

Geology of the Elliston Region,
Powell and Lewis and
Clark Counties, Montana

U.S. GEOLOGICAL SURVEY BULLETIN 2045

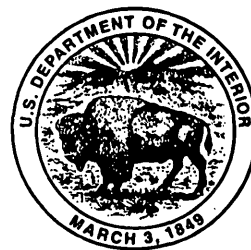


Geology of the Elliston Region, Powell and Lewis and Clark Counties, Montana

By R. G. SCHMIDT, J. S. LOEN, C. A. WALLACE, *and*
H. H. MEHNERT

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*A description of rocks and structures
in the region of the imbricate front
of the Sapphire thrust plate, from a
reconnaissance study*



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GEOLOGY OF THE ELLISTON REGION, POWELL AND LEWIS AND CLARK COUNTIES, MONTANA

By R.G. Schmidt¹, J.S. Loen, C.A. Wallace, and H.H. Mehnert

ABSTRACT

Rocks in the Elliston, Mont., region are mainly units of the Belt Supergroup (Middle Proterozoic), Paleozoic and Mesozoic sedimentary rocks, Tertiary sedimentary and volcanic rocks, and nonconsolidated Quaternary deposits. Cretaceous stocks intruded the older sedimentary rocks at several places in the study area. The frontal imbricate edge of the Sapphire thrust plate overlies the west limb of the northwest-trending Black Mountain syncline in the west-central part of the area. A prominent high-angle right-slip fault of the Lewis and Clark line, the Bald Butte fault, strikes northwest across the north part of the studied terrain.

The oldest sedimentary rocks are units of the Ravalli Group of the Belt Supergroup (Middle Proterozoic), and these rocks are exposed only north of the Bald Butte fault along the northern border of the study area. The Spokane Formation is the oldest unit in the study area, and the Spokane is overlain by the Empire Formation at the top of the Ravalli Group. Above the Ravalli Group is the Helena Formation, which is overlain by the Snowslip, Shepard, and Mount Shields Formations, the Bonner Quartzite, and the McNamara Formation, all of which are in the Missoula Group.

The Paleozoic sequence of limestone, dolomite, shale, and sandstone ranges in age from Middle Cambrian to Permian, and western lithofacies of these units are superposed by the Sapphire thrust plate over an eastern lithofacies. The Middle Cambrian Flathead Quartzite is the oldest unit in the western and eastern lithofacies. On the Sapphire thrust plate above the Flathead Quartzite, the Middle Cambrian Silver Hill Formation and the Upper Cambrian Hasmark and Red Lion Formations represent the western lithofacies. North and east of the Sapphire thrust plate above the Flathead Quartzite, the Middle Cambrian Wolsey Shale, Meagher Limestone, and Park Shale

and Upper Cambrian Hasmark and Red Lion Formations form the eastern lithofacies in ascending order. The Upper Devonian Maywood Formation unconformably overlies the Red Lion Formation. The Jefferson Formation, of Late Devonian age, overlies the Maywood Formation in eastern and western lithofacies. The Three Forks Formation (Lower Mississippian and Upper Devonian) overlies the Jefferson in the eastern lithofacies, but the Three Forks is absent in the western lithofacies. Mississippian rocks of the Madison Group overlie the Three Forks Formation in the eastern lithofacies and overlie the Jefferson Formation of the western lithofacies. Rocks previously assigned to the Amsden Formation are now assigned to the Snowcrest Range Group of Mississippian and Pennsylvanian age. Above the Snowcrest Range Group is the Quadrant Quartzite of Pennsylvanian age, and overlying the Quadrant is the Shedhorn Sandstone and Phosphoria and Park City Formations of Permian age.

Western and eastern lithofacies of Mesozoic formations are composed mostly of shale, siltstone, and sandstone, and these units range from Middle Jurassic to Late Cretaceous in age. The Upper and Middle Jurassic Ellis Group contains the Sawtooth, Rierdon, and Swift Formations, from bottom to top. The Upper Jurassic Morrison Formation is very thin in western lithofacies of the Sapphire thrust plate, but the Morrison Formation is thickest in the Black Mountain syncline east of the thrust plate. The Kootenai Formation (Lower Cretaceous) unconformably overlies the Morrison Formation and Ellis Group. The Blackleaf Formation is Early and Late(?) Cretaceous in age, and this unit overlies the Morrison Formation and Ellis Group. Above the Blackleaf Formation thrust faults have eliminated some units from the Upper Cretaceous succession, so only the Jens and Carten Creek Formation are present.

Tertiary and Quaternary sedimentary rocks and deposits are mostly volcanogenic claystone, siltstone, sandstone, and conglomerate deposited in intermontane

¹Deceased, 1983.

basins. Tertiary deposits are Oligocene, Miocene, and Pliocene in age, and four lithologic units are separated by unconformities. Youngest Tertiary or older Quaternary gravel deposits contain placer gold near Elliston, and alluvium contains placer gold.

Igneous rocks are widespread in the study area, and they consist of volcanic and plutonic rocks. Volcanic rocks consist of mafic- and intermediate-composition lava and breccia of the Late Cretaceous Elkhorn Mountains Volcanics, of intermediate lava and breccia of Late Cretaceous or early Tertiary age, of late Eocene rhyolite flows, and of Oligocene basalt flows. Plutonic rocks are mostly intermediate stocks of the Boulder batholith and isolated stocks of intermediate composition, all of which are Late Cretaceous in age. The Marysville, Blackfoot City, and Scratchgravel Hills stocks are mineralized, and they have produced gold from veins.

The main structures in the study area are thrust faults of the frontal imbricate zone of the Sapphire thrust plate, high-angle faults, and the Black Mountain syncline. Thrust faults of the frontal imbricate zone of the Sapphire thrust plate overlap the west limb of the Black Mountain syncline, and the thrust faults amputated a series of open folds west of the study area. The Bald Butte fault is a high-angle right-slip fault that displaced the northwest end of the Black Mountain syncline about 28 km to the southeast after 97 Ma and before 44–47 Ma. Normal faults, such as the Blackfoot City, Illinois Ridge, and Marsh Creek faults appear to have offset rocks during Late Cretaceous time, and these faults may have reactivated during Tertiary time.

The Garnet line, which has been described as a Middle Proterozoic growth fault, was projected through the study area, but data from this study and from maps of adjoining regions indicate that the Garnet line does not exist.

Much of the normal slip previously attributed to Basin-and-Range tensional faulting is instead Late Cretaceous and early Eocene in age. The essential elements of the modern topography predate middle Eocene volcanic rocks.

INTRODUCTION

A recent study of the Elliston area (fig. 1) presents new stratigraphic and structural data that refine older reconnaissance information (Bierwagen, 1964) and adds data in places where geologic mapping was absent or incomplete (fig. 2). The study area is underlain by sedimentary rocks and deposits that range in age from Middle Proterozoic to Holocene, by plutonic rocks mostly of Late Cretaceous and early Tertiary age, and by volcanic rocks

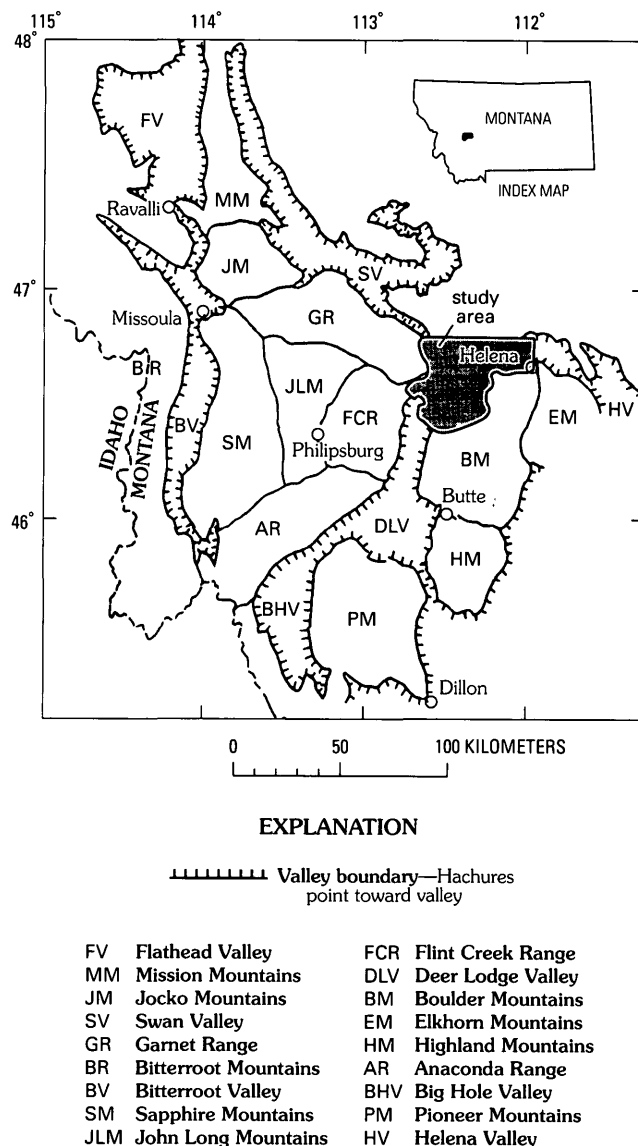


Figure 1. Principal geographic elements near the Elliston study area, Montana.

of Late Cretaceous and Tertiary age. The accompanying map (pl. 1) shows that the eastern edge of the Sapphire thrust plate overlaps the west limb of the Black Mountain syncline, and the north end of the Black Mountain syncline is cut by the Bald Butte fault, which is a prominent strike-slip fault of the Lewis and Clark line (Wallace and others, 1990). This report revises older stratigraphic usage applied to this area, it contrasts stratigraphic differences between the Sapphire thrust plate and the nonthrust terrane, and it presents evidence that supports slip on steep faults before 77 Ma. This report also shows that the Garnet line of Winston (1986) does not exist in the study area, and this report presents data that indicate that much extensional slip predates the Oligocene and younger Basin-and-Range extensional event described by Reynolds (1979).

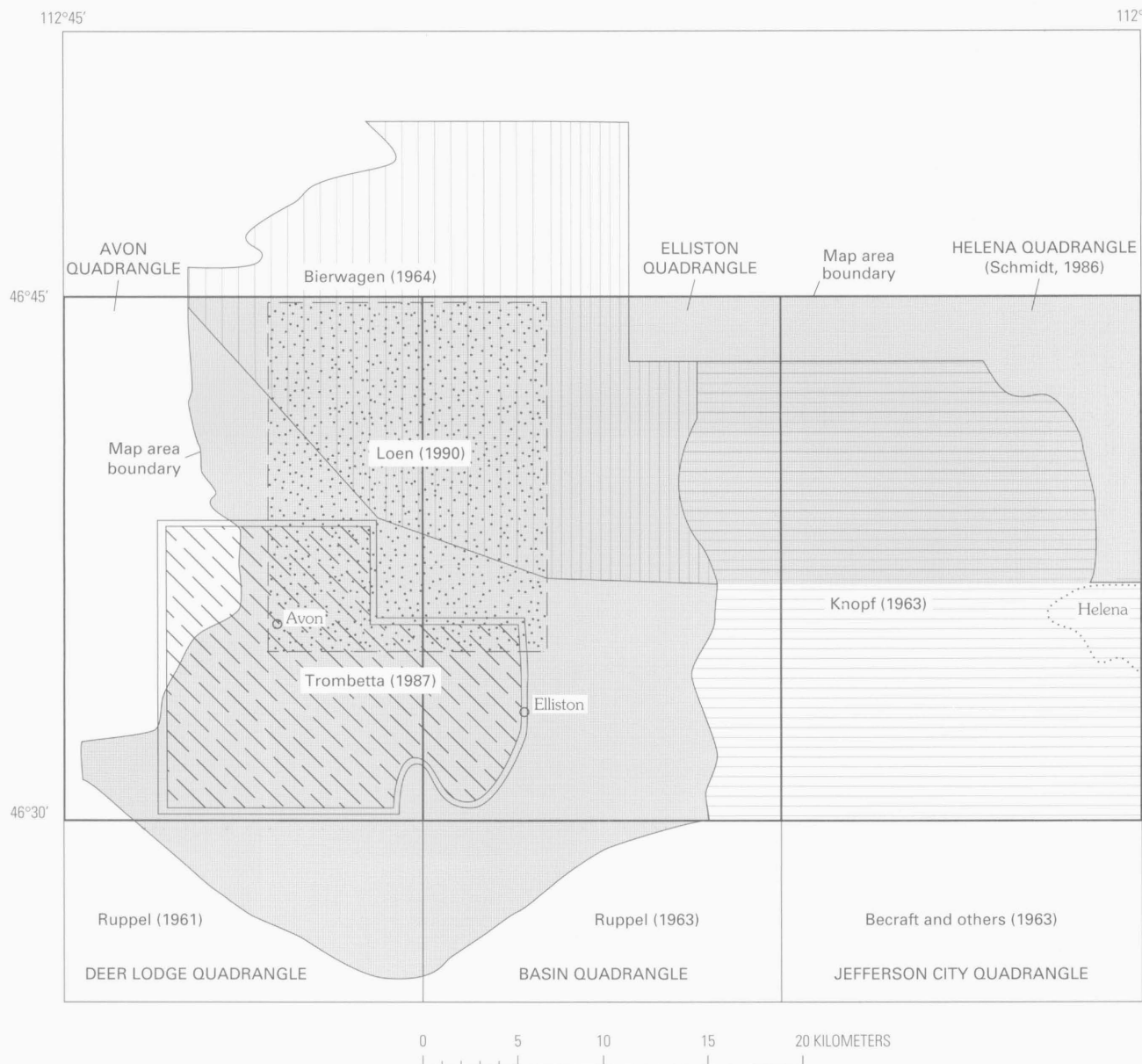


Figure 2. Index to geologic mapping in the Elliston region, Montana.

