sheets of the Sapphire plate are exposed. The structure consists of Late Cretaceous thrust faults and related folds and crosscutting, Tertiary high-angle faults.

Three types of thrust faults are distinguished in the study area: (1) sheet thrusts, (2) imbricate thrusts, and (3) out-of-syncline thrusts. These terms for thrust faults are discussed by Dahlstrom (1977, p. 411–414). The Georgetown, East Fork, and Cutaway thrusts are sheet thrusts (pl. 1) that are laterally extensive, generally parallel bedding, and commonly show some shearing and cleavage in fault zones. The Georgetown and East Fork thrusts are folded, whereas the structurally lower Cutaway thrust is not folded in the study area. The imbricate thrust faults in the study area (pl. 1) occur mainly as an anastomosing zone of faults associated with the tight to isoclinal, overturned folds in the western part of the area. Coherent and noncoherent breccia and gouge commonly occur along the imbricate thrust faults. Out-of-syncline thrust faults occur along the east limbs of most of the synclines (pl. 1), as single faults and zones of imbricate faults that locally are anastomosing. They are characterized by fault zones similar to the zones of the imbricate thrust faults in the western part of the area, and they have relatively small separations.

The large-scale folds in the area consist of: (1) open, slightly asymmetric folds, such as the Rock Creek anticline, Rock Creek syncline, and Dry Creek anticline; and (2) tight to isoclinal, overturned folds, which occur on the west side of the study area in the zone of imbricate thrust faults. Field observations indicate that the folds are primarily parallel folds.

The youngest faults in the area are high-angle faults of Tertiary age. These faults cut the thrust faults, folds, and Cretaceous to early Tertiary stocks, but they do not cut Quaternary glacial deposits. The high-angle faults trend mainly north-northeast and west-northwest and incline about 75°–90°. Their fault zones commonly contain coherent and noncoherent breccia and gouge, similar to the breccia and gouge along the imbricate thrust faults.

Sheet Thrusts

The Georgetown, East Fork, and Cutaway thrusts separate thick, laterally extensive sheets of sedimentary rocks, each of which has internal stratigraphic continuity. The sheet thrusts commonly follow bedding planes or truncate them at a low angle. Locally, along the principal sheet thrusts, subsidiary sheet thrusts merge with them to form lensoid thrust slices consisting of either footwall or hanging-wall rocks. The internal stratigraphic continuity of rock units composing individual thrust sheets and the apparent lateral continuity of these structural and stratigraphic features suggest that the thrust sheets are nappe-like and that their bounding faults are consequently sheet-like. The trace of the Georgetown thrust was first mapped by Emmons and Calkins (1913), but the East Fork and Cutaway thrusts were not previously recognized, in part because the stratigraphic succession of Middle Proterozoic rocks was poorly known and in part because some fault-related implications of the large-scale folds were not recognized. The East Fork thrust is structurally beneath the Georgetown thrust, and the Cutaway thrust is structurally below the East Fork thrust. The Georgetown and East Fork thrusts are folded harmonically with the strata above and below. The Cutaway thrust is not folded in the study area, but elsewhere in the Anaconda Range, probable north, northeastern, and southern extensions of the Cutaway thrust are folded (Wallace and others, in press). These three thrust faults do not crosscut or join with each other in the Anaconda Range. Instead they bound thrust sheets that consist of separate parts of the Middle Proterozoic and Paleozoic stratigraphic sequence. The Georgetown and Cutaway thrusts place older rocks on younger rocks, whereas the East Fork thrust places younger rocks on older rocks (pl. 1).

Georgetown Thrust

The southern part of the Georgetown thrust, present in the study area, shows large stratigraphic separation and is extensively folded. Regionally, the Georgetown thrust shows large stratigraphic separation, trends north-to-northeast, and can be traced for a minimum of about 40 mi (65 km) (fig. 3) (Poulter, 1956). Emmons and Calkins (1913) first mapped the Georgetown thrust and showed that the fault trace extended nearly 18 mi (29 km) north of our study area to where it is cut by the Philipsburg batholith. Wiswall (1976, p. 42) tentatively extended the Georgetown thrust about 4 mi (6.5 km) south of our study area, and he showed younger-on-older stratigraphic relations along much of the thrust trace. Recent mapping shown by Wallace and others (in press) indicates that the Georgetown thrust is properly located about 1 mi (1.6 km) west of the area studied by Wiswall; the fault identified by Wiswall as the Georgetown thrust is actually the East Fork branch thrust. The mapping of Wallace and others (in press) extends the Georgetown thrust about 3.5 mi (5.6 km) southwest of the study area where the fault shows about the same stratigraphic separation as in the study area. In the western part of the study area, the Georgetown thrust trends north-northeast. Just north of the north border of the area, the Georgetown thrust is folded around the north-plunging nose of the Rock Creek anticline; the thrust reappears in the area east of the Rock Creek anticline, where it is tightly folded into the core of the Rock Creek syncline and associated with subsidiary thrust faults and thrust slices (pl. 1, section *A–A′*).

Throughout most of the area, the Georgetown thrust places Middle Proterozoic rocks of the Helena Formation on Mississippian and Pennsylvanian formations, which indicates a stratigraphic offset of about 24,000 ft (7,300 m) based on an estimated total thickness for the Helena Formation and the Missoula Group of about 19,000 ft (5,800 m) and a thickness for the Paleozoic rocks of about 5,000 ft (1,500 m). Poulter (1956) estimated a minimum stratigraphic separation of 22,500 ft (6,860 m) along the Georgetown thrust, and he estimated a minimum of 4 mi (6.4 km) horizontal displacement to the east-southeast along the fault. Hyndman (1980) estimated about 37 mi (60 km) of eastward movement of the Sapphire tectonic block, and he interpreted the Georgetown thrust as a principal bounding-fault of that block.

Locally, subsidiary thrust faults merge with the Georgetown thrust and bound lensoid thrust slices of Paleozoic rocks of the lower plate. The most prominent of these subsidiary thrust faults, first mapped by Poulter (1956), bound thrust slices in the core of the Rock Creek syncline (pl. 1, section *A–A′*) where brecciated and fractured Devonian and Mississippian limestone and dolomite occur in fault slices beneath the Helena Formation and above Pennsylvanian quartzite. A thrust slice of Cambrian rocks also occurs along the western part of the fault, about a half mile south of Blue Grotto Spring in the northwestern part of the area.

East Fork Thrust and East Fork Branch Thrust

The East Fork thrust and the East Fork branch thrust are nearly bedding-plane faults that were folded harmonically with both the strata above and below them and with the overlying Georgetown thrust. The East Fork thrust and branch thrust occur entirely within the Proterozoic sequence where they place younger rocks on older ones.

The East Fork thrust can be traced across most of the study area of plate 1. The East Fork thrust puts the upper part of the Garnet Range Formation (Middle Proterozoic) and overlying Paleozoic rocks on the lower part of member two of the Mount Shields Formation (Middle Proterozoic). The thrust can be traced for approximately 11 mi (17 km) in the area, and for about 1 mi (1.6 km) directly east of the area, where it is truncated by the stock of Storm Lake (Wallace and others, in press). Approximately 6 mi (10 km) east of the area, the fault reappears on the east side of the stock (J.E. Elliott unpub. mapping). In the southwestern part of the area, the East Fork thrust is cut by the northern border of a granodiorite stock, but the thrust is also present south of the stock; and Wallace and others (in press) have shown that the thrust extends an additional 2 mi (3.2 km) south of the study area, where the thrust is folded into a syncline and cut by thrust faults of the imbricate zone.

The East Fork branch thrust probably is a subsidiary fault of the East Fork thrust, and the branch thrust is a separate fault only in the southwestern part of the study area; it merges with the East Fork thrust near the junction of Spruce and Page Creeks. The branch thrust places the upper part of member one and lower part of member two of the Mount Shields Formation over older rocks of the lower part of the Snowslip Formation. Wiswall (1976, p. 42) first mapped and described the southernmost segment of the branch thrust, which he erroneously called the Georgetown thrust, from exposures about 2,000 ft (600 m) south of West Pintlar Peak southwest of the study area, where the fault is folded into a syncline.

The fault zones of the East Fork thrust and the East Fork branch thrust vary in thickness and in character of deformation, apparently depending largely on the lithology of the rocks involved. In argillaceous rocks the zones of deformation commonly are thicker than in quartzitic rocks. In the central part of the area, in the headwall and eastern sidewall of the cirque that contains Spruce Lake, where both thrusts are exposed, zones of sheared and cleaved rocks are a few feet to a few tens of feet (1–10 m) thick along the East Fork branch thrust and occur mainly in argillaceous rocks of the Snowslip Formation beneath the fault. However, shear zones are only a few inches (several centimeters) thick along the nearby East Fork thrust, where quartzite lies above and below the fault. Also along the East Fork thrust in this area, thin quartz veins and small quartz pods are present in quartzite of the Mount Shields Formation in the footwall, but these veins and pods do not cross the thin, 1-in. (2.5-cm)-thick shear zone into quartzite of the Garnet Range Formation in the hanging-wall. Along the branch thrust, in the eastern sidewall of the cirque that contains Spruce Lake, quartz veins and pods occur in quartzite of the Mount Shields Formation in the hanging-wall but not in the underlying Snowslip Formation of the footwall. In the northeastern part of the area, along the ridge directly northwest of Rainbow Mountain and in the vicinity of the window along the east flank of the Dry Creek anticline, zones as much as 100 ft (30 m) thick of intensely sheared and partly recrystallized protomylonitic rock occur along the East Fork thrust both in the Snowslip Formation and in quartzite of the Garnet Range Formation. In general, zones of cataclastic rock along the East Fork thrust are thickest east of where the East Fork thrust and branch thrust appear to merge and become a single fault in the central part of the area.

Because the East Fork thrust and the East Fork branch thrust are essentially bedding plane structures that omit stratigraphic section and display younger-on-older stratigraphic relations, evidence for significant amounts of thrust-related displacement along them can only be based on regional structural and stratigraphic information. Regional structural and stratigraphic relations, discussed in the following sections of this report, suggest to us that the East Fork thrust is an extensive, laterally continuous fault that modified the apparent thickness and the stratigraphic sequence of the Missoula Group. Although the amount of displacement along the East Fork thrust is not known, the timing of any significant slip along the thrust is well constrained as predating the formation of large-scale folds, because the East Fork thrust is folded equally with the Georgetown thrust; this suggests that the East Fork thrust formed relatively early during Late Cretaceous compressional deformation, prior to later compression that formed the large-scale folds.

Cutaway Thrust

The Cutaway thrust is exposed in the southern part of the area, where it has an east-northeasterly strike and north-northwesterly dip. This thrust was not previously mapped although its stratigraphic position coincides with the "zone of bedding-plane thrust" that Poulter (1959, p. 24, fig. 2) depicted at the base of his Newland Formation and above his Ravalli Formation. The Cutaway thrust is not folded in the study area, and consequently its surface trace is much less sinuous than the traces of the folded Georgetown and East Fork thrusts. Emplacement of the stocks in the southern part of the study area apparently tilted the country rocks and associated structures to the north, accounting for the present north-northwesterly dip of the Cutaway thrust and the prominent northerly plunge of folds overlying the thrust. The northerly plunging, large-scale folds above the Cutaway thrust are disharmonic above the thrust, which suggests that the thrust acted as a basal detachment that accommodated large-scale folding in a manner similar to that described by Dahlstrom (1977) for buckled concentric folds, or in a manner similar to that described by Suppe (1983) for fault-bend folds.

The Cutaway thrust puts the older Helena Formation on younger rocks of the Mount Shields Formation. These stratigraphic relations suggest a minimum stratigraphic separation of about 8,000 ft (2,440 m). The Cutaway thrust can be traced for about 5 mi (8 km) along the crest of the Anaconda Range in the southeastern part of the area, and for another 3 mi (5 km) east of the area (Wallace and others, in press), for a total length of about 8 mi (13 km). South of Cutaway Mountain, the Cutaway thrust splits to form a branch thrust in the upper plate that is entirely within the Helena Formation. In the southern part of the area, and in several places east of the area (J.E. Elliott, oral commun., 1988), the Cutaway thrust is cut by Late Cretaceous and early Tertiary stocks.

