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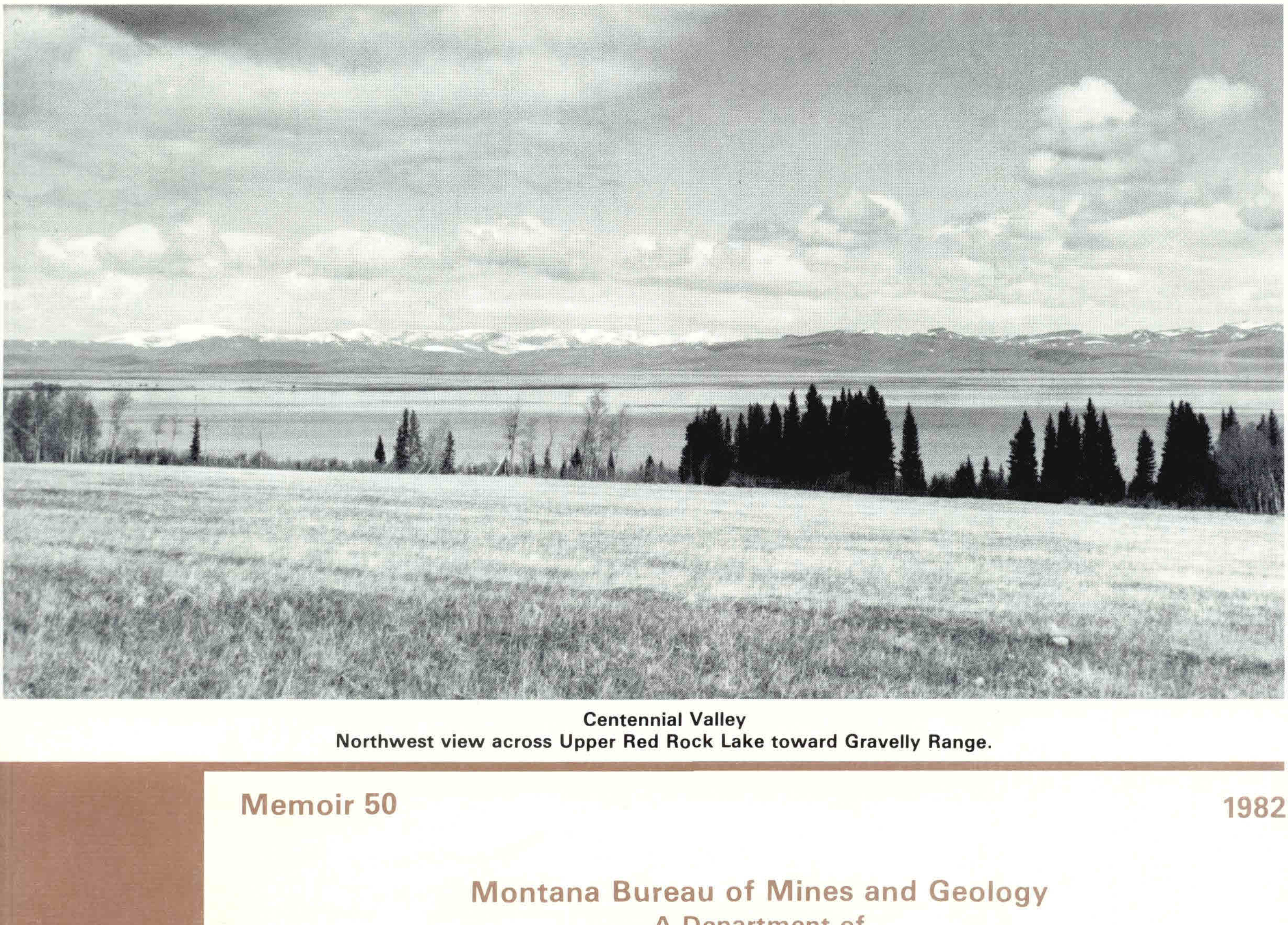
THE UPPER CENTENNIAL VALLEY, BEAVERHEAD AND MADISON COUNTIES, MONTANA

by

John L. Sonderegger, James D. Schofield, Richard B. Berg and Matthew L. Mannick

With a section on
the Madison Valley Thermal Springs
by Gerald J. Weinheimer



Centennial Valley
Northwest view across Upper Red Rock Lake toward Gravelly Range.