PREVIOUS STUDIES

The most detailed maps of this region were by Knopf (1963), who mapped the terrane southwest and west of Helena, Mont., at a scale of 1:48,000, and by Bierwagen (1964), who mapped the terrane north of Elliston, Mont., at a scale of 1:125,000 (fig. 2). These maps showed many of the principal faults and folds in their respective map areas and portrayed the distribution of Phanerozoic rock units, but the stratigraphic subdivisions applied to rocks of the Belt Supergroup were not sufficiently refined to portray Middle Proterozoic rock units nor to locate principal structures in these rocks. The southernmost part of the

study area of this report incorporates some geologic map information from Ruppel (1961, 1963) that mostly shows geologic relations of the plutonic and associated volcanic rocks. Schmidt (1986) published a generalized map of the Helena quadrangle (scale 1:62,500) that showed major structures and surficial geologic units, but that special-purpose map combined all sedimentary and plutonic bedrock, which eliminated structural and stratigraphic details of his mapping. Unpublished reconnaissance study of stream terraces and soils in surficial deposits northeast of Avon by M.R. Waters in 1982 was modified by later detailed mapping by Loen (1990). Trombetta (1987) produced a detailed geologic map of Tertiary volcanic and

sedimentary rocks in the Avon area, and some of his data are incorporated in this map.

PRESENT STUDY

The geologic map (pl. 1) of the Elliston area is based primarily on reconnaissance geologic mapping by R.G. Schmidt, and on detailed mapping by Loen (1990) of an area northeast of Avon. The Elliston study area is about 1,000 km² (square kilometers) in size, and it includes most of the Elliston quadrangle, the northern part of the Helena quadrangle, the northeastern corner of the Deer Lodge quadrangle, the northwestern corner of the Basin quadrangle, and the eastern part of the Avon quadrangle; all quadrangles are 15-minute size (fig. 2). R.G. Schmidt's reconnaissance geologic studies were done during the summers of 1975–76, 1978–79, and 1982. C.A. Wallace wrote the text, drafted the cross sections, and prepared the report for publication. H.H. Mehnert and R.G. Schmidt collected samples from several small stocks in the region for potassium-argon age analysis, and Mehnert analyzed and interpreted his isotopic data for this report. R.G. Schmidt succumbed to cancer in November 1983, but this report incorporates many concepts and hypotheses that he developed during the course of his geologic mapping.

STRATIGRAPHY

Lithofacies, thicknesses of rock units, and nomenclature of some Middle Proterozoic and Phanerozoic rock units reflect structural superposition of the Sapphire thrust plate over the parautochthon north, northeast, and east of the thrust plate (fig. 3). In the study area, formations in the frontal (eastern) part of the Sapphire thrust plate are most similar to the sequences from the western part of the Garnet Range described by Kauffman (1963) and to rocks described by Gwinn (1961) from the northwest part of the Flint Creek Range (fig. 1), whereas rocks in the parautochthonous terrane are most similar to the sequence described by Knopf (1963) southwest of Helena (fig. 2).

PROTEROZOIC ROCKS

BELT SUPERGROUP

The formations of the Belt Supergroup now recognized in the study area are, from oldest to youngest, the Spokane, Empire, Helena, Snowslip, Shepard, and Mount Shields Formations, the Bonner Quartzite, and the McNamara Formation. In the thrust slabs of the leading edge of the Sapphire thrust plate, only the uppermost part

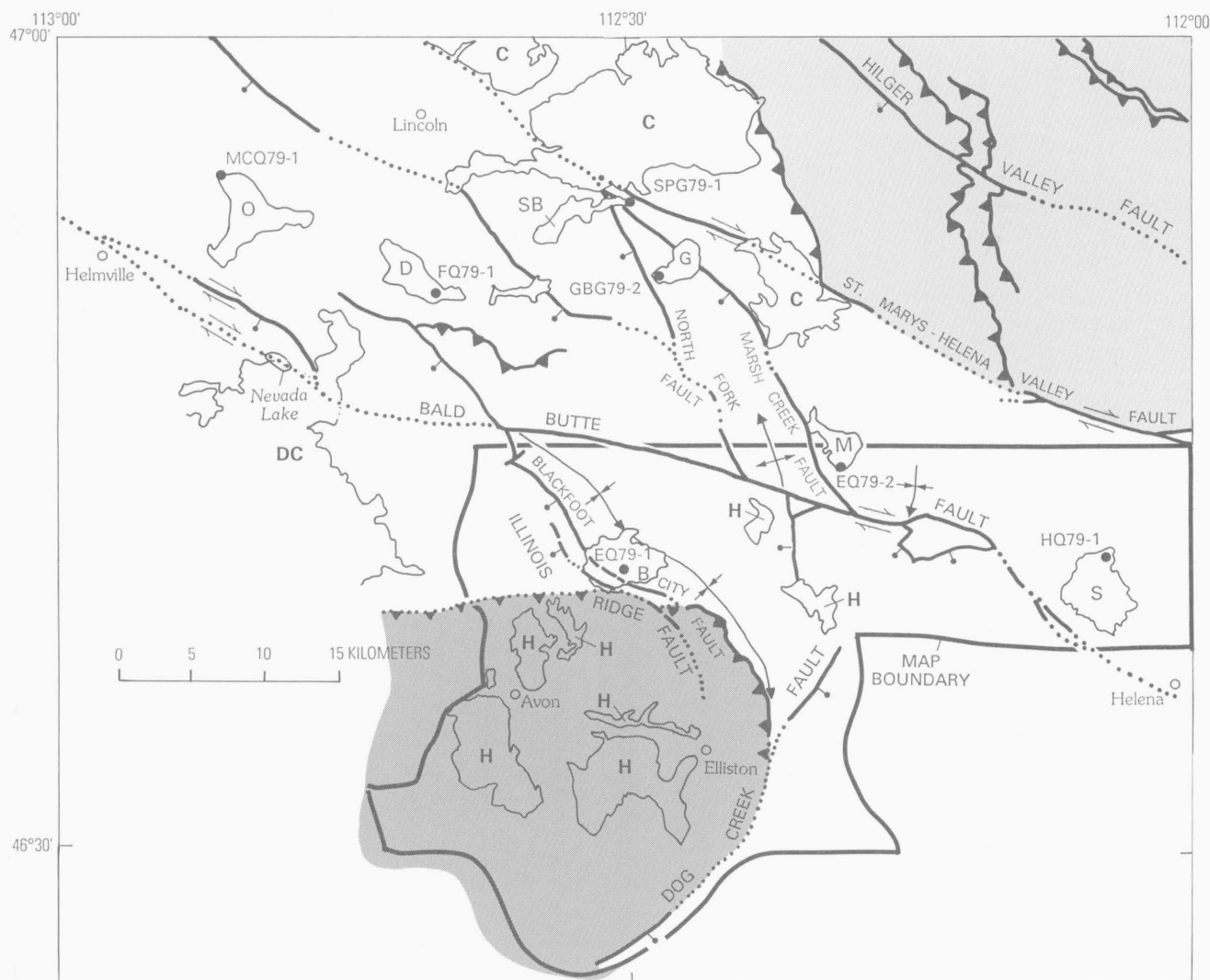
of this sequence, the Bonner Quartzite and the McNamara Formation, is exposed. The nomenclature, as applied in this report, reflects advances in the study of the regional stratigraphy of the Belt basin (Harrison, 1972; Harrison and others, 1986; Wallace and others, 1984; Wallace and others, 1986), and these names differ from those used by Knopf (1963) and Bierwagen (1964).

RAVALLI GROUP

Rocks of the Ravalli Group are represented by the Spokane (unit Ys, pl. 1) and Empire Formations (unit Ye, pl. 1). Typically, the Spokane Formation is a fine-grained unit composed principally of grayish-red, dark-grayish-red, and purplish-red, microlaminated argillite and siltite, and the fine-grained Empire Formation is a calcareous and dolomitic grayish-green, medium-green, and light-green argillite and siltite. Both formations contain thin, lenticular beds of fine-grained quartzite. Thin beds of siltite and fine-grained quartzite may contain 100 ppm (parts per million) or more of copper and silver (Harrison, 1972; Connor and McNeal, 1988). Regionally, the Spokane Formation ranges from about 1,400 to 2,400 m (meters) thick in the Montana disturbed belt north of the St. Marys-Helena Valley fault (fig. 3) (Whipple and others, 1987), but south of that fault the Spokane Formation is only 1,000 m thick. Similarly, the Empire Formation ranges between 300 and 450 m thick north of the St. Marys-Helena Valley fault (Whipple and others, 1987), but south of that fault this unit has a maximum thickness of about 300 m. The Spokane and Empire Formations occur only east, north, and northeast of the Sapphire thrust plate; nowhere do these units occur on the Sapphire thrust plate. Sediment of the Spokane Formation accumulated on a broad mudflat that was drained by broad, shallow channels that filled with silt and sand (Whipple, 1980; J.W. Whipple, U.S. Geological Survey, oral commun., 1990). This peritidal mudflat was periodically exposed above water level. The Empire Formation is principally a subaqueous and subtidal deposit (J.W. Whipple, U.S. Geological Survey, oral commun., 1990), and these fine-grained calcareous and dolomitic beds contain stromatolites at some places.

HELENA FORMATION

The Helena Formation (unit Yh, pl. 1) (name changed from Helena Dolomite of Knopf, 1963, by McGill and Sommers, 1967; Mudge and others, 1982; Harrison and others, 1986; Whipple and others, 1987) is composed mostly of cyclic interlayers of clastic, dolomite, and limestone beds (Eby, 1977). In a complete cycle the basal bed is clastic and is composed of fine-grained quartzite, siltite, or argillaceous siltite, the middle bed is composed of dark-gray or tan, laminated and mottled dolomite, and the top



EXPLANATION

TECTONIC TERRANES		STOCKS	
	Sapphire thrust plate	SB	Silver Belle—52 Ma (biotite), age may be reset
	Montana disturbed belt	B	Blackfoot City—77 Ma (biotite)
	Parautochthonous terrace	M	Marysville—79 Ma (biotite)
VOLCANIC FIELDS		D	Dalton Mountain—79 Ma (biotite)
	Helena volcanic field—39 Ma	O	Ogden Mountain—85 Ma (hornblende), 81 Ma (biotite)
	Douglas Creek volcanic field—44-47 Ma	S	Scratchgravel Hills—97 Ma (hornblende)
	Crater Mountain volcanic field—Oldest rocks are 56 Ma	G	Granite Butte—78 Ma (biotite), 109 Ma (hornblende)
	Contact	● EQ79-1	Locality of sample taken for potassium-argon age determination—Showing number
	Fault—Dashed where approximately located; dotted where concealed; bar and ball on downthrown side; barbs show direction of movement		Anticline—Showing direction of plunge
	Thrust fault—Dotted where concealed; saw teeth on upper plate		Syncline—Showing direction of plunge

Figure 3. Principal structures in the Elliston region and isotopic-age sample localities.

bed is composed of dark-gray, argillaceous or silty, micritic calcite, or calcareous argillite that contains characteristic molar-tooth structure, which is formed from vertical ribbons and veinlets of calcite. Stromatolites commonly occur in the upper calcareous beds. A lithofacies of dark-gray limestone and blackish-gray calcareous argillite occurs in the upper part of the Helena Formation in the hills west of Helena, and this limestone is about 300 m thick. The cyclic, dolomite-rich lithofacies occurs east, north, and northeast of the Sapphire thrust plate, and in these areas the formation is about 1,200 m thick. The dark-gray limestone lithofacies forms most of the Helena Formation on the Sapphire thrust plate west and southwest of the Elliston study area, where the formation is estimated to be at least 2,500 m thick (Ruppel and others, 1981). The Helena Formation is thinner (775–1,000 m) northeast of the St. Marys-Helena Valley fault (fig. 3) in the Montana disturbed belt (Whipple and others, 1987) than in the Elliston study area (1,200 m) and to the southwest and west on the Sapphire thrust plate (about 3,000 m) (Ruppel and others, 1981). The cyclic rocks of the Helena Formation represent deposits on a carbonate platform that formed a rim along the east side of the Belt basin (Eby, 1977), and the dark-gray limestone lithofacies in the upper part of the Helena Formation represents a platform-slope deposit that accumulated in a slightly deeper part of the basin west of the carbonate platform (Wallace, Harrison, and others, 1989).

MISSOULA GROUP

The nomenclature applied to rocks above the Helena Formation conforms to the regional scheme proposed by Harrison (1972); the Missoula Group in the study area is composed of the Snowslip (unit Ysn, pl. 1), Shepard (unit Ysh, pl. 1), and Mount Shields Formations (unit Yms, pl. 1), the Bonner Quartzite (unit Ybo, pl. 1), and the McNamara Formation (unit Ym, pl. 1), from base to top (Wallace and others, 1984). Formations now known to be Snowslip, Shepard, and Mount Shields were previously defined as the Marsh Formation (Walcott, 1899; Knopf, 1950), a term later abandoned by Harrison (1972, p. 1219). Knopf (1963) correctly recognized that rocks from the type locality of the "Marsh shales" of Walcott (1899) in Marsh Creek occurred under Knopf's "Helena Dolomite" and were part of the Spokane Formation. However, Knopf (1963) still retained the name "Marsh Formation" for rocks above his Helena Dolomite and below his Greenhorn Mountain Quartzite, which formed the top of the sequence (Knopf, 1963, p. 2), even though Walcott's original type locality applied "Marsh" to rocks now known to be part of the Ravalli Group.