The Cutaway thrust, best exposed south of Cutaway Mountain and along the ridge between Queener and Rainbow Mountains, is marked by an uneven zone of sheared and cleaved rock that ranges from as much as 20 ft (6 m) thick to

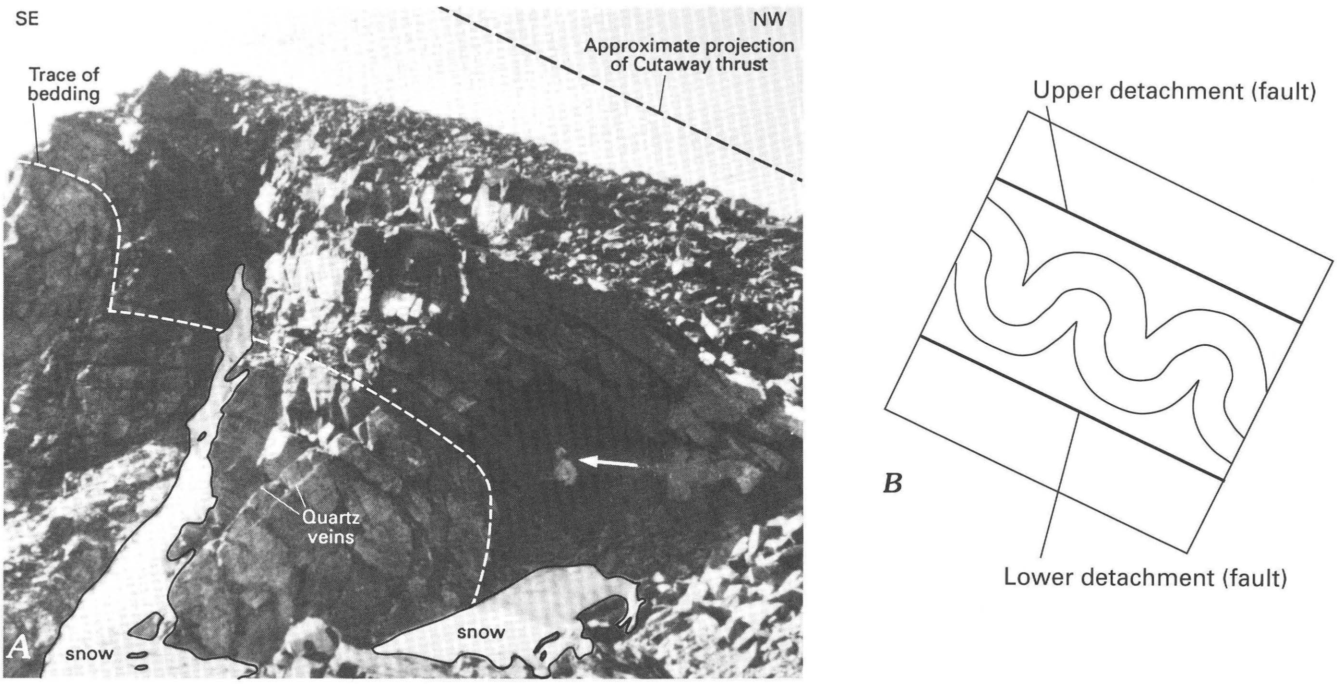


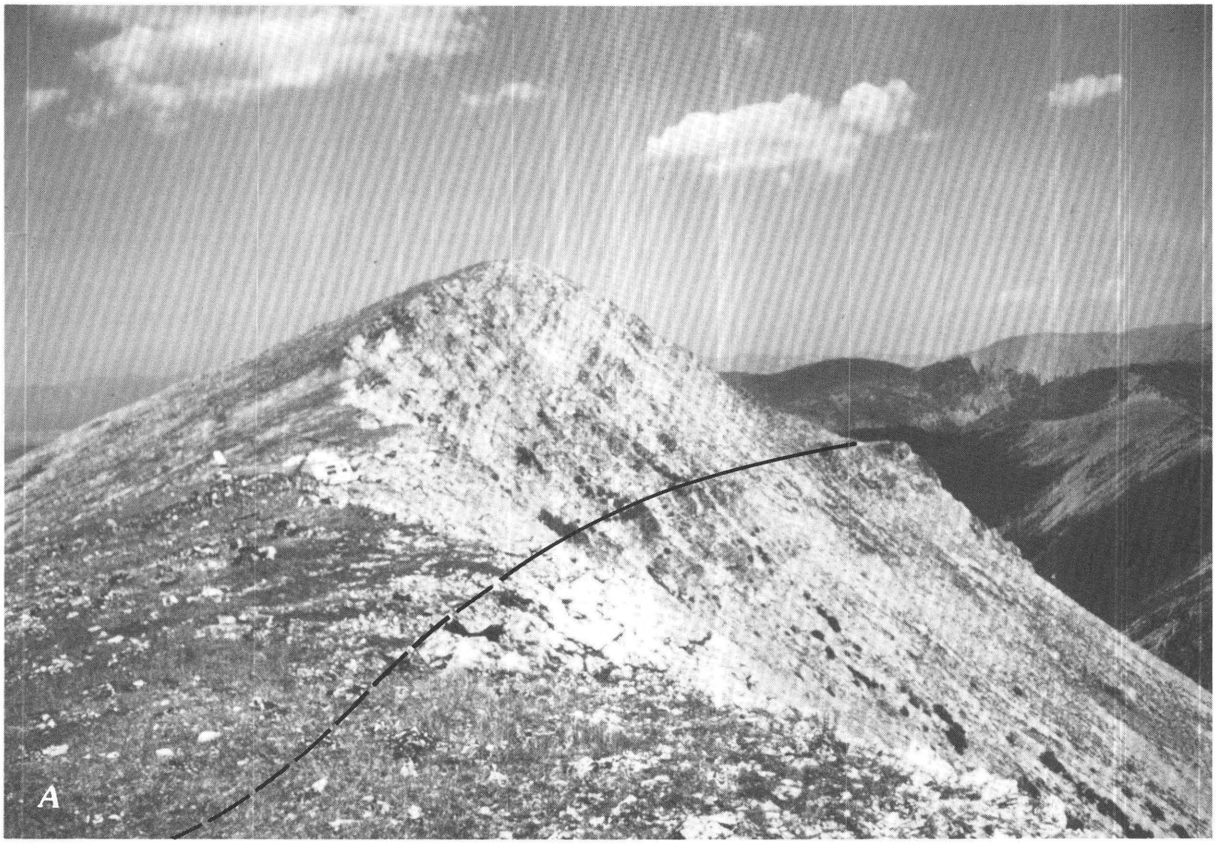
Figure 4. Parallel folds in the Mount Shields Formation (between Queener and Rainbow Mountains). *A*, Mesoscopic folds in member one of the Mount Shields Formation beneath the Cutaway thrust; the folds trend north-northeast, but are tilted to the northwest. *B*, Relation of upper and lower detachments to ideal concentric folds (modified from Dahlstrom, 1977). Cuspate syncline and broad anticlines of *A* suggest the presence of an upper detachment nearby above the folds, and their form reflects that of the upper part of the ideal folds depicted in *B*. The Cutaway thrust probably is the upper detachment for these folds.

as little as a few inches (several centimeters) thick. Between Queener and Rainbow Mountains, mesoscopic folds in the Mount Shields Formation beneath the thrust have amplitudes and wavelengths of about 50–200 ft (15–60 m), and based on the form of the anticlines and synclines the folds appear to be disharmonic beneath the Cutaway thrust (fig. 4). Similarly, large-scale folds above the Cutaway thrust, such as the Dry Creek anticline and the Rock Creek syncline, do not involve rocks beneath the thrust. The branch of the Cutaway thrust, which displaces beds in the Helena Formation, is well exposed on the ridge south of Sauer Lake (fig. 5). At this place, the position of the fault is marked by an increase in crenulation and shear and especially by a small but consistent angular discordance between strata above and below the fault.

Thrust in Southeastern Part of Area

The thrust structurally beneath the Cutaway thrust in the southeastern part of the area may be a small segment of a once-more-extensive sheet thrust, or it may be an imbricate thrust. This thrust trends northeast, dips moderately northwest, and can be traced a distance of about 4 mi (6.4 km). At its southwest end the thrust is cut by a monzogranite stock, and its northeast end appears to be cut by another. This thrust puts member one of the Mount Shields Formation on member two, indicating a minimum stratigraphic offset of about 4,000 ft (1,220 m) based on the minimum thicknesses of these members in this area. At most places the fault is covered, but sheared argillitic rocks in the hanging-wall were seen in several places. Other segments of thrusts that structurally underlie the Cutaway thrust are recognized elsewhere in the Anaconda Range, south and east of the study area (Wallace and others, in press; J.E. Elliott, oral commun., 1986). Those segments of thrusts are separated from the thrust present beneath the Cutaway thrust in the study area by one or more stocks, and some segments occur in roof pendants. Although structural, stratigraphic, and plutonic relations are not entirely clear, the thrust in the southeastern part of the study area and the discontinuous segments of thrusts south and east of the area might all have been part of a single, extensive sheet thrust beneath the Cutaway thrust.

Figure 5 (facing page). Branch of Cutaway thrust (about 8,500 ft southwest of Cutaway Mountain); Helena Formation is thrust on Helena Formation; note angular discordance in bedding of the Helena across the fault. *A*, fault looking north, dashed where approximately located (helicopter for scale); *B*, fault looking west. Fault dips about 35°–40° WNW.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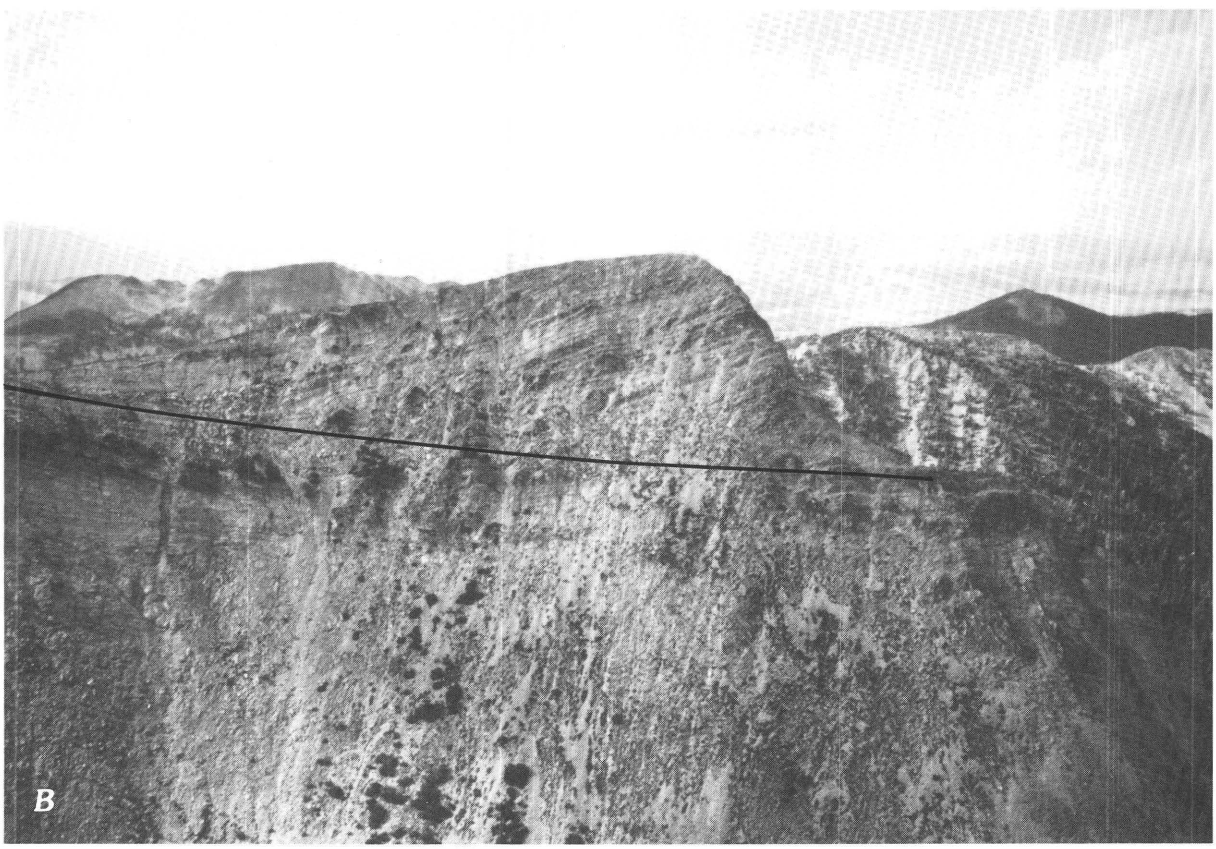
B

Imbricate Thrusts

A zone of anastomosing, imbricate thrust faults cuts tight and isoclinal folds between the western part of the Georgetown thrust and the core of the Rock Creek anticline in the western part of the area. The zone of imbricate faults trends north-northeast. Individual faults in the zone dip 25°–75° west-northwest. Apparent offset along individual faults ranges from about 500 to 5,000 ft (150 to 1,525 m), based mainly on stratigraphic separations in the tight and isoclinal folds. Individual fault strands are poorly exposed, but in a few places coherent and noncoherent breccia, gouge, and some boxwork breccia occur in shear zones of the faults. The zone of imbricate thrust faults extends for about 9 mi (14 km) across the study area; the zone is traceable for about 8 mi (13 km) southwest of the area, where some of the faults are truncated by Late Cretaceous stocks and where some of the faults appear to join a segment of the Cutaway thrust (Wallace and others, in press). Directly north of the area, Tertiary and Quaternary deposits mask any continuation of the zone.