Memoir 50

1982

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An investigation of resources utilizing geological, geophysical, hydrochemical and geothermal methods

by

John L. Sonderegger, James D. Schofield, Richard B. Berg and Matthew L. Mannick

with a section on

The Madison Valley Thermal Springs

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Gerald J. Weinheimer

Prepared in cooperation with the U.S. Department of Energy.

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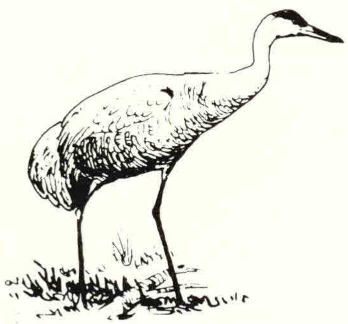

About the cover . . .

Red Rocks Lakes National Wildlife Refuge was established in 1935 to protect the rare trumpeter swan, largest of all the North American waterfowl. On this 40,000-acre refuge located in southwestern Montana, 350 trumpeter swans may be found, which is one-third of the North American swan population outside of Alaska. It is one of the most important nesting areas for these majestic birds.

The refuge lies in the eastern end of the Centennial Valley, near the headwaters of the Missouri River. The towering Centennial Mountains thrust abruptly 10,000 feet to the Continental Divide, which borders the refuge on the south and east. These rugged mountains catch the heavy snows of winter, providing a constant supply of water that replenishes the refuge's 14,000 acres of marsh and water. The flat and marshy lands of the valley floor merge into the rolling foothills of the Gravelly Range to the north. This is the habitat that provides the solitude and isolation so necessary to the trumpeter swan.

The refuge's lakes, marshes and creeks also provide an attractive habitat for a multitude of ducks. Eighteen different kinds of waterfowl, including the relatively uncommon Barrow's goldeneye, raise their young here each year.

In August and September, more than 50,000 ducks and geese congregate on the refuge before their southward migration. Between 1,500 and 2,000 whistling swans make their appearance in October. They are known locally as "cowboy geese" because of the fancied resemblance of their calls to the distant yells of cowpunchers gathering cattle.

Each spring, greater sandhill cranes nest in the refuge meadows and marshes. These grand-limbed birds are most easily observed on grasslands near Upper Red Rock Lake from May through August. Their courtship display and dance take place in April and May. *(Written commun., National Wildlife Service, 1982.)*

(Cover photo by H. L. James, MBMG.)

Preface

The Montana Bureau of Mines and Geology is a research entity of the Montana College of Mineral Science and Technology. By State law, the Bureau is charged to evaluate geologically related resources. This study of the geothermal, geologic and hydrologic resources of the upper Centennial Valley, and the geothermal resources of the Madison Valley is intended to meet that goal.

Financial support for this research was initially provided by the United States Energy Research and Development Administration under Contract No. EY-76-S-06-2426, and later support was provided by the United States Department of Energy under Cooperative Agreement No. DE-FC07-79ID12033.

The authors acknowledge Irving J. Witkind for reviewing the geologic map, and Duncan Foley for reviewing the maps and text. The authors also thank the landowners for their cooperation and assistance, especially Les Staudenmeyer and Tobe Morton. Cooperation by the United States Fish and Wildlife Service and the United States Forest Service is also appreciated.

The report is organized into sections on geologic framework, geophysical investigations, hydrology and thermal-spring sites and models, and conclusions. Each author has contributed either directly or indirectly to almost every section, particularly the portion dealing with the geologic framework. Berg mapped and described the Precambrian rocks; Mannick mapped and described the Cenozoic volcanics of the eastern half of the map area; Schofield performed the geophysical data collection and interpretation; and Sonderegger conducted the hydrologic studies, mapped the geology of the western half of the area and synthesized the final report from manuscripts provided by the coauthors and from Gerald Weinheimer's thesis (1979) on the geology and geothermal potential of the Madison Valley.