The Snowslip Formation, at the base of the Missoula Group, is a grayish-red and purplish-red, laminated and microlaminated, silty argillite, argillaceous siltite, and

fine-grained quartzite that contains some beds of grayish-green argillite and siltite. In the study area the Snowslip Formation contains more zones of grayish-green beds than to the west and southwest on the Sapphire thrust plate, and the lower part of the formation contains more beds of tan-weathering, coarse-grained quartzite than on the Sapphire thrust plate (Schmidt and others, 1983). In the Black Mountain syncline and northwest of Helena, the Snowslip is about 400 m thick, as compared to a thickness of about 1,070 m to the west and southwest of the study area on the Sapphire thrust plate, and as compared to a thickness of about 300 m northeast of the St. Marys-Helena Valley fault (fig. 3) (Whipple and others, 1987). Grayish-red argillite and siltite beds of the Snowslip Formation were deposited on mudflats that were drained by shallow channels that are now represented by fine-grained quartzite. Grayish-green beds were deposited in a dominantly shallow-water environment (Wallace and others, 1984).

The Shepard Formation contains a basal stromatolite zone in the study area, and that zone is overlain by the dolomitic and calcareous argillite and siltite that characterize the unit. In the study area the Shepard is more calcareous and dolomitic than to the west and southwest on the Sapphire thrust plate. This formation is about 335 m thick in the study area, and the unit thins to about 150 m to the northeast of the St. Marys-Helena Valley fault (fig. 3) (Whipple and others, 1987). In the northern part of the Sapphire thrust plate the Shepard is about 760 m thick. The Shepard Formation was deposited in a shallow-water environment and mostly was above wave base (Wallace and others, 1984).

The Mount Shields Formation consists locally of the lower three of the six members that are recognized regionally (Whipple, 1992; Harrison and others, 1992). In the study area, the lower of the three, referred to as member 1 (unit Yms1, pl. 1), is about 400 m thick and consists mostly of zones of interbedded grayish-red argillite and tan siltite and zones of tan fine-grained feldspathic and nonfeldspathic quartzite; on the Sapphire thrust plate to the southwest and west, the thickness of member 1 is about 1,000 m. Member 2 (unit Yms2, pl. 1), in the study area, is about 300 m thick and is composed mainly of fine-grained interbedded feldspathic and nonfeldspathic quartzite. This member thickens and coarsens to the west on the Sapphire thrust plate, where it is represented by about 4,000 m of fine- to coarse-grained, conglomerate-bearing feldspathic quartzite. Member 3 (unit Yms3, pl. 1), the uppermost unit, is dominantly grayish-red microlaminated argillite and reddish-gray or tan-weathering siltite, and fine-grained quartzite; the member is about 400 m thick in the study area, but it is about 1,070 m thick on the Sapphire thrust plate to the southwest and west. The total thickness of the three members is about 1,100 m in the study area, which contrasts with a total thickness of about 700 m of the Mount Shields Formation northeast of the St.

Marys-Helena Valley fault (fig. 3) (Whipple and others, 1987). The argillite-rich members of the Mount Shields Formation, members 1 and 3, were deposited in shallow-water mudflats that were periodically covered by prodeltaic sheet sands. Member 2 represents deposition of sand in delta-channel, delta-plain, strand-plain, and delta-front environments (Wallace and others, 1984).

The Bonner Quartzite, which overlies the Mount Shields Formation, consists of about 460 m of medium- and coarse-grained pebbly and conglomeratic feldspathic and lithic quartzite that is prominently crossbedded and channeled. Originally, Knopf (1963) named this unit the Greenhorn Mountain Quartzite, but this name was abandoned in favor of the Bonner Quartzite, which is recognized over most of the Belt basin (Harrison, 1972; Harrison and others, 1986; Wallace and others, 1984; Winston and Wallace, 1983). The contact shown by Knopf (1963) as the base of the Greenhorn Mountain Quartzite occurs in member 3 of the Mount Shields Formation, and the disparity in location of the lower contact of the Bonner Quartzite between Knopf's map and this map (pl. 1) accounts for the difference in thickness of 550 m of Knopf (1963, p. 3) and the thickness of about 460 m for the Bonner given here. The Bonner Quartzite is about the same composition and thickness on the Sapphire thrust plate but thins to about 250 m northeast of the St. Marys-Helena Valley fault (fig. 3) (Whipple and others, 1987). The Bonner Quartzite was deposited in a delta-plain and delta-channel environment (Wallace and others, 1984).

Overlying the Bonner Quartzite in Knopf's (1963) study area is the Middle Cambrian Flathead Quartzite, but north of the Elliston area, the Middle Proterozoic McNamara Formation (unit Ym, pl. 1) overlies the Bonner Quartzite in normal succession. The McNamara Formation is mainly a variegated red and green, microlaminated argillite and siltite, and fine-grained quartzite that contains beds and clasts of red and green chert. From the cross sections the McNamara Formation is estimated to be a minimum of about 760 m thick on the Sapphire thrust plate and is absent in the eastern part of the study area because erosion truncated the McNamara before deposition of the overlying Flathead Quartzite (Middle Cambrian), forming a small angular discordance. The McNamara Formation was deposited mostly in shallow water, and sheets of prodeltaic and channel sands periodically covered the argillaceous beds (Wallace and others, 1984). The Garnet Range and Pilcher Formations, which normally occur at the top of the Missoula Group about 80 km west of the study area in the thrust-faulted terrane, were eroded before the Flathead Quartzite was deposited.

PALEOZOIC ROCKS

Paleozoic rocks, composed mainly of limestone and dolomite in the lower part and sandstone, shale, and

limestone in the upper part, are exposed in the Black Mountain syncline and in thrust slabs that truncate the west flank of the syncline. Paleozoic rocks above the basal thrust faults of the Sapphire thrust plate represent lithofacies that are transitional between (1) the eastern lithofacies described by Knopf (1963) from the Helena region and from the Black Mountain syncline, and (2) the western lithofacies described by Emmons and Calkins (1913) that is restricted to the Sapphire thrust plate. These rocks range in age from Middle Cambrian to Permian. They were deposited on a broad miogeoclinal shelf, where most of the deposits accumulated in near-shore, littoral, and open-water marine environments.

CAMBRIAN ROCKS

The Middle Cambrian assemblage of rocks in the study area forms part of the classic miogeoclinal, eastward time-transgressive sequence that typifies early Paleozoic sedimentation in the Rocky Mountains. Depositional environments progressed from strandline clastic deposits at the edge of the Paleozoic sea at the base of the sequence, to intertidal mudflats at the middle of the sequence, to carbonate banks and shallow-water carbonate deposits at the top of the sequence. These Middle Cambrian rock units are central to a dispute about rock-unit correlations, and although this dispute affects only local correlations, it obscures differences in stratigraphy that define the front of the Sapphire thrust plate.

The Flathead Quartzite (unit Cf, pl. 1) (Middle Cambrian) unconformably overlies Middle Proterozoic rocks, and it is mostly a fine-grained, vitreous quartzite that contains some lenticular beds of granules and small pebbles. Glauconite grains occur in some beds. The Flathead Quartzite represents a strandline and shallow-marine deposit that migrated eastward through time (Hanson, 1952; Kauffman, 1963), and this unit rests on an erosion surface that progressively truncated Belt rocks from west to east. At the east side of Avon Valley the quartzite unit overlies the McNamara Formation, but 8 km northwest of Helena, the Flathead was deposited on the upper Snowslip Formation. Before the Flathead was deposited, a minimum of 2,500 m of Middle Proterozoic rocks was eroded over a lateral distance of about 30 km, which produced a regional angular discordance of about 5°.

The Silver Hill Formation (unit Csh, pl. 1) (Middle Cambrian) was first described in the southern Flint Creek and northern Anaconda Ranges by Emmons and Calkins (1913). Kauffman (1963) and Maxwell (1965) extended the identification of this unit northward into the Garnet Range and John Long Mountains. The formation thickens from south to north; in the north-central Anaconda Range this unit is about 60 m thick (Lidke and Wallace, 1993) and in the northern Anaconda Range and southwestern Flint Creek Range the Silver Hill is about 100 m thick

(Emmons and Calkins, 1913; McGill, 1959). In the eastern part of the Garnet Range this unit is about 335 m thick (A.B. French, Mobil Oil Co., written commun., 1978), and in the study area the Silver Hill is about 275 m thick. The principal lithologic characteristics of the Silver Hill Formation are the grayish-green, waxy shale and interbedded gray limestone at the top; a thick zone of gray limestone and irregularly interbedded, laterally discontinuous, thin, orange-weathering siliceous and argillaceous dolomite in the middle; and greenish-gray, grayish-black, and blackish-gray shale, siltstone, and fine-grained sandstone that contains rare beds of dark-gray limestone at the base. In the study area, the 10–30 m thickness of the upper shale exceeds the normal thickness of 2–3 m over most of the Sapphire thrust plate. The medial limestone, which closely resembles the “blue-and-gold” limestone of the Meagher Limestone east of the study area described by Klepper and others (1957, p. 9), is thicker in the study area (240 m) than in the type section at Silver Hill (about 57 m) (Emmons and Calkins, 1913). The lower shale zone, although not present everywhere west of the study area in the Garnet Range, is of similar thickness to the type section (Emmons and Calkins, 1913), and the lithology of this lower zone is similar to that of the type section. The upper and lower shales were deposited in a shallow-water, intertidal-mudflat environment (Kauffman, 1965). The medial limestone accumulated as a calcareous ooze in clear seas, and fine-grained terrigenous material was sporadically transported into the depositional basin to form the terrigenous, orange-weathering interbeds (Kauffman, 1965).

The Wolsey Shale (unit Cw , pl. 1) (Middle Cambrian) is lithologically similar to the lower shale member of the Silver Hill Formation, although the Wolsey contains more beds of limestone distributed through the formation (Klepper and others, 1957) than does the lower shale of the Silver Hill Formation. The Wolsey varies in thickness from about 90 m near Helena (Knopf, 1963) to about 120 m in a fault block 1 km south of Bald Butte. The Wolsey Shale was probably deposited in an intertidal-mudflat environment.

The Meagher Limestone (unit Cm , pl. 1) (Middle Cambrian) is similar in lithic type to the middle limestone described from the Silver Hill Formation. In the study area the Meagher is about 185 m thick, but Knopf (1963, p. 3) reported that this limestone was about 195 m thick. Although Knopf (1963), Klepper and others (1957), and Hanson (1952) did not specify a depositional environment for the Meagher Limestone, the common occurrence of oolites and pisolites in the limestone suggests deposition in a carbonate-bank environment that was locally above wave base and in the swash zone.

The Park Shale (unit Cp , pl. 1) (Middle Cambrian) is a grayish-green waxy shale and interbedded siltstone that is similar in lithic character to the upper shale of the

Silver Hill Formation. The thickness of the Park Shale in the study area (about 80 m) is slightly more than the 61 m reported by Knopf (1963, p. 3) east of the study area. The depositional environment was probably similar to the intertidal-mudflat environment of the upper shale member of the Silver Hill Formation.

The Hasmark Formation (unit Ch , pl. 1) (Upper Cambrian) is a light- and medium-gray, thinly laminated, conchoidal-fracturing dolomite. This unit is about 185 m thick in the study area, but it thins toward the east where Knopf (1963, p. 4) reported a thickness of about 140 m for the Hasmark. To the west in the western Garnet Range, Kauffman (1965, p. 81) indicated that the Hasmark was about 550 m thick, but southwest of the study area at Hasmark near Philipsburg, Emmons and Calkins (1913) reported a thickness of about 280 m. The depositional environment of the Hasmark is uncertain because diagnostic fossils are generally lacking, but the occurrence of oolites near the base of the Hasmark Formation in the Philipsburg region suggests that, at least locally, the lower part of the formation was deposited above wave base in the swash zone.

The regional correlation of the Silver Hill Formation, Wolsey Shale, Meagher Limestone, Park Shale, and Hasmark Formation is subject to dispute (Emmons and Calkins, 1913; Hanson, 1952; Kauffman, 1965; Ruppel and others, 1981; and Zen, 1988), and different interpretations about lithologic and biostratigraphic equivalence obscure important stratigraphic contrasts between the Sapphire thrust plate and the parautochthonous terrane to the east and northeast. Two conflicting regional correlations of Cambrian formations evolved since Emmons and Calkins (1913) first described a new sequence of Cambrian rocks from the southern Flint Creek and northern Anaconda Ranges, and these different schemes are based on different presumptions (fig. 4). Emmons and Calkins (1913) and Ruppel and others (1981) proposed rock-unit correlations that equated the Silver Hill Formation with the Wolsey Shale, Meagher Limestone, and Park Shale and established the Hasmark Formation as a lateral equivalent of the Pilgrim Limestone (fig. 4B). Hanson (1952), Kauffman (1965), and Zen (1988) combined time-stratigraphic and rock-stratigraphic correlations that equated the Silver Hill Formation with the Wolsey Formation, and correlated the Meagher Limestone, Park Shale, and Pilgrim Limestone with the Hasmark Formation (fig. 4A).