Some of the imbricate thrust faults cut folded sheet thrusts. Directly north and south of Carpp Creek, in the southwestern part of the area, segments of the folded East Fork thrust occur in thrust slices of the eastern part of the imbricate zone. The western part of the imbricate zone continues southwest of the study area and cuts hanging-wall and footwall rocks of the Georgetown thrust (Wallace and others, in press). The Georgetown and East Fork thrusts are tightly folded south of the area along the continuation of this zone (Wallace and others, in press), and those folds are offset by thrusts (Wiswall, 1976, p. 43–45, and fig. 15). Several of the imbricate thrusts put younger rocks on older rocks because the rocks were tightly folded before faulting (pl. 1, sections *A–A′*, *B–B′*). Relations among folds and thrusts in the imbricate zone and the sequence of folding and faulting in this zone are discussed in more detail later in the report.

Out-of-Syncline Thrusts

In the central part of the area, several out-of-syncline thrust faults occur along the east limb of the Rock Creek syncline, and less prominent out-of-syncline thrusts occur along the east limb of the Spruce Creek syncline, and probably along the Porter Ridge syncline. These relatively small, fold-related thrusts duplicate or omit rock units along the east limbs of these synclines, but they have little or no expression in the west limbs. These faults may continue in the west limbs as one or more bedding-plane faults that acted as internal flexural-slip planes during parallel folding, but they were not identified in the west limbs. Out-of-syncline thrust faults and the imbricate thrust faults have some similar characteristics, such as: (1) coherent and noncoherent breccia and gouge in fault zones, (2) relatively small displacement, and (3) commonly younger rocks thrust over older; and like the imbricate thrusts, stratigraphic section is omitted along some of these faults where they cut strata that dip more steeply to the west than does the fault (pl. 1, sections *A–A′* and *B–B′*). Out-of-syncline thrust faults are commonly associated with concentric folds according to Dahlstrom (1977); the thrusts form during folding to accommodate bed-length inconsistencies produced by concentric folding, or form after folding is waning or has ceased, to accommodate continued shortening in a brittle manner.

Two sets of out-of-syncline thrusts occur along the east limb of the Rock Creek syncline and two others just northwest of the Rainbow Mountain fault: (1) the western set on the syncline cuts Upper Mississippian and Pennsylvanian rocks, (2) the eastern set on the synclinal limb cuts Cambrian and Devonian rocks, and (3) the two faults just northwest of the Rainbow Mountain fault cut the lower part of the Cambrian section and the underlying Garnet Range Formation (Middle Proterozoic). Poulter (1956) mapped one of the out-of-syncline thrusts of the western set as a down-to-the-east normal fault; however, the trace of the fault on both his map and our map shows that the fault dips to the west and is actually a reverse fault. Other normal faults on the flank of the Rock Creek syncline that duplicate parts of the Cambrian section, as mentioned by Poulter (1956), are here mapped as out-of-syncline thrusts.

An out-of-syncline thrust fault along the east limb of the Spruce Creek syncline puts the Silver Hill Formation (Middle Cambrian) on the Garnet Range Formation (Middle Proterozoic); the Flathead Quartzite is omitted (section *B–B′*). At this same locality the Hasmark Formation (Upper Cambrian) is faulted on the lower part of the Silver Hill Formation; all but the lower flaser-bedded shales of the Silver Hill Formation are omitted along this out-of-syncline thrust. A probable out-of-syncline thrust fault along the east limb of the Porter Ridge syncline places the Silver Hill Formation over the Garnet Range Formation and locally omits the Flathead Quartzite and the lower shales of the Silver Hill.

Folds

Large-scale, open, upright folds are present in the central and eastern parts of the area (Rock Creek and Dry Creek folds), and large-scale, tight to isoclinal, overturned folds, cut by the imbricate thrust faults, occur in the western part of the area. The Rock Creek syncline and Cable Mountain anticline (Dry Creek anticline of this report) were named by Emmons and Calkins (1913), and Poulter (1956) named Rock Creek anticline, Spruce Creek syncline, and Dry Creek syncline. We changed the name Cable Mountain anticline to Dry Creek anticline because the Dry Creek anticline is not continuous to Cable Mountain, which is north of the study area. Large-scale and small-scale folds trend north-northwest to north-northeast. The large-scale folds are dominantly parallel folds that generally have a concentric form in cross section. Both parallel and similar small-scale folds occur in the area and are most common in the anticlinal cores of large-scale folds.

Open Folds

The Rock Creek and Dry Creek anticlines, the Rock Creek, Porter Ridge, and Spruce Creek synclines, and the Sauer Creek anticline are large-scale, upright and open folds that plunge northward. These folds deform the Georgetown and East Fork thrusts but not the Cutaway thrust. The Rock Creek anticline and syncline and Dry Creek anticline are concentric folds in cross section; they are the largest and most prominent of the open folds, appear to be part of the same fold system, and have amplitudes of about 7,500 ft (2,300 m) and wavelengths of about 26,000 ft (8,000 m). The Spruce Creek and Porter Ridge synclines and the Sauer Creek anticline are less prominent, their wavelengths and amplitudes are not as readily determined, and their relations to the other large-scale folds are not entirely clear.

The Spruce Creek syncline trends northeasterly, has a smaller amplitude and wavelength than the Rock and Dry Creek folds, and appears to be a large-scale subsidiary fold that deforms the east limb of the Rock Creek anticline. The Spruce Creek syncline plunges north and northward merges into the west limb of the Rock Creek syncline. To the south, the Spruce Creek syncline becomes tighter, and a southward projection of the axial trace of the syncline suggests that it intersects the axial trace of the Rock Creek anticline, near the north side of a granodiorite stock. These relations suggest that the Spruce Creek syncline is a late-forming fold that developed on the broad folds, and it may be related more closely in origin and timing to the formation of the tight folds and the zone of imbricate thrust faults directly west of the syncline than to the period of formation of the open folds.

The Sauer Creek anticline is located between the Spruce Creek and Rock Creek synclines; it shares its west-dipping, west limb with the Spruce Creek syncline. Along the northern part of the fold, the east-dipping, east limb of the fold appears to be superposed on the larger west limb of the Rock Creek syncline to form a very gentle anticline between the Rock Creek and Spruce Creek synclines. Along the southern part of the Sauer Creek anticline, the east limb is not distinguishable from the west limb of the Rock Creek syncline, and consequently, the southern part of the Sauer Creek fold is mainly a north-plunging monocline. The Sauer Creek anticline merges with the Spruce Creek syncline in the valley of the East Fork of Rock Creek, and neither fold appears to extend further to the northeast. Like the Spruce Creek syncline, the Sauer Creek anticline may be a late-forming fold that was superposed on the west limb of the Rock Creek syncline.

The Porter Ridge syncline is an open, upright syncline that occurs near the southwest corner of the study area, south of the zone of imbricate thrusts and tight folds; it appears to be isolated from the other open folds. In its northern extent the Porter Ridge syncline is partly or entirely overridden by the zone of imbricate thrust faults.

Tight and Isoclinal Folds

Tight and isoclinal folds are poorly exposed in the western part of the area along the west flank of the Rock Creek anticline (pl. 1, sections *A–A′* and *B–B′*) in the zone of imbricate thrust faults. Amplitudes and wavelengths of the folds are not easily determined because they are overturned both to the east and west and are offset by faults of the imbricate zone.

Individual, unnamed folds in the northern part of the zone have been previously mapped (Emmons and Calkins, 1913; Poulter, 1956). Emmons and Calkins (1913, structure section *E–E′*) first showed the presence of a tight syncline and anticline west of the Rock Creek anticline in the northern part of the zone. Poulter (1956, structure section *B–B′*) showed thrust faults offsetting rocks in the core of the syncline and showed the folds as slightly overturned to the east at depth. Our mapping indicates that the folds are also overturned to the east at the surface, and we mapped more imbricate thrusts in the limbs and axial regions of the folds in the northern part of the zone than had previously been mapped. Tight folds and imbricate thrusts had not previously been mapped in the southern part of the zone; our mapping, however, indicates they are present in the southern part of the zone as well, but most of the folds are overturned to the west in this part of the zone. The axial trace of most of the tight folds in the zone is difficult or impossible to locate with precision, because exposures are poor and because the axial regions of many folds are faulted; furthermore, and particularly important in the southern part of the zone, the crests and troughs of the folds do not coincide with the hingeline of the folds (pl. 1, map and structure sections), which is characteristic of overturned folds (Hobbs and others, 1976, p. 169–170). Because of these complications, some special symbols are used in this zone (pl. 1) to indicate certain relations, such as a thrust in the same approximate position as the axial trace of a fold; and locally, a symbol is used for overturned folds that indicates the dip direction of rocks in the hinge region of the fold.

The tight and isoclinal folds are overturned to the east in the northern part of the zone of imbricate thrusts (pl. 1, section *A–A′*), but in the southern part, near Carpp Ridge, the folds are overturned to the west (pl. 1, section *B–B′*). The opposed direction of vergence of folds along strike suggests that northern and southern parts of the zone responded, in