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Montana Bureau of Mines and Geology

Butte
October 28, 1982

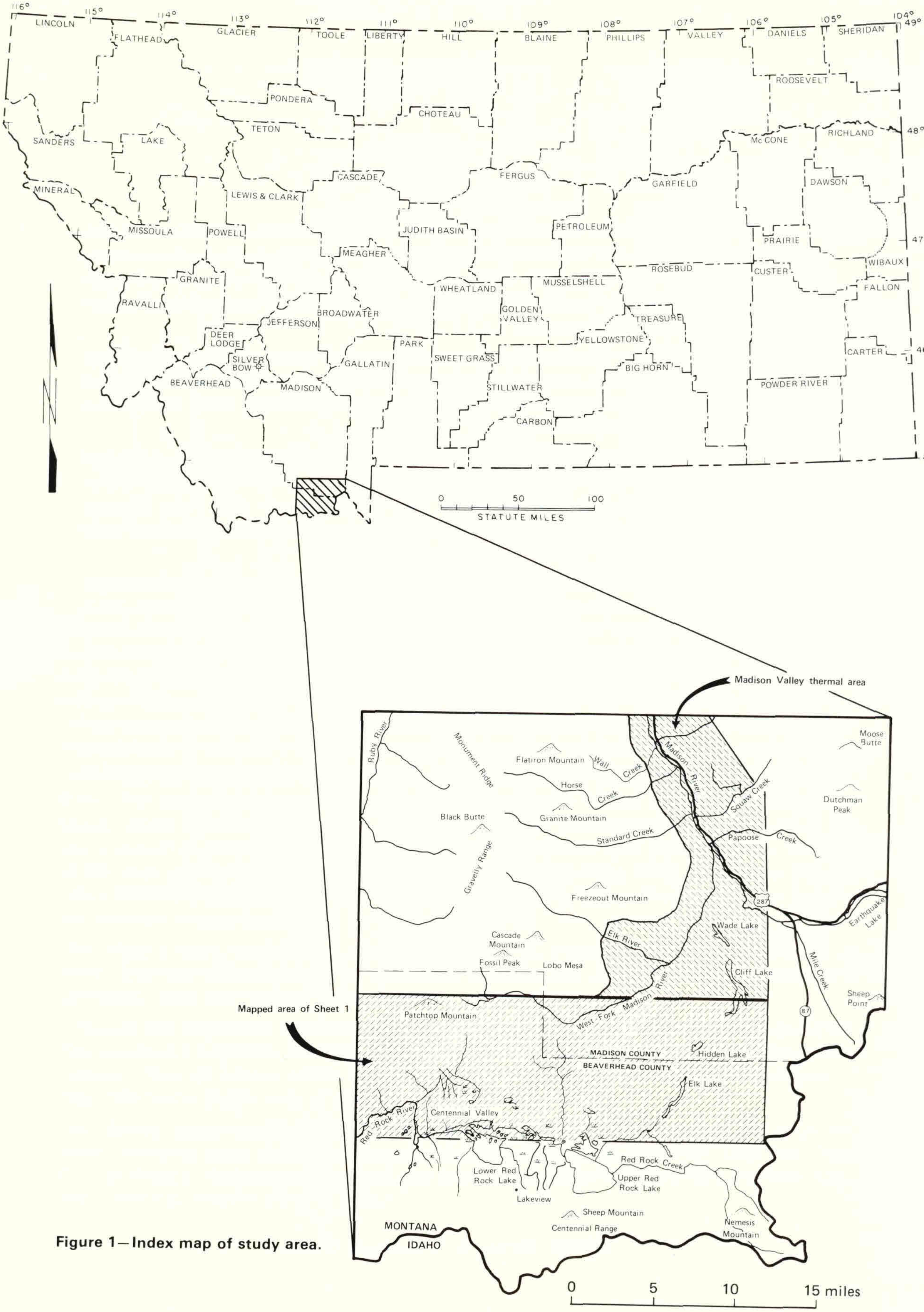


Figure 1—Index map of study area.