The rock-stratigraphic correlation proposed by Emmons and Calkins (1913) was based on similarity of composition and sequence of rock units in the lower Paleozoic assemblage; the composition, texture, and bedding characteristics of the lower member of the Silver Hill Formation closely resemble the lithology of the Wolsey Shale, the middle limestone member is similar to the Meagher Limestone, and the thin upper member is lithologically similar to the Park Shale. The Hasmark

Formation, which overlies the Silver Hill Formation, is lithologically similar to the Pilgrim Formation of the Elkhorn Mountains (Klepper and others, 1957; Klepper and others, 1971). Emmons and Calkins (1913, p. 64) recognized that some Cambrian formations crossed time lines, but they lacked detailed biostratigraphic control to demonstrate that all units were diachronous.

Evolution of the combined time-stratigraphic and rock-stratigraphic correlation clearly showed the diachronous nature of Middle and Late Cambrian rock units (Hanson, 1952; Kauffman, 1965), but this scheme presumed a rock-unit correlation between a locally occurring shale in the Hasmark Formation and the Park Shale. Biostratigraphic data and interpretations of Hanson (1952) showed that the Silver Hill Formation was diachronous with respect to the Wolsey Shale, Meagher Limestone, and Park Shale, and that the Middle Cambrian rock units were time-transgressive to the east. Hanson, however, correlated a shale that occurs locally near the middle of the Hasmark Formation (Emmons and Calkins, 1913) with the Park Shale, a correlation that had been evaluated and rejected by Emmons and Calkins (1913, p. 63–64). Moreover, Hanson's correlation of a shale in the Hasmark Formation with the Park Shale had no biostratigraphic support. As a result, Hanson equated the lower part of the Hasmark Formation with the Meagher Limestone (Hanson, 1952). Kauffman's (1965) biostratigraphic data from the Garnet Range confirmed the diachronous relations among the Middle Cambrian units in his study area, and he followed Hanson's correlation that equated the lower Hasmark Formation and Meagher Limestone. However, Kauffman (1963) provided no new biostratigraphic data to support correlation of these different lithologic units. Zen (1988) established similar diachronous biostratigraphic relations for Middle and Late Cambrian rocks in the Pioneer Mountains, but Zen presented no new biostratigraphic data to support Hanson's (1952) correlation between the Park Shale and a medial shale in the Hasmark Formation, which, in the Pioneer Mountains, is a silty dolomite rather than a shale. Zen (1988, p. 9) stated that the Silver Hill Formation is homotaxic with respect to the Wolsey Shale, Meagher Limestone, and Park Shale because the *Glossopleura* zone occurs in the Wolsey Shale in eastern exposures (Hanson, 1952), and the *Glossopleura* zone occurs near the top of the medial limestone of the Silver Hill Formation in western exposures (Wells, 1974); these same relations, however, may have resulted from a simple diachronous relation among rock units, so Zen's argument is not compelling.

The regional correlation adopted here is the scheme first proposed by Emmons and Calkins (1913) because (1) formations are correlated on lithologic similarity and similarity of rock sequences in accordance with the North American Stratigraphic Code (1983, p. 851) and not on biostratigraphic data as promulgated by Hanson (1952),

Kauffman (1963), and Zen (1988); (2) the medial shale of the Hasmark Formation is a local unit that is not lithologically similar to the Park Shale (Klepper and others, 1957), and correlation of the medial shale in the Philipsburg area to a thin silty dolomite in the Pioneer Mountains (Zen, 1988) is of dubious validity; (3) the Hasmark Formation is lithologically distinct from the Meagher Limestone, and none of the biostratigraphic data of Hanson (1952), Kauffman (1963), or Zen (1988) suggest even a partial time equivalence of the two formations; and (4) biostratigraphic data of Hanson (1952), Kauffman (1963), and Zen (1988) are compatible with the original lithologic correlation proposed by Emmons and Calkins (1913), and those data indicate that the Middle and Late Cambrian rocks represent a transgressive sequence that becomes younger toward the east. Zen's suggestion that the Silver Hill Formation is homotaxic is not proved, and the only proof that would support his contention would be the occurrence of the *Bathyriscus* zone in the lower part of the Hasmark Formation. Available biostratigraphic data provide no new information about the age of the Hasmark Formation, so no reason exists to discard the original assignment made by Emmons and Calkins (1913).

CAMBRIAN AND DEVONIAN ROCKS

The combined Maywood(?) (Upper Devonian) and Red Lion (Upper Cambrian) Formations (unit D ϵ mr, pl. 1) overlie the Hasmark Formation. As mapped here, the unit consists mostly of strata of the Red Lion Formation, but the uppermost beds of the unit may include a lateral equivalent of the Upper Devonian Maywood Formation. In the study area the lower part of the Red Lion consists of 10–15 m of grayish-red unevenly bedded shale, siltite, limestone, and dolomite, and an upper part that is 85–90 m of limestone interbedded with thin planar layers of siliceous and argillaceous dolomite, which is the same sequence originally described by Emmons and Calkins (1913) from the Philipsburg area. Kauffman (1965) separated the lower 10 m of shale and siltstone from overlying limestone and called the lower clastic beds the Dry Creek Shale Member of the Snowy Range Formation. Loen (1990), following Schmidt's original unpublished reconnaissance mapping, included lower clastic beds that could be the Dry Creek Shale in the Red Lion Formation. Knopf (1963, p. 4) suggested that the uppermost 10 m of thinly bedded limestone and grayish-red siltstone included with the Red Lion Formation in his report could represent the Maywood Formation (Upper Devonian), which overlies the Red Lion Formation to the west and southwest on the Sapphire thrust plate. Nelson (1963, p. 36) reported Late Devonian fossils from the uppermost limestone beds above rocks he designated Red Lion and below rocks he assigned to the Jefferson Dolomite 80 km to the west in

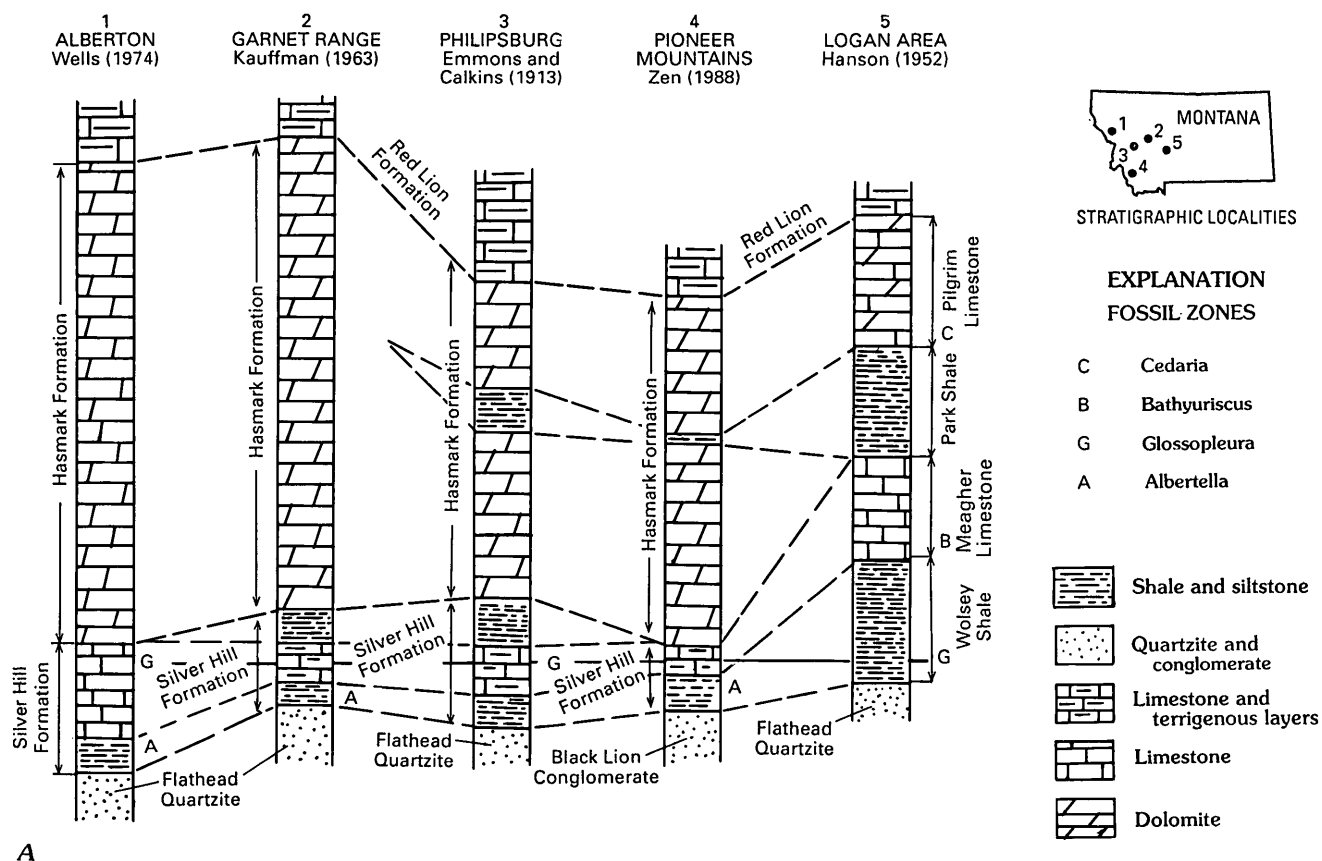


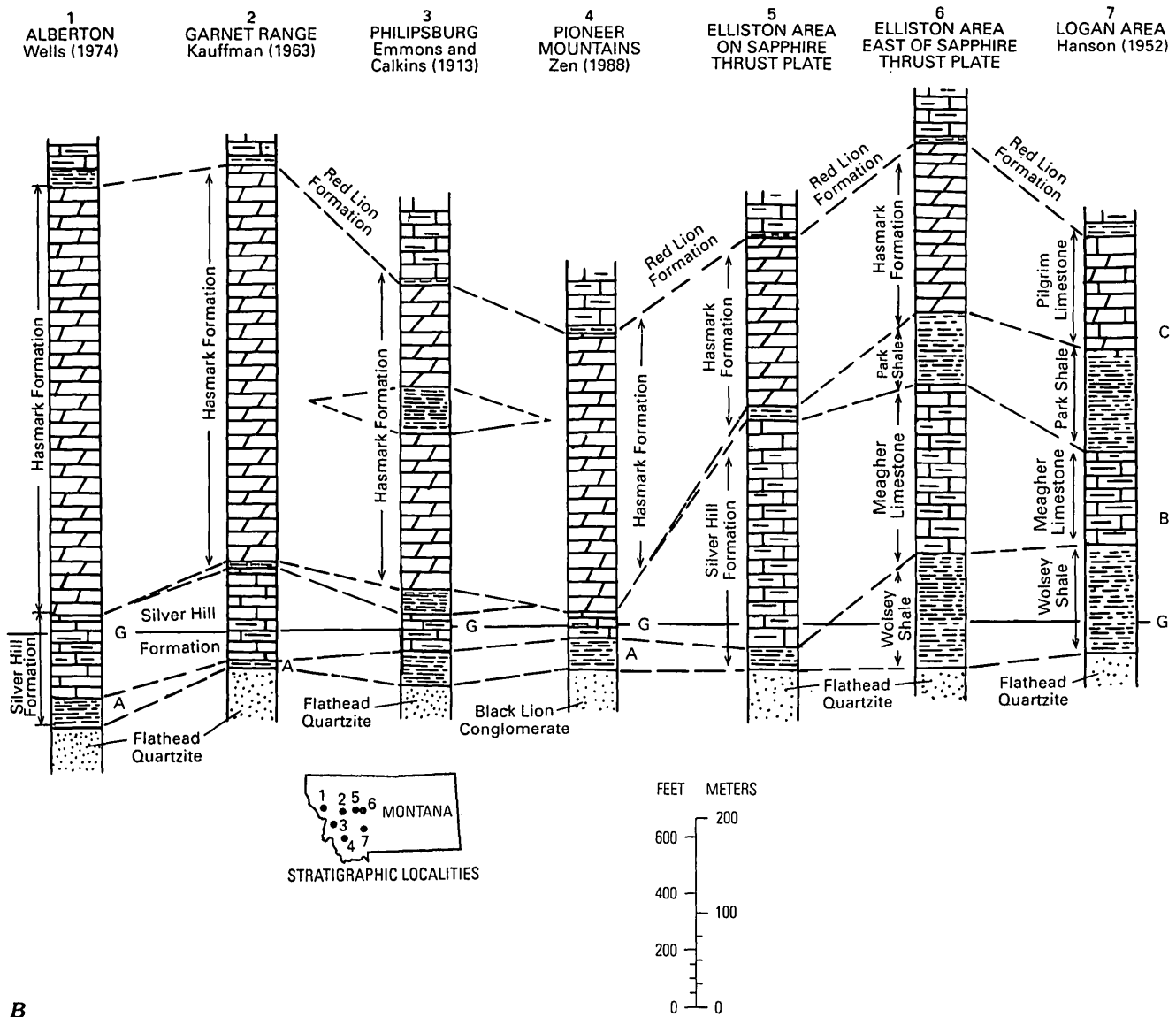
Figure 4 (above and facing page). Correlations of lower Cambrian units in the Elliston area. **A**, Combined biostratigraphic and lithostratigraphic correlation proposed by Zen (1988). Modified from Zen (1988). **B**, Lithostratigraphic correlation proposed in this report. Formation and member thicknesses are different from those shown by Zen (1988) because Zen's thicknesses are not the same as those given by Wells (1974), Kauffman (1963, p. 6), and Emmons and Calkins (1913, p. 53–61).

the Duck Creek Pass quadrangle, and the limestone beds of Late Devonian age may represent the Maywood Formation. A Late Devonian age is generally accepted for the Maywood Formation in central and northwestern Montana (Sloss and Laird, 1947). Although poor exposures precluded fossil collections and detailed lithologic study of calcareous beds below the Jefferson Formation in the study area, we include beds that could be the Maywood Formation with the Red Lion, as did Klepper and others (1957), Knopf (1963), Nelson (1963), and Klepper and others (1971). The combined Red Lion and Maywood(?) rock unit contains the regional disconformity that represents nondeposition or erosion of Ordovician and Silurian deposits. Beds of the lower shale and siltstone unit were probably deposited in a shallow-marine mudflat environment. The limestone and terrigenous dolomite of the Red Lion was probably deposited in a shallow-marine environment into which storms circulated fine-grained terrigenous debris (Lochman, 1957, p. 154). Edgewise conglomerate and fossil hash described by Kauffman (1963, p. 7) supports the occurrence of strong currents above wave base. West and southwest of the study area, the Red

Lion and Maywood Formations thicken to about 100 m each, and these formations can be mapped separately at a scale of 1:62,500 (C.A. Wallace, unpub. map data, 1978).