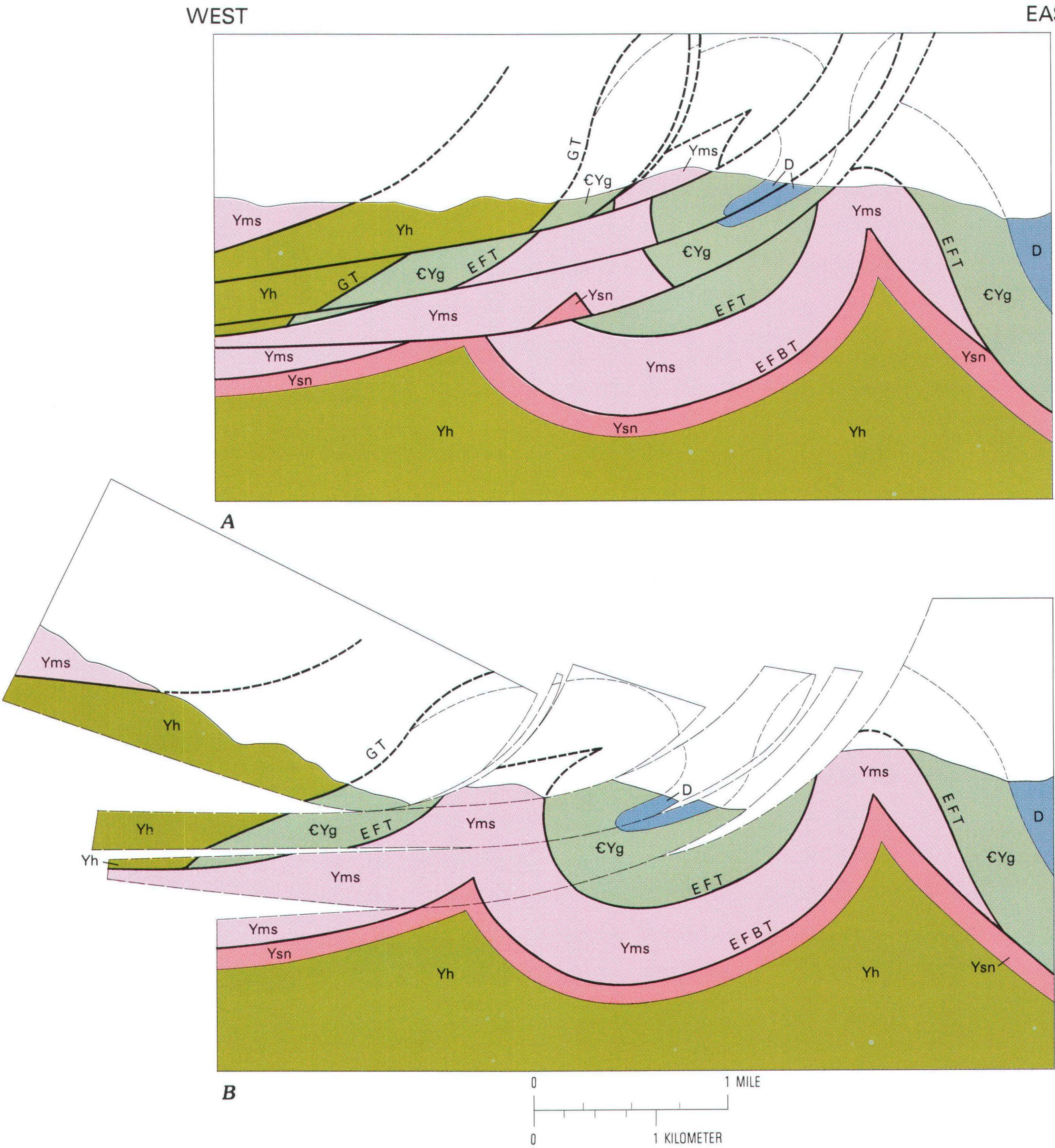

Figure 6. Restoration of footwall to hanging-wall relations along imbricate thrust faults in north part of zone of imbricate thrust faults and tight folds (generalized from western part of section *A–A′*, pl. 1). No vertical exaggeration. *A*, configuration after folding and imbrication; *B*, configuration of folds after footwall to hanging-wall offsets are restored across imbricate thrust faults. Restoration suggests that folds were tightened and overturned to east prior to much movement along imbricate thrust faults that cut and offset fold segments. Compare with restoration (fig. 8) of southern part of zone of tight folds and imbricate thrusts. D, Devonian rocks; €Yg, Cambrian rocks and Garnet Range Formation; Yms, Mount Shields Formation; Ysn, Snowslip Formation; Yh, Helena Formation; GT, Georgetown thrust; EFT, East Fork thrust; EFBT, East Fork branch thrust. Light lines, contacts; heavy lines, faults. Heavy and light short-dashed lines, faults and contacts projected above topographic profile. Long-dashed lines, lines of imbricate thrust fault surfaces after restoration of folded surfaces in hangingwalls of imbricate thrusts to matching folded surfaces in footwalls.

part, differently to the combination of folding and imbricate thrusting. In general, relations between the tight folds and the thrusts in the zone suggest that the tight folds in the northern part of the zone were ductilely overturned to the east and then cut and offset by the imbricate thrust faults. In the southern part of the zone, the tight folds were not ductilely overturned to the east; instead, they were cut by the imbricate thrusts and moved eastward up the steep-dipping, listric fronts of those faults, which apparently caused the folds to rotate counterclockwise and overturn to the west. These changes along strike in the geometry of the folds and the relative position and importance of thrusts are similar in many respects to the characteristics of Dahlstrom's (1977, p. 422) "transfer zones," which are zones of folds and thrusts that collectively accommodate an equal amount of shortening along the entire line of strike in each zone, the shortening being accommodated by varying amounts of folding versus thrusting along the strike.

Relations of the imbricate thrust faults to the tight east-verging folds in the northern part of the zone are simpler to analyze and restore than are relations of the imbricate faults and tight west-verging folds in the southern part. In the northern part of the zone, the folds appear to have been tightened and overturned to the east prior to much movement along the imbricate faults, as shown in the partial palinspastic restoration (fig. 6*B*); later, the imbricate faults cut and offset the limbs and axial regions of the folds (fig. 6*A*).

In the southern part of the zone, relations among the tight folds and imbricate thrusts are more complicated than those in the northern part, and the most obvious complication is an unusual but characteristic west vergence of the folds (pl. 1, section *B–B′*). The relations among the west-verging folds and west-dipping thrusts suggest to us that the folds formed initially as nearly upright folds and later acquired their west vergence when the folds were cut by imbricate thrusts and the folded fault blocks then moved up the more steeply dipping frontal parts of the imbricate thrusts. This movement up the frontal parts of the thrusts caused the folds to be rotated counterclockwise, and consequently overturned to the west (fig. 7). The relations among the west-vergent folds and west-dipping thrusts in the area are similar to the relations among some thrust faults and folds near Jasper, Alberta, where the west vergence of folds also is interpreted to be the result of rotation of folded fault blocks above steeply west dipping frontal parts of thrust faults (Mountjoy, 1959; Dahlstrom, 1977).

Another structural complication in the southern part of the imbricate zone involves an imbricate thrust in the back (west) part of the zone that apparently cut and offset previously rotated folds (fig. 8*A*, *B*); this relation implies, in part, a "break-back" rather than a "break-forward" (Morley, 1988) sequence of fault development for the southern part of the imbricate zone. Therefore, we interpret details of the sequence of folding and faulting in the southern part of the tight fold and imbricate zone, essentially as shown in the restoration (fig. 8): first, the rocks were broadly folded (fig. 8*D*); then, the folds were amputated from their bases by the imbricate thrusts, tightened, and overturned to the west by rotation of the folded fault blocks when they moved up the more steeply dipping frontal segments of the middle and eastern imbricate thrust (fig. 8*B*, *C*); the youngest deformation probably was in the back part of the zone where the western imbricate thrust cut and offset the folds that apparently had been previously rotated (fig. 8*A*). By analogy with the northward plunge of folds throughout the area and the northerly dip of the Cutaway thrust, the low-angle sole fault of the imbricate zone (fig. 8) may also have a northward dip and project to the surface in the region now occupied by the granodiorite stock in the southwest part of the area. If this interpretation is correct, then the Porter Ridge syncline lies beneath the sole fault and may be the base of a broad syncline whose upper part was amputated and further deformed in the fault blocks of the overlying tight fold and imbricate zone.

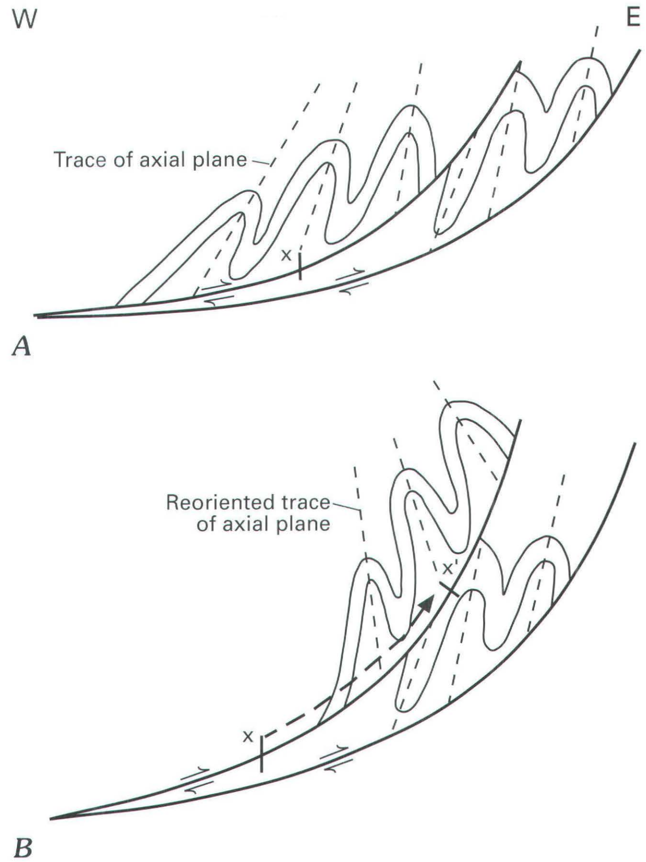

Figure 7. Generalized interpretation of the development of west-vergent folds, southern part of zone of imbricate thrust faults and tight folds. Barbs show direction of relative movement along faults. *A*, Formation of folds accompanied by some slip along listric thrusts. *B*, Rotation and overturning to the west of some folds as a result of movement of a folded fault block up the steeply dipping frontal part of a listric thrust fault. Axis of the fold at point X in *A* moved to point X′ in *B*.

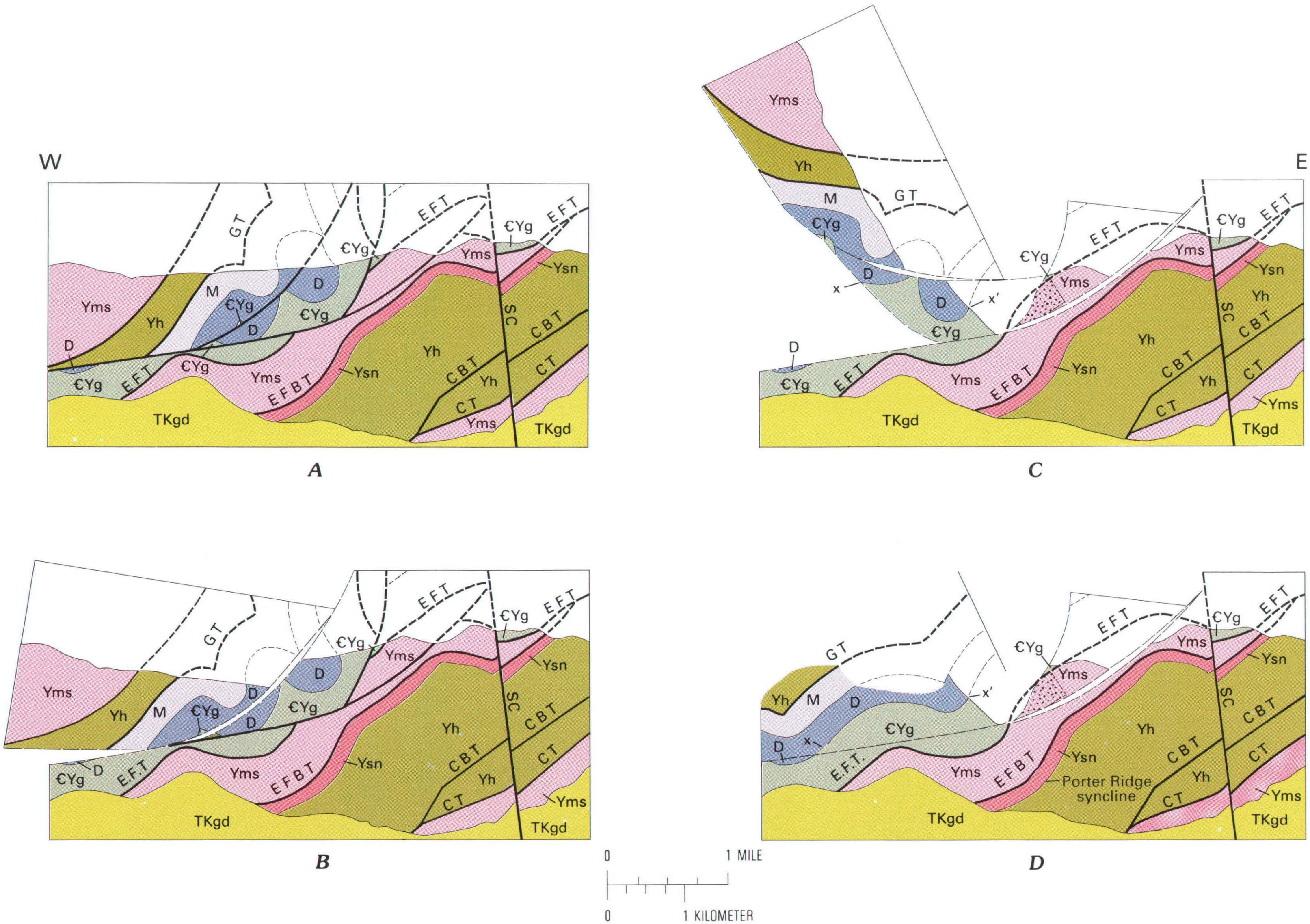
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Figure 8 (facing page). Sequential restoration of footwall to hanging-wall relations along imbricate thrust faults in southern part of zone of imbricate thrust faults and tight folds (generalized from western part of section *B–B′*, pl. 1). No vertical exaggeration. *A*, Present configuration after folding and imbricate thrust faulting. *B*, Configuration after restoration of footwall to hanging-wall offsets across imbricate fault in back part of zone. *C*, Configuration after restoration of footwall to hanging-wall offsets across two frontal imbricate thrust faults. *D*, Tight folds partly unfolded to match broad folds, such as the Porter Ridge syncline (see pl. 1 and discussion), southwest and structurally beneath zone of imbricate thrusts. Cambrian-Devonian contact (**x–x′**) is labeled in both *C* and *D* for comparison of line-lengths. Restoration sequence suggests the folds were tightened, but upright, when thrust up the steeply dipping parts of the frontal imbricate thrusts, which caused folds to be rotated counterclockwise and overturned to the west; rotated folds were then later cut and offset by the imbricate thrust fault in back part of the zone. TKgd, granodiorite stock; M, Mississippian rocks; D, Devonian rocks; ЄYg, Cambrian rocks and Garnet Range Formation; Yms, Mount Shields Formation; Ysn, Snowslip Formation; Yh, Helena Formation; GT, Georgetown thrust; EFT, East Fork thrust; EFBT, East Fork branch thrust; CT, Cutaway thrust; CBT, Cutaway branch thrust; SC, Spruce Creek fault. Light lines, contacts; heavy lines, faults. Heavy and light short-dashed lines, faults and contacts projected above topographic profile. Long-dashed lines, lines of imbricate thrust fault surfaces after restoration of folded surfaces in hanging-walls of imbricate thrusts to matching folded surfaces in footwalls. Stipple (*C, D*), area of overlap.

High-Angle Faults

High-angle faults are the youngest structures in the study area and offset the thrust faults, folds, sedimentary rocks, and granitic stocks. These faults are of Tertiary age, and most strike northeast, although west- and northwest-striking high-angle faults also occur. The high-angle faults have relatively small displacement compared with displacement on the thrust faults. Most of these faults appear to be normal faults, but some dip so steeply that their dip direction is difficult to discern. The Dry Creek fault, present in the eastern part of the area, shows about 1,500 ft (460 m) of stratigraphic separation, but most show only a few hundred feet (a couple hundred meters) or less. Coherent and noncoherent breccia and gouge occur along some high-angle faults, and some are marked mainly by zones of shear.

The Spruce Creek fault, which is best exposed in the saddle about 1 mi (1.6 km) southwest of Spruce Lake, is a north-northeast-striking, high-angle, normal fault that has a stratigraphic separation of about 400 ft (120 m) down on the eastern block; the mapped trace of the fault suggests that the fault surface dips moderately to steeply to the east. In the headwall of the cirque about 1 mi (1.6 km) southwest of Spruce Lake, several small northeast-trending faults are present but are not shown on plate 1; they downdrop small fault blocks a few feet to a few tens of feet (1–10 m) to the east. These small faults appear to be subsidiary to the Spruce Creek fault. Adjacent to the area, about 4 mi (6.4 km) south-southwest of the saddle, a high-angle fault is aligned with the Spruce Creek fault (Wallace and others, in press). Between the Spruce Creek fault and this high-angle fault adjacent to the study area, the granodiorite stock that lies across the trend of the two faults shows some northeast-trending foliation and alignment of porphyry bodies, which suggests that these faults may connect and cut the stock.

The Page Creek fault is a north-northeast-striking high-angle fault that offsets the stocks in the southernmost part of the area as well as the sedimentary sequence north of the stocks. This fault shows a stratigraphic separation of about 300–700 ft (90–215 m). About 2,000 ft (600 m) southeast of Cutaway Pass, in the south-central part of the area, the granitic stocks are sheared along the fault and a granodiorite stock is displaced against the monzogranite stock.

The Dry Creek fault is a prominent, arcuate, high-angle fault in the eastern part of the study area. The north end of the fault strikes north-northeast and the south end strikes north-northwest. In the saddle near the head of Dry Creek, the fault appears to be vertical or else to dip steeply to the west; the fault shows a stratigraphic separation of about 1,500 ft (460 m) down to the east. At Kurt Peak, in the southeastern part of the area, many small, high-angle faults intersect the Dry Creek fault at small angles; these faults form an internally faulted keystone graben. Farther to the southeast, beyond the area, the Dry Creek fault is represented by Elliott and others (1985) as a shear zone in a granodiorite stock. The fault-directed stratal tilts in the hanging-wall block of the Dry Creek fault and its arcuate trace suggest that it has a listric geometry, as shown in section *A–A′* (pl. 1).

West- and northwest-striking high-angle faults are not nearly as common as the northeast-striking faults. Most west- and northwest-striking faults that were mapped have a stratigraphic separation of only about 100 ft (30 m) or less. A zone of northwest-striking high-angle faults about 0.5 mi (0.8 km) southwest of Spruce Lake is marked by northwest-trending shear zones in rocks of the Belt Supergroup; thrust faults and rock units are downdropped on the northeast. The Rainbow Mountain fault, an east-striking high-angle fault, cuts the east limb of the Rock Creek syncline, cuts Belt and Cambrian rocks, and forms a local breccia and gouge zone. Geometric relations of the Rainbow Mountain fault with the out-of-syncline thrust faults differ north and south of the Rainbow Mountain fault, suggesting that this high-angle fault is a tear fault along which some left-lateral slip occurred during folding and movement along the out-of-syncline thrust faults. A west-northwest-striking high-angle fault in the northeastern part of the area cuts the nose of the Dry Creek anticline; this high-angle fault has a net slip of about 50–100 ft (15–30 m) down to the north.

NEW INTERPRETATIONS OF STRATIGRAPHY AND STRUCTURE

This report differs from earlier reports by Emmons and Calkins (1913), Poulter (1956), and Wiswall (1976) in two principal aspects: (1) we subdivide units of the Missoula Group below the Flathead Quartzite and above the Helena Formation, which requires new structural interpretations, and (2) we identify rock units below the Helena Formation along the Continental Divide as a southern lithofacies of the Mount Shields Formation, which also has structural consequences not previously recognized. The differences, then, result mostly from differences in stratigraphic identification, which are based on improved understanding of the regional stratigraphic framework in the Belt basin that was advanced by geologic maps of the Wallace, Choteau, Butte, and Dillon 1°×2° quadrangles (Harrison and others, 1986; Mudge and others, 1982; Wallace and others, 1987; Ruppel and others, 1983, respectively), and by the stratigraphic synthesis of Harrison (1972).

Missoula Group Below the Flathead Quartzite

Our changes to stratigraphic assignments of rock units below the Flathead Quartzite build on Poulter's (1956) revisions to Emmons and Calkins' (1913) stratigraphic assignments; Poulter (1956) recognized a thin Flathead Quartzite above rocks he assigned to the Missoula Group at places where Emmons and Calkins (1913) showed a thick Flathead Quartzite overlying their Spokane Formation (fig. 2). Although Emmons and Calkins (1913, p. 51–52) estimated a thickness of 140–200 ft (43–61 m) for the Flathead Quartzite, thicknesses calculated from their map locally exceed 500 ft (152 m) in the drainage of the East Fork of Rock Creek. Rocks similar to the Spokane Formation were included in the basal part of the Flathead Quartzite in some areas by Emmons and Calkins (1913, p. 51–52). Poulter's mapping resolved Emmons and Calkins' (1913) discrepancy between described and map thicknesses of the Flathead Quartzite by restricting rocks of the Flathead Quartzite to well-sorted, fine- to medium-grained, thickly bedded, light-grayish-white, light-gray, and tan, silica-cemented quartzite, and by assigning moderately to poorly sorted, fine- to medium-grained, mottled, grayish-green, grayish-red quartzite, argillaceous quartzite, and silty and sandy argillite to the underlying Missoula Group. Our revisions to the stratigraphic sequence in the study area (pl. 1) consisted principally of subdividing Poulter's (1956) Missoula Group into regionally recognized formations (Harrison, 1972).

The southwesternmost part of the area slightly overlaps previous mapping of Wiswall (1976). Wiswall included all the siliciclastic rocks above the middle Belt carbonate (Helena and Wallace Formations) in his Flathead Quartzite, and he concluded that rocks of the Missoula Group had been entirely eroded and an unusually thick Flathead Quartzite had been deposited above the unconformity, directly above the carbonate-rich rocks that he mapped as the Wallace Formation (Helena Formation of this report). We extended the Missoula Group as defined by Poulter (1956) into the area mapped by Wiswall, and found that Missoula Group rocks of the Garnet Range and Mount Shields Formations comprise most of what Wiswall had previously mapped as the Flathead Quartzite.

Our subdivision of the thin sequence of Missoula Group rocks directly beneath the Flathead Quartzite into (in descending order) the Garnet Range, Mount Shields, and Snowslip Formations (fig. 2) suggests that stratigraphic or structural discordances are present in the Missoula Group that were not previously recognized. The thin Missoula Group sequence was previously thought to have been reduced in thickness principally by erosion prior to deposition of the Middle Cambrian Flathead Quartzite (Emmons and Calkins, 1913; Poulter, 1956; Wiswall, 1976), implying that only the lower part remained of a Missoula Group sequence of unknown original thickness. Our identification of thin and apparently incomplete formations of the upper, middle, and lower parts of the Missoula Group underlying the Flathead Quartzite, combined with the absence of other regionally recognized, intervening formations of the Missoula Group, implies that erosion, represented by the unconformity at the base of the Flathead, did not cut very deeply into the top of the Missoula Group (fig. 2). These relations suggest to us two alternative explanations for the unusually thin Missoula Group, which includes the central lithofacies of the Mount Shields Formation: (1) the thin Missoula Group contains major unconformities between formations; or (2) folded thrust faults separate formations of the Missoula Group. We favor the latter interpretation, that folded thrust faults separate rock units of the Missoula Group below the Flathead Quartzite and above the Helena Formation, on the basis of local and regional relations discussed in the following paragraphs.

The possibility that local unconformities separate the Garnet Range and Mount Shields Formations, and separate the Mount Shields and the Snowslip Formations, is considered unlikely because: (1) no lithologic evidence of shoals is preserved, such as an abrupt change in quartzite-argillite ratios or change in the types of sedimentary structures that reflect changes in depositional currents; (2) no local, abnormally coarse grained rocks or lateral coarsening of clastic material occurs at the base of the formations that might record proximity to a local sediment source; and (3) no local changes occur in bedding style, bed thickness, or sedimentary structures that suggest proximity to a local strandline.

Rocks of the uppermost part of the Garnet Range Formation and rocks of the lower part of member two of the Mount Shields Formation (fig. 2) form the upper part of the abnormally thin and incomplete Missoula Group sequence that directly underlies the Flathead Quartzite in the study area. Quartzite, argillaceous quartzite, and interbedded sandy argillite are characteristic of the Garnet Range Formation in the study area and closely resemble lithologies of the uppermost part of the Garnet Range Formation in the Flint Creek Range to the north (Winston and Wallace, 1983) and the John Long Mountains to the northwest (Wallace and others, 1987) (fig. 1). Blocky, fine-grained, feldspathic quartzite and sparse, thin argillite beds and laminations are characteristic of member two of the Mount Shields Formation in the area and closely resemble lithologies in the lower part of member two in those nearby mountain ranges to the north and northwest, as well. Normally, however, as exposed in those mountain ranges, the uppermost part of the Garnet Range is separated from the lower part of member two of the Mount Shields by several thousand feet (a few thousand meters) of strata consisting of the remainder of the Garnet Range Formation, the McNamara Formation, the Bonner Quartzite, and member three and the upper part of member two of the Mount Shields Formation (fig. 2). The regional similarities among lithofacies argue against the presence of a shoal, proximal sediment source, or strandline in the vicinity of the Anaconda Range during deposition of rocks of the middle and upper units of the Missoula Group. Consequently, the presence of a major unconformity between rocks of the upper part of the Garnet Range Formation and the lower part of member two of the Mount Shields Formation is unlikely in the Anaconda Range.