DEVONIAN ROCKS

The Jefferson Formation (unit Dj, pl. 1) (Upper Devonian) overlies rocks of the combined Red Lion and Maywood(?) Formations, and the Jefferson consists mostly of fine- to coarse-grained dark-gray and grayish-black, fetid dolomite and limestone as described by Knopf (1963, p. 4). The three informal members mapped by Kauffman (1963) in the western Garnet Range were not separated in the study area; the Jefferson Formation is only about 230 m thick in the region of the Black Mountain syncline, but the formation thickens to about 520 m on the Sapphire thrust plate to the west in the western Garnet Range (Kauffman, 1963). The Jefferson Formation was deposited in a shallow-water marine environment that was far from a source of terrigenous debris but subject to strong current action (Robinson, 1963, p. 28).



DEVONIAN AND MISSISSIPPIAN ROCKS

About 80 m of the Three Forks Formation (unit MDt, pl. 1) (Lower Mississippian and Upper Devonian) overlies the Jefferson Formation in the Black Mountain syncline, which is structurally below thrust faults of the Sapphire thrust plate, but the yellowish-gray and brownish-gray weathering shale, siltstone, and limestone of the Three Forks is absent, or very thin, 70 km west of the study area in the Garnet Range on the Sapphire thrust plate. The Three Forks Formation is rarely well exposed, and only about 25 m of Three Forks Formation might be present above frontal thrusts of the Sapphire thrust plate along the southeast side of the Black Mountain syncline. On these thrust slices a swale occurs at the base of a prominent limestone unit of the Madison Group, but exposures are poor and we are not certain that the swale represents shale

of the Three Forks Formation. A thin or absent Three Forks Formation is a typical feature of the Paleozoic sequence on the Sapphire thrust plate (Ruppel and others, 1981). Analysis of brachiopods collected from the Three Forks by J.T. Dutro, Jr. (U.S. Geological Survey, written commun., 1980) suggests that part of the formation represents the Trident Member near Logan, Mont. According to Robinson (1963, p. 38), the Three Forks Formation was deposited in a shallow-water, near-shore, marine environment that was sometimes restricted from free circulation of water with the open ocean.

MISSISSIPPIAN ROCKS

The Madison Group (unit Mm, pl. 1) (Upper and Lower Mississippian) consists mostly of a lower cherty,

thin-bedded limestone and an upper thick-bedded, coarse-grained gray limestone. In the Helena area, Knopf (1963, p. 5) assigned the lower 100 m of the Madison Group to the Lodgepole Limestone and the upper 330 m to the Mission Canyon Limestone. In the Elliston area, the Lodgepole Limestone may be absent or very thin. The depositional environment of the Madison Group was a shallow-water, open-marine setting into which fine-grained clastic material was transported.

MISSISSIPPIAN AND PENNSYLVANIAN ROCKS

Rocks formerly assigned to the Amsden Formation (Upper Mississippian) by Knopf (1963) and by Bierwagen (1964) are now assigned to the Snowcrest Range Group (unit IPms, pl. 1) (Lower Pennsylvanian and Upper Mississippian) by B.R. Wardlaw (U.S. Geological Survey, written commun., 1987). The yellowish-gray and grayish-red calcareous shale, siltstone, fine-grained sandstone, and lenticular gray limestone and dolomite are rock types similar to parts of the Amsden Formation, but the Snowcrest Range Group is mostly equivalent to the Big Snowy Group that underlies the Amsden Formation (Wardlaw and Pecora, 1985). The Snowcrest Range Group could be as young as Early Pennsylvanian, according to Wardlaw. In the Garnet Range in the northern part of the Sapphire thrust plate and in the Elkhorn Mountains, the Snowcrest Range Group is reported to be about 100 m thick, but Knopf (1963, p. 5) reported about 113 m for the thickness of the Quadrant Quartzite, in which he included about 40 m of rocks now assigned to the Snowcrest Range Group. The Snowcrest Range Group was probably deposited in marine, supratidal, strandline, and subtidal environments (B.R. Wardlaw, U.S. Geological Survey, oral commun., 1991).

PENNSYLVANIAN AND PERMIAN ROCKS

The Quadrant Quartzite (unit IPq, pl. 1) (Pennsylvanian), which overlies the Snowcrest Range Group, was combined with the Shedhorn Sandstone and the Phosphoria and Park City Formations (Permian) because only the Quadrant was well exposed, and the Phosphoria Formation and related units were too thin to separate at this map scale. The Quadrant Quartzite is about 100 m thick in most of the study area, but Knopf (1963, p. 5) suggested that the Quadrant was about 75 m thick near Helena, Mont. According to Robinson (1963, p. 51), the Quadrant Quartzite is a shallow-water and strand-plain deposit. The Phosphoria Formation is poorly exposed in the study area, so thickness determinations were not made; Knopf (1963, p. 5) estimated that about 6 m of the Phosphoria

Formation occurred near Helena. Rocks of the Phosphoria Formation are of shallow-water marine origin (Robinson, 1963, p. 54).

MESOZOIC ROCKS

Mesozoic rocks are composed mostly of Jurassic and Cretaceous clastic rocks and rare carbonate beds that are exposed in the Black Mountain syncline and in an imbricate thrust terrane south and southwest of Elliston. On the Sapphire thrust plate, lithofacies of Jurassic and Lower Cretaceous rocks are different from lithofacies of correlative rocks northeast and east of the thrust plate. Jurassic rocks are thicker and contain more coarse-grained clastic beds on the Sapphire thrust plate than correlative rocks east of the thrust plate, and Lower Cretaceous rocks contain thicker zones of limestone and thicker beds of sandstone on the thrust plate than east of the thrust plate. In general, Jurassic sediment below the Morrison Formation accumulated in marine and near-shore depositional environments, and sediment deposited after the Morrison Formation accumulated in continental, lacustrine, brackish-marine, and fluvial environments. Continental deposits accumulated during Late Cretaceous time while eruption of the Elkhorn Mountains Volcanics and related extrusive rocks heralded emplacement of the Boulder batholith at depth (Robinson and others, 1968; Gwinn and Mutch, 1965).

JURASSIC ROCKS

Formations of the Ellis Group (Upper and Middle Jurassic) are the oldest Mesozoic rocks in the study area, and although Triassic rocks are absent in the study area, an angular discordance between Permian and Jurassic rocks cannot be demonstrated. Mesozoic rocks are represented by formations of the Jurassic Ellis Group (unit J_e, pl. 1), which is composed, in ascending order, of the Sawtooth (unit J_{sa}, pl. 1), Rierdon (unit J_r, pl. 1), and Swift (unit J_s, pl. 1) Formations. Overlying the Swift Formation at most places is the Jurassic Morrison Formation (unit J_m, pl. 1). The Morrison Formation (J_m) and Ellis Group (J_e) were combined (unit J_{me}, pl. 1) because the formations were too thin (total of about 45 m thick) to represent at this scale in most of the study area, but near Elliston in the thrust-faulted terrane, the Sawtooth, Rierdon, and Swift Formations of the Ellis Group and the Morrison Formation thicken to an aggregate of about 185 m, and individual formations were separated at some places.

The Sawtooth Formation is principally fine grained sandstone and siltstone that was deposited under shallow-marine conditions (Imlay, 1957), and the Rierdon Forma-

tion is black shale, muddy siltstone, and limestone also deposited under marine conditions (Imlay, 1957). Both formations form swales and are poorly exposed, so thickness estimates presented here may be exaggerated.

The Swift Formation is fossiliferous, calcareous, yellowish-weathering, lithic sandstone and sandy conglomerate that contains abundant crossbedding and channels. Interbeds of dark-gray shale and limestone occur between beds of sandstone and conglomerate, but at some places conglomeratic sandstone of the Swift forms massive cliffs and prominent ridges. Black chert and phosphate grains and pebbles are common in rocks of the Swift Formation, and these beds were probably deposited in a shallow-marine environment (Imlay, 1957).

The Morrison Formation is principally a continental deposit that was formed on flood plains and in stream channels; red and green shaly and silty beds are common in the upper and lower parts, and an olive-drab, fine- to medium-grained sandstone is common in the middle of the unit (Kauffman, 1963). Although the Morrison Formation appears to be the thickest Jurassic formation in the Black Mountain syncline, the Swift Formation is the thickest on the Sapphire thrust plate.

CRETACEOUS ROCKS

Cretaceous rocks are exposed in the core of the Black Mountain syncline and south and southwest of Elliston. The lower part of the Cretaceous sequence accumulated in a continental setting dominated by fluvial and lacustrine environments. The upper part of the Cretaceous sequence is unique to the region south of the Bald Butte fault and consists of brackish-water, strandline, and continental deposits (Gwinn, 1961). The Kootenai Formation (unit Kk, pl. 1) is Early Cretaceous in age, and it overlies Jurassic rocks on an unconformity. The Blackleaf Formation (unit Kb, pl. 1) is Early and Late Cretaceous in age, and this unit overlies the Kootenai. The Jens (unit Kj, pl. 1) and Carten Creek (unit Kcc, pl. 1) Formations (Gwinn, 1961) form the uppermost part of the Mesozoic assemblage. The Coberly Formation, which is present west of the study area, has been eliminated from the study area by thrust faulting, and the Golden Spike Formation, which occurs to the southwest of the study area, probably was not deposited here. The Elkhorn Mountains Volcanics (unit Kem, pl. 1) overlie thrust-faulted Upper Cretaceous rocks, and a younger basalt unit (TKb, pl. 1) overlies the Elkhorn Mountains Volcanics.

Lower Cretaceous rocks of the Kootenai Formation form a thinner sequence in the study area than to the west on the Sapphire thrust plate; individual members are not easily separated in the study area. The Kootenai unconformably overlies Jurassic rocks, but an angular discordance is not evident in the study area. In the study area,

the thickness of the Kootenai Formation is about 365 m, whereas west of the Avon Valley in the Garnet Range the thickness of the Kootenai Formation is about 520 m. The five informal members of the Kootenai distinguished in the Garnet Range (A.B. French, Mobil Oil Co., written commun., 1978) are not as easily identified in the study area because (1) the uppermost quartzite (member e) appears to be absent, (2) the "gastropod limestone" that distinguishes "member d" of Kauffman (1963) is represented by thin, laterally discontinuous beds of biosparite rather than the thick (30–40 m) zones of massive biosparite typical of "member d" of the Kootenai (Kauffman, 1963) on the Sapphire thrust plate, and (3) the micritic limestone of "member b" mapped and described by Kauffman (1963) is thin or absent in the study area. The conglomerate, sandstone, siltstone, and shale beds are of fluvial origin, and the limestone beds accumulated in freshwater environments (DeCelles, 1986).

The Blackleaf Formation is generally divisible into upper and lower parts that regionally correspond to the Taft Hill and Flood Members (unit Kbt, pl. 1) (Lower Cretaceous) at the base, and the Vaughn Member (unit Kbv, pl. 1) (Upper Cretaceous) at the top (Wallace and others, 1986). According to W.A. Cobban (U.S. Geological Survey, oral commun., 1990), the uppermost beds of the Blackleaf Formation are probably earliest Cenomanian in age, but inasmuch as the uppermost Blackleaf Formation is absent in the study area, the Blackleaf that is present is probably Albian in age. The distinctive medial conglomerate and sandstone of the Vaughn Member (unit L of Mudge, 1972) that is common to the west and southwest on the Sapphire thrust plate might represent the boundary between Albian and Cenomanian beds. According to Gwinn (1965) the Flood Member was deposited in a barrier-sandbar and lagoon environment under brackish-water and freshwater conditions, the Taft Hill Member was deposited in a shoreface and lagoonal environment, and the Vaughn Member was deposited in a fluvial and strand-plain environment.

At the western edge of the study area, the Jens and Carten Creek Formations have been thrust over lower members of the Blackleaf Formation along the Little Blackfoot River. The upper part of the Jens Formation along the river is mostly olive-drab and dark-grayish-green shale and siltstone interbedded with tan- and grayish-green-weathering sandstone, bentonite, and porcellanite. The Jens Formation was deposited in a sublittoral marine and brackish-water environment (Gwinn, 1965; W.A. Cobban, U.S. Geological Survey, oral commun., 1990). The lowest beds of the Carten Creek Formation along the river are zones of blackish-gray shale, siltstone, and limestone and zones of tan-weathering, grayish-green, lithic sandstone. Biostratigraphic data from the eastern Garnet Range west of the study area (W.A. Cobban, U.S. Geological Survey, written commun., 1977 and 1983) bracketed the

age of the Jens Formation; the base of the Jens overlies middle Turonian rocks, and the base of the overlying Carten Creek Formation is middle Coniacian in age. The lowest beds of the Carten Creek Formation were deposited in a marine environment, but most of the formation was deposited in a brackish-water and strandline setting (Gwinn, 1965).