Rocks of the upper part of member one of the Mount Shields Formation and rocks of the lower part of the Snowslip Formation compose the remainder of the thin Missoula Group sequence that directly underlies the Flathead in the area (fig. 2). These incomplete units of the Missoula Group are similar in bedding characteristics, grain size, sorting, mineral composition, and sedimentary structures to those same parts of the Mount Shields and Snowslip Formations described on Flint Creek Hill, 7 mi (11 km) north of the study area (Winston and Wallace, 1983, p. 76), and to those parts described from the southern Sapphire Mountains, about 10 mi (16 km) west and northwest of the area (Wallace and others, 1989). These regional similarities neither suggest shoals in the vicinity of the Anaconda Range during deposition of rocks between the upper part of member one of the Mount Shields Formation and the lower part of the Snowslip Formation, nor suggest the occurrence of a nearby strandline. Based on these local and regional stratigraphic characteristics of the Missoula Group, we have concluded that major unconformities within the Missoula Group are unlikely in the study area.

Thrust faults, then, best explain the thin or absent formations of the Missoula Group in the study area, although evidence that supports the occurrence of major interformational thrust faults is not conclusive because large separations cannot be demonstrated. Shear zones, truncated quartz veins, and cataclastic rock present along segments of the East Fork thrust and the East Fork branch thrust demonstrate that tectonic movement occurred along those zones. Poulter (1956) mapped a segment of the East Fork thrust within the study area. Wiswall (1976, p. 42) mapped a folded segment of our East Fork branch thrust that he called the Georgetown thrust southwest of the area, and he described stretched pebble conglomerate and sheared rocks (1976, p. 14 and p. 22) nearby, which occur along the contact between his Flathead Quartzite and Wallace Formation. Wiswall's sheared Flathead-Wallace contact was also mapped as a folded continuation of the East Fork branch thrust by Wallace and others (in press). The site of a photograph by Emmons and Calkins (1913, plate VIIA) that shows an unconformity between the Flathead Quartzite and their Spokane Formation in the cirque wall southwest of Spruce Lake could not be relocated, but in that same cirque wall we mapped a segment of the East Fork branch thrust that places rocks of the Mount Shields Formation over rocks of the Snowslip Formation. Based on our mapping, the angular discordance of 25°–35° between rocks of the Spokane Formation and Flathead Quartzite that was shown by Emmons and Calkins (1913, plate VIIA) probably occurs entirely within the Missoula Group, and it may result from tectonic dislocation and not from erosion. Elsewhere in this region the unconformity beneath the Flathead Quartzite generally is disconformable or shows only slight angular discordance in outcrops (Winston and Lonn, 1988, p. 20). The East Fork and East Fork branch thrusts best explain the thin and erratic Missoula Group section beneath the Flathead Quartzite, which juxtaposes the Mount Shields Formation over either the Snowslip or Helena Formations in the western part of the area, but juxtaposes the Garnet Range Formation over the Snowslip Formation in the eastern part of the area. Because these faults omit stratigraphic section, large thrust-related displacement along them is not obviously demonstrable. However, our interpretation agrees, in part, with structural and stratigraphic observations and interpretations of Poulter (1956) and Wiswall (1976), and we concluded that the Missoula Group rocks beneath the Flathead Quartzite contain extensive folded sheet thrusts, along which any large displacements could have occurred only prior to folding of these faults that occurred during the late pulses of Late Cretaceous thrust faulting and folding in this region.

Rock Units Below the Helena Formation Near the Continental Divide

Rocks mapped as the Newland Formation by Emmons and Calkins (1913) and Poulter (1956) are now known to be the Helena Formation, based on the stratigraphic position of these carbonate rocks below the Snowslip Formation and on lithologic similarity to rocks mapped as Helena Formation in the Sapphire Mountains and Flint Creek Range (Wallace and others, 1987). Emmons and Calkins (1913) assigned the Belt rocks beneath their Newland Formation (now the Helena) along the Continental Divide, to their Ravalli Formation (fig. 2). Poulter (1956) followed this assignment of rock units to the Newland and Ravalli Formations, and assigned the Belt rocks overlying his Newland Formation to the Missoula Group. That assignment of the overlying rocks to the Missoula Group placed the Newland Formation in the stratigraphic position now known to be occupied regionally by the Helena or Wallace Formations, as was recognized by Flood (1974) and Wiswall (1976). Although the Helena Formation overlies the Ravalli Group over most of the Belt basin, in the study area we identify rocks below the Helena Formation as the southern lithofacies of the Mount Shields Formation of the Missoula Group. This identification requires a near bedding-plane thrust fault, the Cutaway thrust, between the Helena Formation and underlying Mount Shields Formation. The lower member of the Helena Formation is absent above the Cutaway thrust in the study area, but about 1.5 mi (2.5 km) south of the area, west of Rainbow Lake, nearly 800 ft (250 m) of the lower member of the Helena overlies the Cutaway thrust (Wallace and others, in press). A bedding-plane thrust between the Helena Formation and the underlying rocks is not an entirely new proposal; Poulter (1959, fig. 2) showed a bedding-plane thrust zone at the base of his Newland Formation and above his Ravalli Formation. Furthermore, as discussed in following sections, the presence of a thrust at deeper levels in the stack of thrust sheets is predicted by mechanical requirements for parallel folding in the upper levels of the thrust stack.

The rocks below the Helena Formation near the Continental Divide are contact metamorphosed to gray, grayish-green, and grayish-black colors, and they superficially resemble rocks of the Ravalli Group; however, these rocks more closely resemble members one and two of the Mount Shields Formation, as exposed in the Sapphire and John Long Mountains and Flint Creek Range, than any formations of the Ravalli Group. The distinctive alternation of zones of rhythmically bedded, fining-upward, fine-grained quartzite, siltite, and argillite, with zones of well-sorted, feldspathic, planar-laminated, and ripple cross-laminated quartzite characterizes member one of the Mount Shields regionally (Wallace and others, 1989; Schmidt and others, 1983). These characteristics are clearly unlike the lithologies of the Empire and St. Regis Formations (Ravalli Group, fig. 2), which should stratigraphically underlie the Helena Formation (Harrison, 1972). The closest exposures of the Empire Formation, about 65 mi (106 km) northeast of the study area, consist of laminated and microlaminated, calcareous and dolomitic green argillite and siltite (Whipple and others, 1987). About 90 mi (145 km) northwest of the area, the St. Regis Formation consists of laminated and microlaminated purple and gray argillite and siltite and is dolomitic near the transitional contact with the overlying Helena Formation (Wells, 1974; Harrison and others, 1986). The thick-bedded, planar-laminated, and ripple cross-laminated feldspathic quartzite identified as member two of the Mount Shields Formation in the area also contrasts with the Empire and St. Regis Formations. The Spokane Formation (Harrison, 1972), which underlies the Empire Formation in a normal stratigraphic sequence in the eastern part of the Belt basin, shares some similarities with rocks of the Mount Shields Formation, as does the Revett Formation (Harrison, 1972), which underlies the St. Regis in the northwestern part of the Belt basin; but these similarities are overshadowed by lithologic differences. In the nearest exposures of the Spokane Formation, about 65 mi (106 km) northeast of the study area, the formation is characterized by fining-upward couplets 0.5–2 in. (1.3–5 cm) thick of grayish-red, calcareous and noncalcareous siltite overlain by laminated and microlaminated purple-red argillite separated by fine-grained sandstone and siltstone beds that are about 3–10 in. (7.5–25 cm) thick. The thickest sandy beds in the Spokane Formation occur about 235 mi (375 km) to the northwest of the area, where the middle part of the formation contains fine-grained quartzite beds as much as 50 ft (15 m) thick (Harrison and others, 1986). Thus, the Spokane Formation does not contain the alternating thick zones of well-sorted quartzite and rhythmically interbedded, fining-upward, fine-grained quartzite and laminated argillite that characterize member one of the Mount Shields Formation. About 65 mi (106 km) northwest of the area, the Revett Formation is characterized by upper and lower quartzitic members and a medial argillite and siltite member (Wells, 1974). The quartzite members are composed of fine-grained, purplish-gray and light-gray feldspathic quartzite that is interbedded with grayish-green siltite and argillite in zones that are generally less than 20 ft (6 m) thick, but rarely as much as 100 ft (30 m) thick. The middle member of the Revett Formation is composed of grayish-green, laminated and microlaminated, fining-upward argillite and siltite. The rhythmic alternation of zones of well-sorted, fining-upward, fine-grained quartzite with zones of red argillite characteristic of member one of the Mount Shields Formation is not common in the Revett Formation. Quartzite members of the Revett Formation may superficially resemble the quartzite of member two of the Mount Shields Formation, but member two does not contain thick zones of interbedded siltite and argillite; further, quartzite beds of member two characteristically have flat, aggradational basal contacts whereas quartzite beds of the Revett exhibit prominent channeled basal contacts and wedge-shaped beds. Member one and member two of the Mount Shields Formation contain calcareous and dolomitic beds, whereas rocks of the Revett Formation rarely contain carbonate rocks.

Poulter (1956 [map sheet]) stated that the "Newland and Ravalli Formations are in fault contact along the Continental Divide," but he did not expand on this observation. In a later report, Poulter (1959, fig. 2, p. 24) showed a zone of bedding-plane thrust at the base of his Newland Formation and above his Ravalli Formation. We mapped the Cutaway thrust in the same stratigraphic position as Poulter's thrust zone; the Cutaway thrust accounts for the presence of rocks of the Mount Shields Formation beneath older rocks of the Helena Formation along the Continental Divide, and explains the large-scale folds, which deform the thrust sheets above the Cutaway thrust.

INTERPRETATION OF STRUCTURE

The principal types of structures and the main elements of the sequence of Cretaceous and Tertiary deformation in the Anaconda Range were identified by Emmons and Calkins (1913). Subsequent studies in the Anaconda Range by Poulter (1956), Flood (1974), Wiswall (1976), Desmarais (1983), and Heise (1983) have added details and refinements to these initial observations. All these previous studies recognized multiple stages, or phases, of compressional deformation but differed on the number of stages or phases. Our interpretation of structures in the north-central Anaconda Range stresses the importance of the sheet thrusts, which we interpret as deformed segments of overthrusts, as a key to understanding the sequential development of polyphase compressional structures in this region.

Emmons and Calkins (1913, p. 141–142) summarized structural relations in the Philipsburg quadrangle, the southwestern part of which is this study area, and they identified the structures in the quadrangle as mainly (1) overthrusts, (2) folded overthrusts, (3) thrust faults, (4) high-angle faults (reverse and normal), and (4) large-scale folds. Their discussion of the relations between these structures implied a general sequence of compressional deformation that consisted of (1) overthrusting, (2) folding, and (3) more thrust faulting. The principal types of structures and the general three-fold sequence of deformation discussed briefly by Emmons and Calkins are now recognized as characteristic of the much larger thrust-sheet terrane of the Sapphire thrust plate (Lidke and Wallace, 1988). The Anaconda Range is a critical area in the entire thrust-sheet terrane because structural relations there indicate that (1) compressional deformation can be divided into three main phases, and (2) continued evolution of overthrusts accounts for structures of all three phases.

Kinematics of Thrust Faulting and Folding

The structure in the Anaconda Range is primarily the product of Late Cretaceous compressional deformation, and although the deformation probably was partly a continuum, it can be divided into three, partly overlapping phases based on relative age relations among the thrust faults and folds: (1) an early phase of overthrusting, during which thrust sheets were stacked; (2) an intermediate phase of folding, during which the stacked thrust sheets were folded, and (3) a final phase, during which some folds were tightened and overturned and offset along imbricate and out-of-syncline thrust faults. The early phase of overthrusting is represented by the folded sheet thrusts and folded branches of sheet thrusts; the Georgetown and East Fork sheet thrusts and branches were formed in this early phase. The intermediate phase of folding is represented by the large-scale parallel folds. The Georgetown and East Fork thrusts and adjacent thrust sheets, which formed during the early phase, were folded harmonically into the Rock Creek and Dry Creek folds during the intermediate phase. The final phase is represented by the tight and isoclinal folds of the imbricate zone and by the imbricate and out-of-syncline thrust faults that cut and offset the folds. The Cutaway thrust is a sheet thrust, but it formed (or slip occurred along it) during the intermediate phase (folding) because the Rock and Dry Creek folds are detached above it.

We will first discuss the relation of the Cutaway thrust to the second and third phases of deformation, because the relation of the folds to the Cutaway thrust can be used to predict geometric relations of the Cutaway thrust at depth. Large-scale parallel folds in thrust terranes geometrically require the occurrence of a fault beneath them; the fault serves as a lower detachment for large-scale buckle folding (Dahlstrom, 1977), or the fault is essentially an overthrust that forms fault-bend folds in the hanging-wall above ramps in the footwall (Suppe, 1983). Lidke (1985) initially interpreted the large, parallel folds that deform the Georgetown and East Fork thrusts in the study area as buckled concentric folds like those described by Dahlstrom (1977), and he inferred that the folds formed simply by buckling of the stacked, thrust sheets above an essentially flat Cutaway thrust. However, it is also possible that the Rock Creek and Dry Creek folds are mainly fault-bend folds as described by Suppe (1983), in which the anticlines approximately mark the positions of buried ramps in the footwall of the Cutaway thrust.

Recognition of ramps along the Cutaway thrust, particularly a ramp beneath the Dry Creek anticline, could provide a footwall cutoff (Suppe, 1983) along this thrust that broadly constrains a minimum restoration of the Cutaway thrust (fig. 9); however, a minimum restoration of the thrust, with that cutoff inferred, results in a mismatch of Missoula Group formations across this thrust. Although the restoration shown in figure 9 is imprecise and includes some inference, it illustrates the fundamental aspects of a minimum restoration for the Cutaway thrust, and it shows that a stratigraphic mismatch of Missoula Group units is unavoidable using standard techniques of minimum restoration. A minimum restoration of the Cutaway thrust (fig. 9) would incorrectly match two sequences of Missoula Group rocks that differ greatly; a thin Missoula Group sequence consisting only of the Garnet Range and Snowslip Formations in the hanging-wall of the

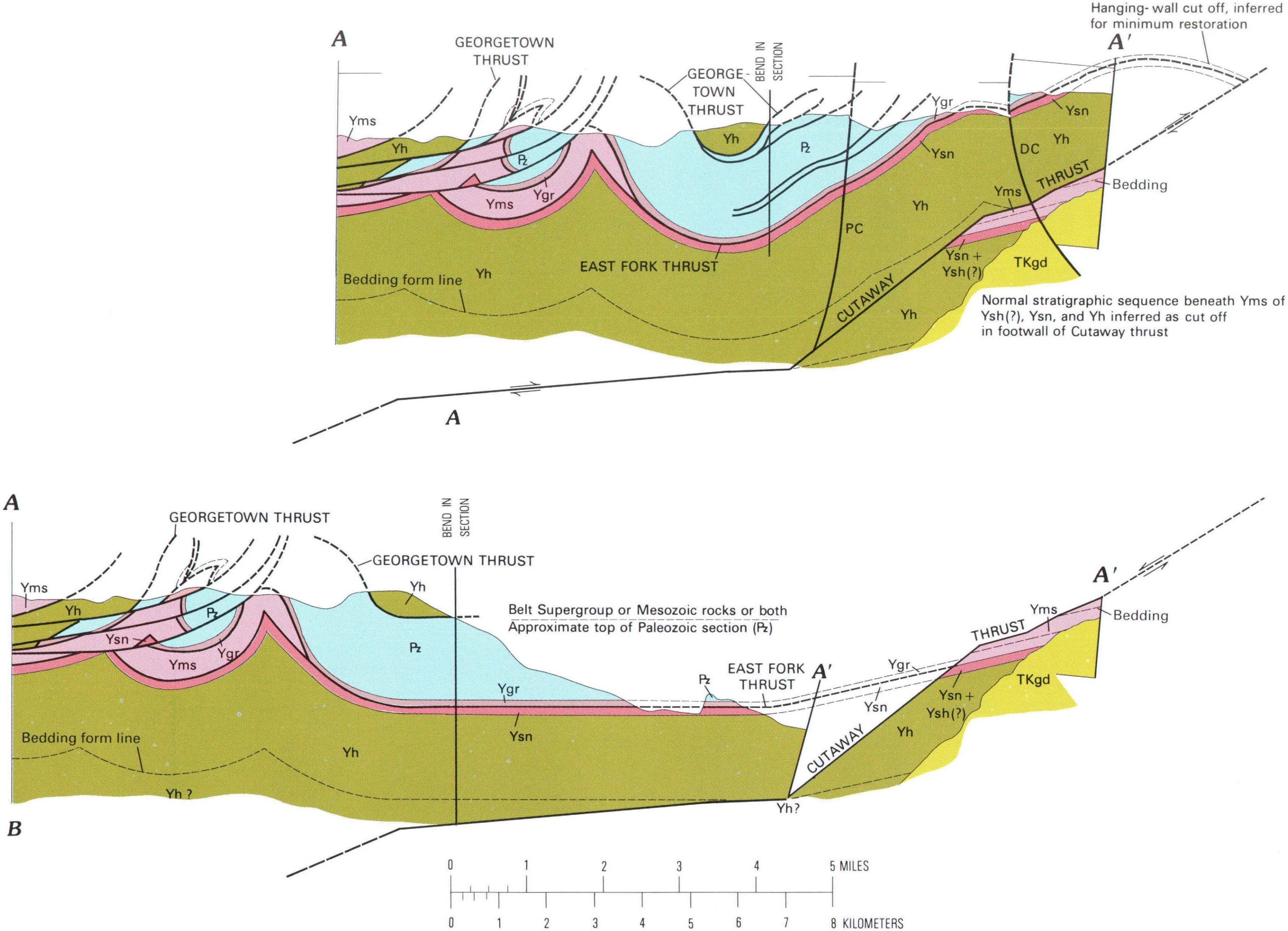
A
GEORGETOWN THRUST
GEORGE-TOWN THRUST
BEND IN SECTION
Hanging- wall cut off, inferred for minimum restoration
A′
Yms
Yh
Pz
Ygr
Ysn
DC
THRUST
Bedding
PC
EAST FORK THRUST
Ysn + Ysh(?)
TKgd
Bedding form line
CUTAWAY
Normal stratigraphic sequence beneath Yms of Ysh(?), Ysn, and Yh inferred as cut off in footwall of Cutaway thrust
A
A
GEORGETOWN THRUST
BEND IN SECTION
GEORGETOWN THRUST
Belt Supergroup or Mesozoic rocks or both
Approximate top of Paleozoic section (Pz)
EAST FORK THRUST
A′
Yh ?
Yh?
B
0 1 2 3 4 5 MILES
0 1 2 3 4 5 6 7 8 KILOMETERS

Figure 9 (facing page). Diagram showing that a minimum restoration of Cutaway thrust is an unlikely solution (generalized from section *A–A′*, pl. 1). No vertical exaggeration. *A*, Restoration of (1) footwall-to-hanging-wall offsets across Page Creek and Dry Creek high-angle faults and (2) positions of inferred hanging-wall and footwall cutoffs along Cutaway thrust, both of which are necessary to infer for the minimum restoration of the Cutaway thrust. *B*, Sketch of an approximate minimum restoration along the Cutaway thrust, which matches the Helena Formation and Missoula Group units across the footwall and hanging-wall of the Cutaway thrust. Missoula Group rocks of differing formations and of differing combined thicknesses are juxtaposed, which suggests minimum restoration to be untenable. Figure 10 presents a more tenable solution of Cutaway thrust. TKgd, granodiorite; Pz, Paleozoic rocks; Ygr, Garnet Range Formation; Yms, Mount Shields Formation; Ysh(?), Shepard Formation(?); Ysn, Snowslip Formation; Yh, Helena Formation; PC, Page Creek fault; DC, Dry Creek fault. Light lines, contacts; heavy lines, faults. Heavy and light short-dashed lines, faults and contacts projected above topographic profile.

Cutaway thrust does not match to a thicker Missoula Group sequence consisting of the southern lithofacies of member one and member two of the Mount Shields, and perhaps the Shepard and Snowslip Formations at depth, in the footwall of the thrust. Because this stratigraphic mismatch suggests structural or stratigraphic complications that techniques of a minimum restoration do not accommodate, and because none of the folded sheet thrusts exhibit hanging-wall and footwall relations that tightly constrain minimum restorations, these techniques were not used to interpret structural and stratigraphic relations among the stacked thrust sheets.

Our interpretation of thrust faults and folds in the north-central Anaconda Range uses graphical modeling techniques (Boyer and Elliott, 1982) to portray the development of a folded thrust-sheet stack within a multistage, ramp-flat system of overthrusts (fig. 10). In figure 10, stacked thrust sheets, folds, and late-stage thrust faults resulted from the sequential development of three overthrusts. The Georgetown, East Fork, and Cutaway thrusts are the three overthrusts shown, and they cross competent units abruptly to form ramps and flatten at incompetent stratigraphic horizons to form long flats, as is characteristic of overthrusts (Rich, 1934). The Georgetown, East Fork, and Cutaway thrusts are nearly parallel to bedding in Middle Proterozoic and Paleozoic rocks throughout the study area. The segments of these overthrusts that formed in the study area are shown as flat segments of overthrusts in figure 10 because bedding in the Middle Proterozoic and Paleozoic rocks was nearly flat lying when these faults formed. These flat overthrust segments of the study area are connected to inferred ramps at depth, west of the area, and these ramps cut bedding and locally cut flat segments of previously formed overthrusts (fig. 10). Those relations of the ramps to bedding and to flat fault segments caused stratigraphic sequences to be repeated or omitted above individual flat segments of faults within the stacked thrust sheets (fig. 10*D*). Branch thrusts of the Georgetown, East Fork, and Cutaway thrusts and thrust slices related to those branches are not shown, but the occurrence of the branches and slices in the study area suggests that parts of the long, flat fault segments include small ramps, at which the overthrusts branched to form the branch thrusts and thrust slices. Changes to inferred locations and angles of ramps that formed west of the area or additions of small ramps to flat overthrust segments that formed in the area do not alter the basic relations of the overthrust system shown.

The sequence of overthrusting and the crosscutting relations among the three overthrusts shown in figure 10 are together primarily responsible for the arrangement of fault segments in the thrust-sheet stack that formed (fig. 10*D*); although that stack resulted from a single repetitive sequence of overthrusting, the stack contains five flat segments of the three overthrusts, which collectively show both "in-sequence" and "out-of-sequence" thrust relations as described by Morley (1988). In-sequence thrust relations occur in the stack where younger faults underlie older faults, and out-of-sequence thrust relations occur where younger faults overlie older faults. The most typical thrust-fault sequence is in-sequence (or break-forward) development of new thrust faults (Morley, 1988). New thrusts progressively develop in front of, or beneath, the preceding thrust fault by undercutting the footwall of the preceding thrust to form a new thrust fault, as was illustrated in the duplex models of Boyer and Elliott (1982). In figure 10, the western part of each new overthrust (East Fork and Cutaway thrusts) developed in a position that was in-sequence, because the western part formed in front of, and beneath, the preceding thrust by undercutting the footwall-ramp of the preceding thrust. In contrast, however, the eastern parts of the East Fork and Cutaway thrusts formed out-of-sequence thrust relations where these faults cut across, stepped over, and duplicated the preceding thrusts. Consequently, in the thrust-sheet stack that formed (fig. 10*D* and *E*), the Cutaway thrust and overlying western segments of the Georgetown and East Fork thrusts show in-sequence relations, whereas the Cutaway thrust and underlying eastern segments of the Georgetown and East Fork thrusts show out-of-sequence relations.