Plutons of the Boulder batholith (units Kg, Kbu, and Ku, pl. 1) intrude thrust-faulted Cretaceous sedimentary rocks and the Elkhorn Mountains Volcanics (unit K_{em}, pl. 1), which were erupted during early phases of emplacement of the Boulder batholith (Robinson and others, 1968). The Elkhorn Mountains Volcanics consist of mafic and intermediate flows, flow breccia, ash-flow tuff, and volcanoclastic rocks. An isotopic age of about 79 Ma from the Elkhorn Mountains Volcanics and of about 78–70 Ma from most plutonic rocks of the Boulder batholith (Robinson and others, 1968) indicate that volcanism predominated in this region during late Campanian and early Maastrichtian time.

CRETACEOUS OR TERTIARY ROCKS

In the headwaters of Spotted Dog Creek, a sequence of basalt and andesite of uncertain affinity (unit TKb, pl. 1) unconformably overlies the lower part of the Elkhorn Mountains Volcanics. This unit is composed mostly of lava flows, flow breccia, tuff, plugs, and dikes, and it contains subordinate amounts of volcanoclastic rocks. Ruppel (1963) provisionally included these rocks in the Elkhorn Mountains Volcanics and indicated that these rocks are older than the batholithic rocks but not much younger than andesite of the Elkhorn Mountains Volcanics. These volcanic rocks are most likely Late Cretaceous in age, but they could be as young as Paleocene.

CENOZOIC ROCKS AND DEPOSITS

Cenozoic rocks and deposits are represented by Tertiary volcanic rocks, by Tertiary lake beds and related alluvial deposits, and by glacial and alluvial deposits in modern drainages.

LATE EOCENE RHYOLITE

The oldest Cenozoic rocks occur in the Avon Valley and in tributaries to Dog Creek where late Eocene volcanic rocks of the Helena volcanic field (units Tr and Tri, pl. 1) (Chadwick, 1978, 1981; Trombetta, 1987) and Eocene or Oligocene(?) volcanoclastic rocks (unit Tt, pl. 1) are exposed. The Helena volcanic field consists of isolated patches of flows, breccia, and plugs of sanidine-quartz rhyolite in the region southwest, west, and

southeast of Helena (Chadwick, 1981, p. 305). Without the benefit of isotopic ages, Knopf (1963, p. 9) suggested that the rhyolite was younger than Oligocene clay beds, but he did not report evidence that supported an Oligocene age for the clay beds. Isotopic ages reported by Chadwick (1981) are 37.4 Ma for rhyolite along the Continental Divide southwest of Helena, 37 Ma for rhyolite flows at Mullan Pass, 37 Ma for rhyolite in Hope Creek north of Mullan Pass (Blackwell and others, 1975), and a newly determined age of 39.4 ± 1.6 Ma for rhyolite 1.8 km west of Avon. The last age corrected a previous determination of 19.6 Ma for a rhyolite flow in the Avon Valley (Chadwick, 1981, p. 305). Trombetta (1987) reported an age of 39.8 Ma (L.W. Snee, U.S. Geological Survey, personal commun. to Trombetta, 1986) from rhyolite in the Avon Valley, which Trombetta designated the "Avon volcanic complex" of the Helena volcanic field. Tuffaceous volcanoclastic deposits are exposed south of Warm Spring Creek, east of Avon, and these deposits are probably contemporaneous with, or slightly younger than, eruption of late Eocene rhyolite flows.

OLIGOCENE(?) BASALT

The basalt (unit Tb, pl. 1) consists mainly of black, brown, and red, thin-bedded vesicular basalt; about 20 m of claystone, sandstone, and gravel are interlayered with northeast-dipping basalt flows on the east side of Ophir Creek. Basalt and the interbedded sediments evidently were tilted and eroded prior to deposition of younger Oligocene sedimentary beds. Basalt flows and flow breccia of the Helena volcanic field overlie rhyolite of the Helena volcanic field, and the basalt may be early Oligocene in age. However, isotopic ages are lacking for these mafic volcanic rocks.

OLIGOCENE, MIOCENE, AND PLIOCENE SEDIMENTARY ROCKS AND DEPOSITS

Sedimentary rocks and deposits of middle to late Tertiary age in the Elliston region make up four distinct sequences that are separated by unconformities. In the parts of the study area that were mapped by reconnaissance, these sequences are shown as one unit (Ts_{gs}, pl. 1). The relative age relations and depositional environments of the units have been worked out in greatest detail in the Avon Valley, but uncertainties remain about ages because fossil evidence is lacking.

The oldest and most extensive sedimentary beds consist of gravel, siltstone, sandstone, bentonitic claystone, and lignite. These deposits represent the accumulation of volcanic ash in ponds as well as deposition of sand and gravel derived from dominantly rhyolitic source areas in low-gradient stream channels. Pardee and Schrader (1933)

designated these Tertiary lakebeds as Oligocene in age, based on fossil data reported by Douglass (1902), and this Oligocene age is retained for these beds because the lower part of the sequence unconformably overlies basalt (Oligocene?) along Ophir Creek northeast of Avon. The sequence in the southeast part of the Avon Valley, which may be as much as several hundreds of meters thick, is informally known as the Blackfoot City deposits (unit Tbc, pl. 1) of Loen (1990), which is similar in age and rock type to the Oligocene Renova Formation of southwestern Montana (Fields and others, 1985). Beds similar to these make up most of the deposits mapped as Tertiary sediment (unit Ts, pl. 1) and Tertiary sediment and gravel (unit Tsg, pl. 1) in other parts of the study area. Schmidt (1986, p. 9) indicated that interbedded brown, reddish-orange, tan, light-gray, and greenish-gray claystone, siltstone, sandstone, gravel, and volcanic ash of units Ts and Tsg underlie most of the western flank of the Helena Valley below Quaternary alluvial and colluvial deposits.

Oligocene(?) sedimentary rocks in the Snowshoe Creek area, northwest of Elliston, are locally overlain by brown nonlaminated to faintly bedded, moderately sorted silt and fine-grained sand that is interlayered with discontinuous lenses of brown trough-crossbedded conglomerate, sandstone, and pebble gravel (unit Tss, pl. 1). These beds are interpreted as deposits formed by braided streams that drained the mountainous area to the north. The deposits are tentatively correlated on the basis of stratigraphic position and rock type with deposits in other intermontane basins that have yielded middle Miocene age fossils (Rasmussen, 1977; Hanneman and others, 1985; Loen, 1986). Sequences of this age are correlated regionally with the Miocene and Pliocene Sixmile Creek Formation (Fields and others, 1985). The overall extent of Miocene deposits in the study area is poorly known, but they likely occur in areas mapped as units Tsg and Ts.

Alluvial-fan deposits (unit Tf, pl. 1) of probable late Miocene age locally overlie Oligocene(?) and middle(?) Miocene beds along the southeast side of the Avon Valley. The deposits consist mainly of coarse matrix- and clast-supported gravel composed mostly of limestone, quartzite, hornfels, slate, and granitic rocks, although sandstone, siltstone, and claystone are interspersed with the coarse deposits. The fans consist of detritus transported to the Avon Valley from the mountains to the northeast by debris flows, mudflows, and braided streams, and these fans may be several hundreds of meters in thickness. A late Miocene age is tentatively assigned to these deposits because fans in the Snowshoe Creek area overlie middle(?) Miocene siltstone and sandstone and are locally overlain by Pliocene gravel (unit Tg, pl. 1). The distribution of fan deposits in valleys outside of Avon Valley is unknown, although Tertiary fan deposits probably occur combined with other Tertiary deposits in areas mapped as units Tsg and Ts.

Gravel deposits on pediment surfaces (unit Tg, pl. 1) are the youngest Tertiary deposits in the Avon Valley and probably are of Pliocene age (Loen, 1990, p. 39). The deposits are generally a few meters thick and are composed of coarse-grained clast-supported boulder and cobble stream-channel gravel that was probably deposited by braided streams on an extensive erosional surface. Later erosion of these deposits spread the gravel down slopes, making it difficult to estimate the distribution and thickness of these deposits. In some areas bouldery colluvium conceals the lower contact, which makes the gravel deposits seem to be as much as several hundred meters thick. The vertical differences in elevation of abandoned stream channels are as much as several hundred meters, which indicates that the deposits are of different ages. Development of a pediment surface in the Avon Valley probably correlates with a regional Pliocene erosional event that affected most of southwestern Montana (Fields and others, 1985, p. 19). Veneers of boulder and cobble gravel that commonly overlie Oligocene sediment occur in many parts of the study area where Tertiary sequences were not mapped in detail; these gravel deposits are included in units Tsg or Ts.

TERTIARY AND QUATERNARY DEPOSITS

Terrace gravels (unit QTg, pl. 1), in stream valleys near the Avon Valley and west of Helena could be Pleistocene or Pliocene, or both, in age. These clast-supported terrace gravels occur along Ophir, Carpenter, and Snowshoe Creeks in the Avon area, and in the Greenhorn and Silver Creek drainages west and northwest of Helena. The gravel generally consists of poorly sorted stratified pebbles, cobbles, and boulders in a sandy matrix. Because this gravel unit contained gold in placers, the terrace deposits were mostly removed by mining in the drainages of Ophir, Carpenter, and Snowshoe Creeks. These terrace deposits may have formed during the latest period of regional pediment development (Loen, 1990).

QUATERNARY DEPOSITS

Pleistocene till (unit Qt, pl. 1) and outwash (unit Qo) occur locally, and Holocene and Pleistocene landslides (unit Ql) and alluvium and colluvium (unit Qac) form widespread veneers in the study area. Holocene alluvial fans in drainages (unit Qf) and alluvium (unit Qal) form the youngest deposits in the study area.

Till (unit Qt, pl. 1) and outwash (unit Qo, pl. 1) deposits occur mainly in Threemile Creek, where alpine glaciers moved southwest down the previously established

drainage from the Continental Divide, and in the Little Blackfoot River drainage, where glaciers moved in established drainages toward the northwest from the Boulder Mountains. Till consists of poorly sorted and nonsorted debris that ranges in size from clay particles to boulders that are several meters in diameter. Recessional moraines are well developed in the Little Blackfoot River drainage north, northeast, and southeast of Elliston. Outwash deposits form a prominent apron south and southwest of the mountain front in the Threemile Creek drainage, and these deposits consist of stratified, rounded, moderately to well-sorted, framework-supported sand and gravel. Presumably, outwash deposits formed in the Little Blackfoot River have been reworked into younger alluvial deposits after the glaciers receded.

Landslide deposits (unit Ql, pl. 1) are mostly confined to poorly consolidated and clay-rich materials along the margin of the Avon Valley. Slumps and rotational slides occur most commonly in Oligocene and Miocene lake and alluvial deposits, although a large slide developed in rhyolite 2 km west of Elliston. The well-defined morphology of some slides, such as the slide on the west side of Snowshoe Creek, suggests that some slides moved recently. Two other landslides east of Snowshoe Creek formed from downslope movement of surficial deposits and locally derived rock debris.

Poorly sorted surficial debris, which consists mainly of alluvium and colluvium (unit Qac, pl. 1) occurs widely in the study area as a veneer; this debris was mapped as a unit where the debris was thick enough to obscure underlying deposits or where the debris was thick enough to mask the origin of underlying poorly consolidated deposits. Along mountain fronts much slope wash covers the ground surface, and this material grades into alluvium of stream channels farther down the slopes.

Alluvial-fan deposits (unit Qf, pl. 1) occur in some principal valleys where tributary streams that have steep gradients converge with main water courses. These fans are mapped in Threemile, Carpenter, and Snowshoe Creeks, the Little Blackfoot River, and in the northwestern part of the Helena Valley. The fan deposits, composed mostly of poorly stratified sand, silt, and clay, are interbedded with rare layers of gravel. These deposits are probably Holocene in age because they overlie Pleistocene outwash in Threemile Creek.

Alluvium (unit Qal, pl. 1) occurs in all active drainages, but only the largest deposits are shown on plate 1. Most alluvium consists of sand and gravel, although organic-rich silt and clay are common at some places. The deposits are well stratified and show local changes in grain sizes from layers of clay to boulder beds within one vertical meter. Alluvium in stream valleys that drain lode gold-mining districts was mined by ground-sluice methods during the late 1800's. Later, alluvium in the lower

parts of Carpenter Creek was dredged for placer gold during the 1930's, and Silver Creek Valley was mined by a dry-land dredge during the 1930's and 1940's (Lyden, 1948; Loen, 1990).

PLUTONIC ROCKS

Plutonic rocks in the study area consist chiefly of the Cretaceous Boulder batholith (units Kbu and Ku, pl. 1) along the southern border of the study area and of small stocks that intrude the sedimentary cover north of the Boulder batholith (unit Kg, pl. 1). Plutons of the Boulder batholith have been mapped, described, and discussed by Knopf (1963), Becraft and others (1963), and Ruppel (1963), whereas the small stocks north of the batholith have received less intense study. North and northwest of the Boulder batholith, several small stocks occur between a pair of northwest-trending faults, the Bald Butte fault on the south and the St. Marys-Helena Valley fault on the north (fig. 3); the southeast end of this zone is in the study area, and the principal stocks in the zone are the Scratchgravel Hills (unit Ksg, pl. 1), Blackfoot City (unit Kbc, pl. 1), and Marysville (unit Kmv, pl. 1) stocks (fig. 3). These three stocks have been mineralized, and lode occurrences of gold have been exploited in the past (Elliott and others, 1992). Erosion of the stocks and their metamorphic aureoles released gold to placers in Tertiary gravel and to alluvium of modern drainages (Loen, 1990).

Age and contact relations between the large stocks of the Boulder batholith indicate that the Unionville Granodiorite (unit Ku, pl. 1), which is about 78 Ma², is slightly older than the Butte Quartz Monzonite (unit Kbu, pl. 1), which is about 77–73 Ma² (Robinson and others, 1968), but the relative ages of the smaller stocks that occur in the northwest-trending zone north of the batholith are not clear from the data collected in the study area. Knopf (1963, p. 8) regarded the small stocks as satellitic stocks of the Boulder batholith. Isotopic data available to him (Knopf, 1963, p. 8) indicated that the Marysville stock (unit Kmv, pl. 1) crystallized at about the same time as did the large mass of the Boulder batholith, and other stocks, such as the Scratchgravel Hills stock (unit Ksg, pl. 1), were considered to be similar in age to the Marysville stock. Klepper and others (1957, p. 44–45) described older intrusive rocks of the Boulder batholith as ranging in composition from quartz monzonite or monzonite to gabbro, and younger intrusive rocks of the batholith as granodiorite and monzogranite and small stocks of mafic rocks. Smedes (1966) separated early mafic stocks from the intermediate and siliceous main body of the Boulder

²Potassium-argon ages corrected using revised decay constant of Steiger and Jaeger (1977).

Table 1. Analytical data for potassium-argon isotopic-age determinations on samples from stocks in the Elliston region, Montana.

[Some samples from stocks shown only on fig. 3. See pl. 1 for some sample localities. Determinations by H.H. Mehnert. Decay constants: $^{40}\text{K}\lambda_e + \lambda_e' = 0.581 \times 10^{-10} \text{ yr}^{-1}$, $\lambda_\beta = 4.962 \times 10^{-10} \text{ yr}^{-1}$. Atomic abundance: $^{40}\text{K}/\text{K} = 1.167 \times 10^{-4}$ atom/atom (Steiger and Jaeger, 1977)]

Sample No.	Name of stock	Mineral dated	K ₂ O (percent)	Moles $^{40}\text{Ar}^*$ per gram of sample ($\times 10^{-10}$)	$^{40}\text{Ar}^*$ as percent of total ^{40}Ar	Age (Ma $\pm 1\sigma$)
HQ79-1	Scratchgravel Hills.	Hornblende	1.83, 1.86	2.643	91.4	97.0 \pm 4.2
MCQ79-1	Ogden Mountain	Biotite	9.42, 9.45	11.265	89.0	81.1 \pm 2.8
		Hornblende	0.66, 0.66	.831	77.9	85.4 \pm 3.9
FQ79-1	Dalton Mountain	Biotite	8.33, 8.13	9.614	94.3	79.4 \pm 2.6
EQ79-2	Marysville	Biotite	8.77, 8.77	10.135	85.6	78.5 \pm 2.6
GBQ79-1	Granite Butte	Biotite	4.60, 4.65	5.300	90.0	77.9 \pm 2.6
		Hornblende	0.88, 0.92	1.458	71.0	109.1 \pm 4.7
EQ79-1	Blackfoot City	Biotite	7.24, 6.99	8.014	77.7	76.6 \pm 2.6
SPQ79-1	Silver Belle	Biotite	8.83, 8.93	6.740	70.2	52.0 \pm 1.8

* Radiogenic argon.

batholith, as did Klepper and others (1957) and Knopf (1963). New isotopic data (table 1) suggest that granodiorite and monzogranite of the Marysville stock (79 Ma) and quartz monzonite, quartz monzodiorite, and granodiorite of the Blackfoot City stock (unit Kbc, pl. 1) (77 Ma) crystallized at about the same time as the quartz monzonite, granodiorite, and monzogranite of the Boulder batholith (80–72 Ma). Augite quartz monzonite and gabbro-norite of the Scratchgravel Hills stock are older than the Boulder batholith, based on age determinations of about 86 Ma (Daniel and Berg, 1981, p. 64) and about 97 Ma (table 1) in this report.

The Blackfoot City stock extends under the Avon Valley at a shallow depth beneath younger deposits, and only small separations are assigned to normal faults that cut the stock (pl. 1, cross section B–B'). Loen (1990, p. 20) determined that the positive aeromagnetic anomaly that is centered on the outcrop area of the Blackfoot City stock extends southwest into the Avon Valley, presumably marking the buried part of the stock.

STRUCTURE

The principal structural elements in the study area are northwest-trending folds, a few north- and northeast-trending folds, imbricate thrust faults within the frontal part of the Sapphire thrust plate, a northwest-trending

strike-slip fault, and north-, northeast-, and northwest-trending normal and reverse faults. Broad, open folds of Late Cretaceous age are cut by faults and plutons throughout the study area. Evidence that would establish absolute time relations between slip on thrust faults and slip on strike-slip faults is lacking in this study area, but interpretation of map data from the region west of the study area suggests that the principal strike-slip faults had some slip after thrust faults were active in Late Cretaceous time (Wallace and others, 1990). The axial surface of the Black Mountain syncline is cut by the Bald Butte fault north of the study area and by thrust faults of the Sapphire thrust plate, but the orientation of the fold axis suggests that the fold could be contemporaneous with thrust faults.

The principal differences among structures depicted on this map (pl. 1) and those on the maps of Knopf (1963) and Bierwagen (1964) are that the Bald Butte strike-slip fault has been extended through the northern part of Knopf's study area, the east-dipping thrust and reverse faults that Bierwagen (1964) placed along the west flank of the Black Mountain syncline have been eliminated, and west-dipping thrust faults were mapped along the east side of the Avon Valley.

FOLDS

The most prominent fold in the study area is the Black Mountain syncline (Bierwagen, 1964), but smaller

folds occur in rocks adjacent to the Bald Butte fault. The Black Mountain syncline plunges toward the southeast. The oldest rocks folded in the syncline are Middle Proterozoic rocks of the Snowslip Formation, and the youngest unit is the lower part of the Vaughn Member of the Blackleaf Formation (Albian), so the fold postdates at least 97 Ma, which approximates the end of Albian time according to Palmer (1984). The fold predates the Blackfoot City stock, which has a potassium-argon age of about 77 Ma (table 1). The fold projects to the southwest and west under the frontal thrust faults of the Sapphire thrust plate. Emplacement of the Sapphire thrust plate and slip on faults of the Lewis and Clark line occurred during the period 97–77 Ma (Wallace and others, 1990). The Black Mountain syncline could have been folded during Late Cretaceous time before or during the period in which thrust faults of the Sapphire thrust plate cut rocks on the west limb of the fold. Later, but still during Late Cretaceous time, right slip on the Bald Butte fault offset the nose of the syncline about 28 km to the east-southeast.

Smaller folds occur in fault blocks adjacent to the Bald Butte fault, but the origin of these smaller folds cannot be determined from the mapping data. Several plutonic bodies penetrate anticlines, as shown in the cross sections (pl. 1).

A series of open folds occur in the thrust-faulted terrane south and southwest of Elliston, and this series of folds is cut by thrust faults (cross section *E-E'*). These folds could have preceded thrust faulting or they could have formed during slip on thrust faults.

THRUST FAULTS

SAPPHIRE THRUST PLATE

Imbricate thrust faults mapped along the east side of the Avon Valley represent the frontal edge of the Sapphire thrust plate. These thrust faults are low-angle structures that dip west or southwest at 10° to 20° (cross sections *B-B'*, *C-C'*, and *D-D'*). The thrust faults juxtapose western lithofacies of Paleozoic and Mesozoic formations against eastern lithofacies of equivalent rock units. Most thrust faults place uppermost Middle Proterozoic rocks and middle and lower Paleozoic rocks over lower Cretaceous rocks on the west flank of the Black Mountain syncline. The thrust faults are not exposed and are mapped by stratigraphic and structural discordances. Paleozoic rocks of the upper plate include a thick Silver Hill Formation, a thin or absent Three Forks Formation, and thick representatives of the Ellis Group and Kootenai Formation, and they most closely resemble the lithotype and sequence of the western lithofacies of rocks described by Emmons and Calkins (1913) and Kauffman (1963) rather than the eastern lithofacies described by Knopf (1963). The rocks in

the thrust slabs along the eastern side of the Avon Valley show some differences in thickness when compared to rock units in thrust sheets to the southwest near Philipsburg (Emmons and Calkins, 1913) and to the west in the Garnet Range (Kauffman, 1963), so we consider these rocks transitional into the western lithofacies, which was thrust eastward on the Sapphire thrust plate.

Thrust faults in the study area may record two distinct periods of compressional deformation based on differences in dip angles and apparent truncation of some thrust faults. About 2.5 km northeast of Elliston, steeper north-striking thrust faults appear to be cut by gently west dipping thrusts. Cross sections *C-C'* and *D-D'* show more steeply dipping thrusts below gently dipping thrusts. The gently dipping thrusts may be younger than the underlying, more steeply dipping thrust faults. Similar multiple events of compressional deformation elsewhere in the region have been documented by Emmons and Calkins (1913), Lidke and others (1987), Wallace, Lidke, and others (1989), and Lidke and Wallace (1993).

The most eastern thrust faults juxtapose strata that cannot be matched using a minimum restoration technique because the western lithofacies does not match the eastern lithofacies in thicknesses of formations and members. Balanced cross sections suggest that thrust faults amputated a series of open folds that are buried beneath Tertiary volcanic and sedimentary cover of the Avon Valley. Because thrust faults cut preexisting structures, or because some folds formed during thrust faulting and were later cut by thrusts, some of the thrusts form younger-over-older stratigraphic relations. The total amount of shortening in the frontal zone of the Sapphire thrust plate cannot be estimated from the cross sections inasmuch as the throw on the easternmost thrust fault cannot be determined and because restorations are not possible for thrust slabs covered by younger Tertiary and Quaternary rocks.

OTHER THRUST FAULTS

Two thrust faults occur south of Greenhorn Creek and north of plutons of the Boulder batholith (Knopf, 1963). These thrusts have short traces and small displacements. They could be local structures that probably formed during intrusion of plutons of the Boulder batholith.

HIGH-ANGLE FAULTS

STRIKE-SLIP FAULT

The Bald Butte fault, which was first mapped by R.G. Schmidt in 1976, trends west-northwest through the northern part of the study area from the south side of Helena, Mont., to beyond the northern limit of the study area.

The Bald Butte fault is one of several strike-slip faults that form the Lewis and Clark line, a long-lived fault system that has been recurrently active since Middle Proterozoic time (Harrison and others, 1986; Wallace and others, 1990). The Bald Butte fault displaced Middle Proterozoic rocks along most of its trace, but a horse of Cambrian rocks is adjacent to the fault in the headwaters of Dog Creek. Although Reynolds (1979) showed the fault as a left-slip structure, Wallace and others (1990) estimated right-lateral separation of about 28 km on the Bald Butte fault using a restoration that reattached the Helena and Snowslip Formations exposed in a syncline at the head of Threemile Creek, which is 16 km northwest of Helena, to the north end of the Black Mountain syncline, which is in Threemile Creek drainage north of Avon, Mont. The period during which right slip occurred on the Bald Butte fault is not closely constrained, but right separation occurred after rocks were folded into the Black Mountain syncline, which is cut by the Bald Butte fault. The youngest rocks folded in the syncline are members of the Blackleaf Formation, so right slip on the Bald Butte fault postdates about 97 Ma (Albian). Most slip on the Bald Butte fault predates about 44–47 Ma, because volcanic rocks of the Douglas Creek volcanic field are not offset by the Bald Butte fault northwest of the study area (Wallace and others, 1990; fig. 3, this report). The Bald Butte fault may be the most seismically active fault in the Helena area (Schmidt, 1986, p. 13).

NORMAL FAULTS

Normal faults in the study area are controversial because some faults have been interpreted previously as extensional Basin-and-Range faults of Oligocene and Miocene age by Pardee (1950), Reynolds (1979), and Loen (1990). However, our data and interpretations suggest that most displacements on normal faults predate the Basin-and-Range event. The most prominent high-angle faults in the study area are the Blackfoot City and Illinois Ridge faults along the northeast side of the Avon Valley, the Dog Creek fault, the North Fork fault, and the Marsh Creek fault.

Although Pardee (1950) first showed Tertiary range-front faults along both margins of the Avon Valley, and Reynolds (1979) attributed all slip on the range-front faults to middle Miocene and younger extension, our structural interpretations indicate that most slip was Late Cretaceous in age, and only smaller amounts of slip occurred during Tertiary time. Cross section A–A' shows about 1,130 m of slip down on the southwest block across the Blackfoot City fault. Cross section B–B' shows that the Blackfoot City stock is offset about 250 m by the Blackfoot City fault, which probably represents the amount of post-Late Cretaceous slip. Our interpretation of

the amount of offset on the stock was based on relative positions of footwall roof pendants of Madison Group marble northeast and southwest of the Blackfoot City and Illinois Ridge faults, on offset contacts of the stock with Middle Proterozoic and Paleozoic rocks in Snowshoe and Ophir Creeks, and on the relatively flat contact of the roof of the stock at the head of Carpenter Creek. A total of about 880 m of slip on these high-angle faults occurred before about 77 Ma. Our estimate of a total of 250 m of separation on the Illinois Ridge and Blackfoot City faults after 77 Ma is similar to Pardee's (1950, p. 397) estimate of about 300 m of offset on the single Tertiary fault that he mapped along the range front. Beds of the Blackfoot City deposits, alluvial-fan deposits, and flows of late Eocene rhyolite and basalt are tilted to the east and southeast, and intrusive rhyolite that penetrates the Blackfoot City stock is cut by the Illinois Ridge fault, so some renewed tectonic activity occurred during Oligocene and later time along the Blackfoot City and Illinois Ridge faults.