Out-of-sequence thrusts commonly produce atypical stratigraphic relations (Morley, 1988), yet development of out-of-sequence thrust faults is a predictable consequence of the continued sequential formation of new thrust faults in hinterland thrust belts (Boyer and Geiser, 1987). Omission of stratigraphic section along the East Fork thrust (fig. 9) is a logical result of out-of-sequence thrusting, as is the stratigraphic mismatch of a minimum restoration of the Cutaway thrust (figs. 9 and 10). The interpretation that the East Fork and Cutaway thrusts are in part out-of-sequence thrusts implies that (1) Belt Supergroup rocks now present beneath the East Fork thrust were initially part of the hanging-wall of the Georgetown thrust before being stepped over by the East Fork thrust; (2) the Cutaway thrust represents, in part, a reactivated frontal segment of the Georgetown thrust that was

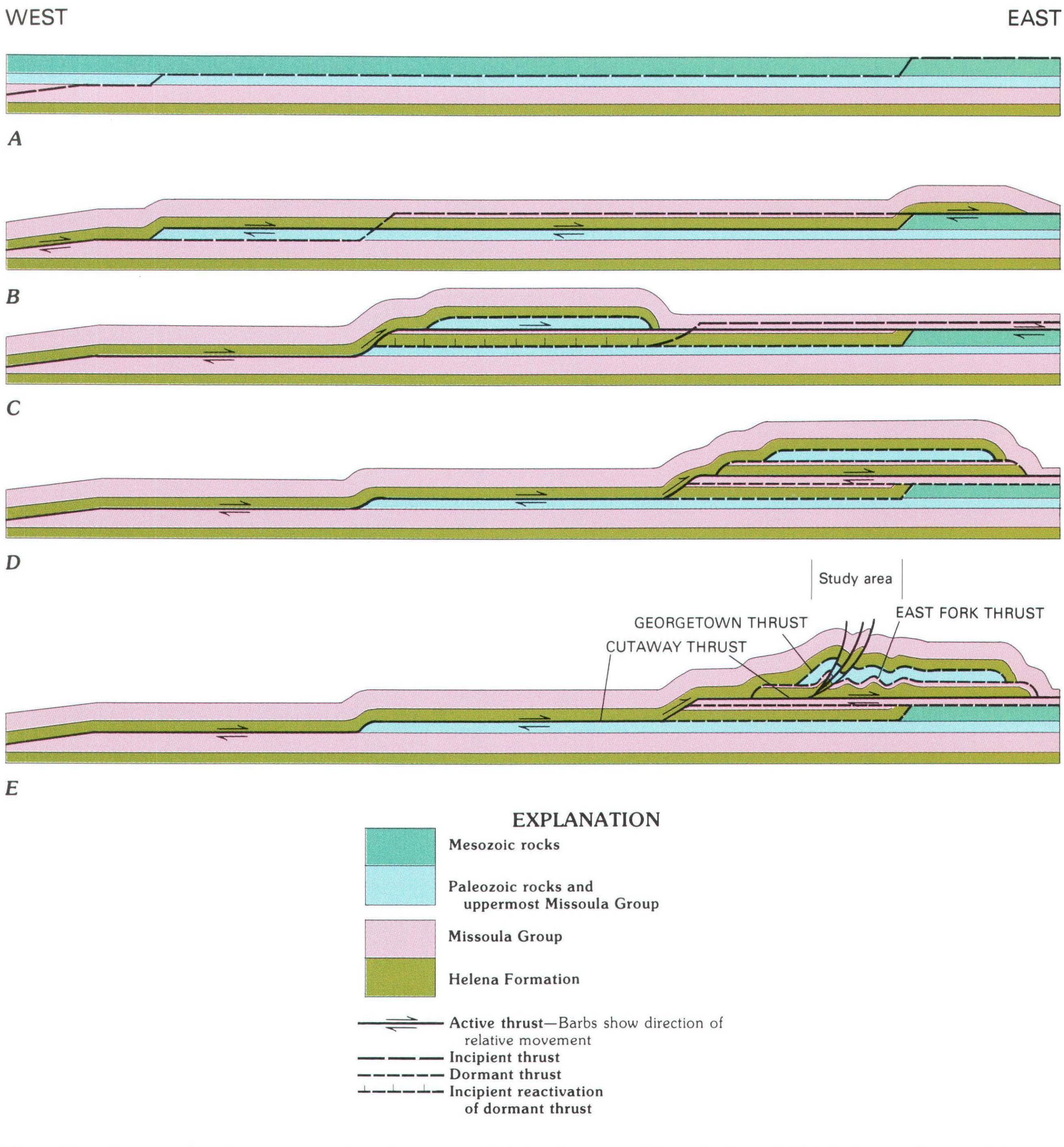

Figure 10. Diagram showing interpretation of the sequential development of thrust faults and folds in the north-central Anaconda Range. Scale is approximate; width of study area in *E*, about 9 mi (14.5 km); rock units are generalized. *A*, undeformed; *B*, movement along the Georgetown thrust; *C*, movement along the East Fork thrust; *D*, movement along the Cutaway thrust; *E*, folding and movement along the imbricate and out-of-syncline thrusts during some movement along Cutaway thrust.

stepped over by the East Fork thrust; and (3) segments of the Georgetown and East Fork thrusts exist structurally beneath the Cutaway thrust (fig. 10). The position and stratigraphic relations of the unnamed thrust in the southeastern part of the area are compatible with the segment of the East Fork thrust shown beneath the Cutaway thrust in figure 10.

Although out-of-sequence thrust faults can create complicated and atypical stratigraphic relations among thrust sheets, the interpretation of figure 10 is simple because all the thrust-sheet stacking, folding, and late-stage thrusting resulted from movement along only three overthrusts that formed in a predictable sequence and show a repetitive style (fig. 10*A–E*). In the interpretation, the East Fork thrust "piggy-backed" a western segment of the Georgetown thrust over an eastern segment of the Georgetown thrust, and the underlying eastern segment of the Georgetown thrust was then reactivated to form the Cutaway thrust, which like the East Fork thrust, cut and piggy-backed western segments of the Georgetown and East Fork thrusts over eastern segments of the Georgetown and East Fork thrusts. Finally, the upper part of the thrust-sheet stack buckled to form large-scale folds during some period of movement along the Cutaway thrust, and locally these folds were tightened, cut, and offset by the late-stage imbricate and out-of-syncline thrusts (fig. 10*E*). Although the large-scale folds may indicate ramp-flat relations along the Cutaway thrust at depth, which could have been primarily responsible for the formation and position of the folds, these folds are shown instead as buckled concentric folds above a flat segment of the Cutaway thrust for convenience of exposition. This interpretation of the folded thrust-sheet stack in the north-central Anaconda Range is also consistent with regional interpretation of thrust faults and folds in the Sapphire Mountains (Wallace and others, 1989) and with interpretations of structural relations in the John Long Mountains, the Flint Creek Range, and southern and eastern parts of the Anaconda Range (Lidke and Wallace, 1988).

CONCLUSIONS

The Anaconda Range provides a unique view of deep structural levels in the stacked thrust sheets that characterize the thrust-sheet terrane of the Sapphire thrust plate; the deep levels are exposed because the stack was uplifted and tilted to the north during, or following, emplacement of the stocks that form most of the crest and southern part of the range. The presence of a relatively undeformed sheet thrust, such as the Cutaway thrust, at deeper levels beneath sheet thrusts that are markedly folded and thrust faulted indicates that (1) the Cutaway thrust is younger than the higher, more deformed sheet thrusts; and (2) the folds and late-phase imbricate and out-of-syncline thrusts above the Cutaway thrust probably formed during, and as a consequence of, movement along the Cutaway thrust. These relations among the stacked thrust sheets, folds, and late-phase thrusts suggest that this region was first thickened and consequently shortened by easterly directed overthrusting to form stacked thrust sheets, and was later further shortened when the stack of thrust sheets was folded and internally thrust faulted during movement on deeper sheet thrusts. Characteristics of thrust sheets in the Anaconda Range and the thrust-sheet terrane of the Sapphire thrust plate (fig. 3) are similar to characteristics of thrust nappes (McClay, 1981, p. 7–9), and the style and sequence of shortening and deformation in these areas are similar to those determined for some nappe terranes.

Previous studies of the north-central Anaconda Range (Emmons and Calkins, 1913; Poulter, 1956) did not explain the formation of the large-scale folds that deform sheet thrusts, like the Georgetown thrust, probably because those studies did not identify the Cutaway thrust at deeper structural levels. Because of the northerly plunge of the folds, the bottoms of the folds and deeper structures are exposed in the southern part of the area near stocks along the crest of the range. In general, the deepest structures are exposed in the highest parts of the range. The Cutaway thrust, exposed along the crest of the range, dips northward beneath folded thrust sheets that overlie it. Movement along the Cutaway thrust explains the formation of the large-scale folds that involve the stacked thrust sheets structurally above. Based on folded and faulted, stacked thrust sheets observed in the Sapphire Mountains to the west and in the Flint Creek Range to the north, Lidke and Wallace (1988) predicted the occurrence of younger sheet thrusts at depth in those areas. The northerly dipping Cutaway thrust apparently projects northward and northwestward beneath the stacked and folded thrust sheets of those mountain ranges as well.

The relations among the stacked thrust sheets, large-scale folds, and late-phase thrust faults suggest that the thrust- and fold-related structural evolution of the area can be broadly divided into two main stages of shortening encompassing three phases of deformation, all of which is related directly to movement along sheet thrusts. The two stages of shortening are (1) an early stage, in which overthrusting thickened the stratigraphic section by stacking thrust sheets, and (2) a later, and partly overlapping stage, in which folding and late-phase thrust faulting compressed and shortened the stack of thrust sheets above the Cutaway thrust. Of these two stages, the early stage of thrust-sheet stacking probably produced by far the largest amount of regional shortening, but this shortening is less obvious in the study area than is the apparently smaller, but more directly observable amount of shortening attributable to folding and late-phase thrusting (compare fig. 10 with pl. 1, sections *A–A′* and *B–B′*). We define the first phase of deformation as synonymous with the first stage of shortening; the second stage of shortening, however, can be subdivided into two phases of deformation, folding followed by late-phase imbricate thrusting. This subdivision is warranted because the late-phase thrust faults truncate and offset the folds and commonly do not appear to be related to the formation of the folds in the manner that thrust faults typically are related to the formation of fault-propagation folds. Furthermore, some of the late-phase thrusts can be considered to be "out-of-sequence" with respect to the folds (Morley, 1988, p. 541) because they cut the same stratigraphic horizon twice in a single anticline or syncline (pl. 1, sections *A–A′* and *B–B′*), a relation which also explains local younger-on-older stratigraphic relations across some of these thrusts.

The characteristics and evolution of thrust sheets in the study area are similar to those of some thrust nappe terranes.

Thrust nappes in many parts of the European Alps, for example, were not recognized in early studies of those regions (Hobbs and others, 1976, p. 409), and some faults that bound thrust nappes were mapped as stratigraphic contacts because they were devoid of deformation (Billings, 1972, p. 203). Hobbs and others (1976, p. 408) speculated, from known examples from the European Alps and other nappe terranes, that incorrect stratigraphic sequences were mapped and accepted in some regions because stacked thrust sheets were not recognized. Similarly in the study area the Cutaway thrust was previously mapped as a stratigraphic contact, and the East Fork thrust and branch thrust were not recognized within the Spokane Formation of Emmons and Calkins (1913) and within the Missoula Group of Poulter (1956). The thrust sheets in the study area, and elsewhere in the Sapphire thrust plate, characteristically are laterally extensive, show internal stratigraphic continuity, and are separated by faults that generally parallel bedding; along parts of these faults the zone of deformation is thin or obscure. The general sequence of movement along sheet thrusts in the study area, in which the deeper faults are relatively younger than higher faults, is a common sequence of fault development in stacked thrust nappe terranes (Ramsey and others, 1983). In the thrust-sheet terrane of the Sapphire thrust plate and in several thrust nappe terranes, two main stages of shortening characterize the structural development: (1) an early stage of stacking thrust sheets (or thrust nappes), and (2) a later stage of folding and faulting of the stack. Nappe terranes characterized by this sequence of deformation include the Pennine nappes of the Alps and the thrust nappe complex of the Atnarpa Range in Australia (Hobbs and others, 1976, p. 411–415).

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