The Dog Creek fault separates thrust-faulted Paleozoic and Mesozoic rocks on the northwest block from the Elkhorn Mountains Volcanics and the Boulder batholith on the southeastern side. The Dog Creek fault has only about 130 m of displacement down on the southeastern block. Slip on the Dog Creek fault postdates eruption of the Elkhorn Mountains Volcanics (about 80–74 Ma), but slip predates Tertiary or Late Cretaceous volcanic rocks in the southwest corner of the study area, and slip predates the small mafic stock of probable Tertiary age (unit Tms, pl. 1) mapped in Dog Creek.

A small segment of the North Fork fault extends into the study area from the northern boundary, but this fault extends at least 37 km to the northwest where it probably merges with the St. Marys-Helena Valley fault (fig. 3). North of the study area this fault follows the North Fork of Prickly Pear Creek, from which the fault derives its name. The North Fork fault shows about 300 m of downward displacement of the southwestern block. This fault offsets Middle Proterozoic rocks in the study area, and slip on the fault predates early Eocene volcanic rocks of part of the Crater Mountain volcanic field (Whipple and others, 1987), which are about 56 Ma (R.A. Zimmermann, U.S. Geological Survey, oral commun., 1986).

The total length of the Marsh Creek fault is about 20 km, but only about 5.5 km of the southeastern end of the fault occurs in the study area (fig. 3). The northwestern end of the fault appears to join the St. Marys-Helena Valley fault, and the southeastern end joins the Bald Butte fault. The Marsh Creek fault shows about 1,000 m of downward displacement of the southwestern block, and separation on this fault predates the 78-Ma Granite Butte stock (table 1) that is about 16 km northwest of the study area (fig. 3). Restoration of 28 km of right separation along the Bald Butte fault aligns the Marsh Creek fault

with the Blackfoot City fault, along which most of the separation predates 77 Ma, so the Blackfoot City and Marsh Creek faults could have been the same structure before they were offset by the Bald Butte fault in Late Cretaceous time.

Numerous small normal and reverse faults (not named) occur in the drainage basin of Greenhorn Creek, and these faults may have formed during intrusion of the adjacent Boulder batholith. West of the Scratchgravel Hills stock (fig. 3) are two normal faults that are probably related to movement along the Bald Butte fault.

REGIONAL STRATIGRAPHIC AND STRUCTURAL IMPLICATIONS

Geologic relations in the study area have important implications for regional structural interpretations, and geologic-map data in this report contradict some hypotheses about structural evolution of this region: (1) The Garnet line as described by Winston (1986) does not occur in the study area and most likely does not exist at all, and (2) much of the extensional slip attributed to the Basin-and-Range tectonic event in the Elliston area during Oligocene and Miocene time (Pardee, 1950; Reynolds, 1979; McMurtrey and others, 1965; Konizeski, 1965; and Csej-tey, 1963) is probably Late Cretaceous to pre-middle Eocene in age. This tensional tectonic event could be related to Late Cretaceous slip on faults of the Lewis and Clark line (Wallace and others, 1990) or to an Eocene tensional tectonic event described in northern Washington and Idaho (Fox and Rinehart, 1988; Rhodes and others, 1989). The ancestry of the modern large-scale topographic features in this region dates from this pre-Basin-and-Range slip on tensional faults.

THE GARNET LINE

The Garnet line, as described by Winston (1986, p. 245–258), is an east-trending Proterozoic growth fault that separates different stratigraphic sequences in the middle and upper Belt Supergroup. The Garnet line strikes east across the study area north of the Blackfoot City stock, north of Greenhorn Mountain, and north of Birdseye, Mont. Winston (1986, p. 250, fig. 5) extended the Garnet line from north of Helena eastward to the northern Bitterroot Range south of Missoula, Mont., a distance of about 300 km. Our data indicate that the Garnet line was not a growth fault, and no Middle Proterozoic structure parallels Winston's trace of the Garnet line west of Helena.

Winston's location (1986, p. 250, and fig. 6) of the Garnet line within the Elliston study area was based on supposed stratigraphic discordances in the Missoula

Group, but our study shows no stratigraphic discordances in the Missoula Group across the location of his Garnet line. North of the Blackfoot City stock Winston misidentified McNamara Formation as Mount Shields Formation, and the aberrant section that he showed in his figure 6 is actually the normal sequence of the Missoula Group (cross section A–A' of this report). West of Birdseye, Winston showed an abnormally thin sequence of Snowslip, Shepard, and Mount Shields Formations at the place where our data (cross section D–D') shows a normal thickness of these units for the terrane east and northeast of the Sapphire thrust plate. Our data show that the stratigraphic sequences and thicknesses in formations of the Missoula Group are the same at places where Winston (1986, p. 250, fig. 6) showed the sections and thicknesses as different.

Soft-sediment deformation structures were located by Winston north of the Blackfoot City stock in rocks he identified as the Mount Shields Formation (Winston, 1986, p. 252, fig. 6), and he interpreted these structures as indicators of active slip on growth faults. The soft-sediment deformation structures are actually in the McNamara Formation, and these structures are common in the McNamara Formation throughout most of the Belt basin; at most places these structures are confined to specific beds or in zones of beds. Furthermore, nearly all formations in the Belt Supergroup contain soft-sediment deformation structures. Soft-sediment deformation records only thixotropic failure; these structures do not preserve evidence for the cause of sediment failure. To attribute sediment failure to seismic shock that resulted from slip on growth faults overextends the evidence, and Winston's conclusion that soft-sediment deformation is associated with nearby growth faults is not tenable.

Winston's Garnet line connects traces of disparate segments of structures along the projected trace of the line: (1) North of Greenhorn Mountain, Winston's (1986) Garnet line is a segment of the Bald Butte fault; (2) west of Drummond, Mont., and southeast of Missoula, Mont. (fig. 1), the Garnet line parallels thrust faults (Wallace and others, 1986); and (3) in the northern Bitterroot Range he projected the Garnet line where Nold (1968) mapped the Lolo fault, but mapping by C.A. Wallace and D.J. Lidke (unpub. data, 1986–89) indicates that stratigraphic sequences and structures are continuous across the Lolo fault of Nold (1968). The above evidence indicates that in the study area and in the terrane west of the study area, Winston's Garnet line is without foundation.

TENSIONAL TECTONIC REGIMES

Previous reports have emphasized the role of Oligocene and younger Basin-and-Range faulting as the principal factor in forming the present ranges and valleys

(Pardee, 1950; Konizeski, 1957, 1965; McMurtrey and others, 1965; Csejtey, 1963; Rasmussen, 1969, 1977; Reynolds, 1979; Fields and others, 1985; Loen, 1986), but interpretations of our data from this study and from the adjoining region suggest that part of the slip on normal faults is Late Cretaceous in age (Wallace and others, 1990) and that the principal ranges and valleys in this region predate volcanic rocks of middle Eocene age. Late Oligocene and Miocene slip on normal faults may have been only one of several Late Cretaceous and Tertiary tensional tectonic overprints that shaped the present system of ranges and valleys. Interpretations developed from the study area and from adjoining areas challenge the popular concept that places faults along most mountain fronts (Pardee, 1950; Reynolds, 1979). Documentation of Late Cretaceous to middle Eocene tensional tectonics in west-central Montana fits the history of slip on the Lewis and Clark line (Wallace and others, 1990) and fits the tensional pre-middle Eocene and middle Eocene tectonic regime that has been well documented west of the study area in Idaho and Washington (Fox and Rinehart, 1988).

Interpretation of geologic and isotopic data from the study area indicate that the Avon Valley was a down-dropped block before 77 Ma. Normal slip that created the Avon Valley on the Blackfoot City fault predates intrusion and cooling of the Late Cretaceous Blackfoot City stock (77 Ma). The metamorphic aureole of the Blackfoot City stock crosses the Blackfoot City fault in Ophir and Snowshoe Creeks, so the fault displaced rocks down on the southwestern block, where the Avon Valley is now located, before the stock was intruded. From cross section A-A', we estimate that about 1,160 m of normal slip predated intrusion of the Blackfoot City stock, and from cross section B-B', we estimate that about 250 m of normal slip occurred after intrusion of the Blackfoot City stock. Late Cretaceous tensional slip on the Blackfoot City fault could have occurred during right slip on faults of the Lewis and Clark line (Wallace and others, 1990).

In the study area, prominent ranges and valleys predate late Eocene time, and we regard these uplifted ranges and down-dropped valleys as a result of a tensional tectonic regime. The Little Blackfoot River and its tributaries, which drain the highland northeast of Avon Valley, were established before late Eocene time; tributaries to the Little Blackfoot River, Dog and George Creeks, drained to the west before rhyolite of the Helena volcanic field (39 Ma) erupted because the basal contact of the rhyolite is at an elevation as high as 6,400 ft in the headwaters, and the basal contact is at 5,200 ft in the Little Blackfoot drainage southwest of Elliston and north of Avon.

In the region adjacent to the study area, emphasis of Oligocene to Miocene Basin-and-Range extension as a principal force in sculpting the modern ranges and valleys of western Montana (Pardee, 1950; Reynolds, 1979) may be misleading because these features and their drainage

systems probably predate Oligocene and Miocene time, and we attribute early uplift of mountainous areas and down-dropping of valleys to local Late Cretaceous tensional tectonics and to regional pre-middle Eocene tensional tectonics. Wallace and others (1990) showed that a tensional tectonic regime characterized the area included in our map during Late Cretaceous time. Callmeyer (1984, p. 23) showed that some tensional faulting in the eastern part of the Garnet Range, which is about 15 km southwest of this study area, is pre-Eocene in age. According to Konizeski (1965, p. 16), the Deer Lodge Valley, about 25 km southwest of the study area (fig. 1), was a broad topographic low before eruption of the middle Eocene Lowland Creek Volcanics (50–52 Ma, Smedes and Thomas, 1965). Geologic maps (Wallace and others, 1986) show that the Garnet Range was a highland and the adjacent Douglas Creek basin (Konizeski, 1961) was a lowland before middle Eocene lavas and flow breccias of the Douglas Creek volcanic field flowed down the flank of the Garnet Range into the basin at about 44–47 Ma (Carter, 1982). In the Clark Fork drainage, volcanic rocks of the Bearmouth volcanic field, which are 44–47 Ma (Carter, 1982), filled tributaries to the ancient Clark Fork, and 2 km west of the junction of Bear Creek with the Clark Fork, the Bearmouth volcanic rocks were deposited on a surface that had about 330 m of local relief. An ancestral Clark Fork appears to have drained westward between highlands that were precursors of the Garnet Range and John Long Mountains. Similarly, in the southern Sapphire Mountains (fig. 1), Hughes (1971, p. 160) showed that a valley similar to those of present Rock Creek and Willow Creek predated middle Eocene volcanic rocks, which have a potassium-argon age of about 49 Ma (Hughes, 1971, p. 174).

The traditional interpretation of tensional tectonic regimes in the vicinity of the study area has been to infer Oligocene and younger Basin-and-Range faults along the fronts of most ranges (Pardee, 1950; Reynolds, 1979; Trombetta, 1987), but many of these faults may not exist at the locations suggested, and some high-angle faults predate Basin-and-Range extensional tectonics. Southwest of our study area, Pardee (1950) placed a fault along part of the southwestern border of the Avon Valley, and Reynolds (1979, fig. 3) placed a fault along the entire southwestern border of the Avon Valley. However, contact relations are not consistent with the interpretation of Pardee (1950) and Reynolds (1979) that a fault occurs along the southwest border of Avon Valley because Tertiary fluvial and lacustrine beds extend up valleys, and contacts between older rock units and Tertiary beds approximately follow topographic contours where Tertiary beds truncate Middle Proterozoic and Phanerozoic formations and volcanic rocks (Callmeyer, 1984; Wallace and others, 1986; A.B. French, unpub. mapping, 1978). Trombetta (1987) showed numerous normal faults that separate volcanic rocks of the

Helena volcanic field from probable Oligocene sedimentary beds and from Holocene alluvium in Spotted Dog Creek and in the Avon area, but in Spotted Dog Creek and in the vicinity of Avon, our map shows depositional contact relations between Tertiary sedimentary beds and pre-middle Eocene volcanic rocks rather than fault contacts as indicated by Trombetta (1987). These relations are similar to those determined for the Douglas Creek basin 55 km to the west-northwest where Konizeski (1961) determined that early Oligocene fluvial and lacustrine beds depositionally overlie mafic and intermediate volcanic rocks of the Douglas Creek volcanic field (44–47 Ma).

Recognition of Late Cretaceous tensional tectonics and a pre-middle Eocene tensional tectonic regime in and adjacent to the study area carries important implications that extend beyond the Elliston area: (1) The early to middle Eocene and pre-middle Eocene tensional tectonic regime in northern Washington and Idaho described by Fox and Rinehart (1988), Fox and Wilson (1989), Rhodes and others (1989), and Miller and Engels (1975) probably affected west-central Montana, and this early event of tensional tectonics could have shaped much of the present topography; and (2) normal faults, previously considered by many geologists to be of Oligocene and younger Basin-and-Range origin, may have numerous tectonic overprints that include Late Cretaceous and pre-middle Eocene periods of slip, and estimates of slip amounts attributed to Basin-and-Range faulting and estimates of the percentage of extension must account for pre-Oligocene normal slip.

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