Abstract

Geologic, geophysical and hydrochemical investigations in the upper Centennial Valley indicate that the thermal-resource reservoir temperature is 40 to 45°C. Spring discharge is approximately six cubic feet per second, with additional discharge occurring into Quaternary sediments. The water is believed to be heated by deep circulation within the Madison Group limestones. Leakage of warm water from the Madison occurs near the crest of the doubly plunging Metzel Creek anticline where it is offset by late Cenozoic, basin-margin, normal faulting. This portion of the valley drained to the northeast along a fault-controlled valley structure prior to the eruption of the Huckleberry Ridge Tuff. The valley-fill materials attain thicknesses of up to 6,000 feet.

The most viable thermal resource evaluated is the Wolf Creek Hot Spring area in the Madison Valley. The calculated reservoir temperature is 77°C at a depth estimated to be 1,500 to 2,000 feet below land surface. The spring currently discharges about 50 gallons per minute of 68°C water; the amount of leakage to Tertiary gravels is not known but could be quite substantial.

Introduction

Purpose and scope

The area involved in the initial geothermal resource assessment study (Centennial Valley only) amounted to 275 square miles. Consequently, a hydrochemically oriented reconnaissance approach was employed to target areas for further study. The original objectives were to evaluate the thickness, areal extent, age and heat content of the young (previously undated) volcanic rocks within the Centennial Valley graben; this required the integration of geologic, geophysical and hydrologic studies and the use of a team approach.

Study of the Upper and Lower Red Rock lakes area in the Centennial Valley and of a portion of the upper Madison Valley, Montana, was initiated in 1977.

In conjunction with the initial approval of the upper project, it was learned that Gerald J. Weinheimer had commenced a mapping and hot-spring investigation along the Madison River to the north. Jerry kindly agreed to have his work integrated into this final report and to advise us concerning our well and spring inventory of the Madison Valley area, which was added to the work scope.

The area investigated in detail is depicted upon the Upper Red Rock Lake and Lower Red Rock Lake 15-minute quadrangle maps and is restricted to the area north of the Centennial Mountains **(Figure 1)**. Work by Weinheimer (1979) is the basis for most of the material presented to the north (Cliff Lake quadrangle) along the Madison Valley; geophysical studies were not conducted in the Madison Valley, but Bureau personnel conducted well and spring inventories in the area.

Field work was conducted during the summer and early fall seasons of 1977 and 1978, as well as some winter geophysical traverses in 1978, a final geologic field check and remeasurement of wells in the summer of 1979, and a winter 1980 trip to obtain additional samples for dating.

Setting

The Red Rock lakes are exceptionally shallow (average depth less than 8 feet) and are rimmed by marshy areas of variable extent depending upon the slopes. Dirt roads provide reasonable access to most of the Centennial Valley area up to the Beaverhead National Forest boundaries; road closures and restrictions limit access within the Beaverhead National Forest. The lake area itself is largely contained within the Red Rock Lakes National Wildlife Refuge, and motorized transport is not permitted within the refuge area. There are no all-weather roads into the study area. The geophysical traverses across the lakes were conducted by snowmobiling to the refuge boundary and skiing across the frozen lake surface.

The major economic factors are pasturing and raising of livestock (predominantly cattle), and the wildlife refuge. There is an outfitter based in Lakeview, and in past years the Elk Lake Lodge (leased by the U.S. Forest Service) has been open in the summer and fall.

Access along the Madison Valley is excellent owing to an all-weather highway. Dirt roads provide summer access to Cliff and Wade lakes and about four miles up the West Fork of the Madison River. Some roads to exist east of the valley on the terrace, but most of them are privately owned and maintained.

Previous work

The first detailed study of the Centennial Valley area was conducted by Honkala (1949). Succeeding papers on the area (Honkala, 1953; McKelvey and others, 1959) were directed toward the phosphate deposits, which could be economically significant. Alden (1953), in his examination of glacial activity in Montana, dealt briefly with the Centennial Mountains. The Centennial fault was one of a number of faults discussed by Pardee (1950) in his study of late Cenozoic block faulting. Interest in the Centennial region was reactivated by the Hebgen Lake earthquake, as demonstrated by the papers published in the years immediately following the earthquake (Cressman and Swanson, 1960; Honkala, 1960; Meyers and Hamilton, 1964). The Meyers and Hamilton paper is particularly important, because it attempts to place the Centennial area into a wider, regional tectonic setting. The thesis of Christie (1961) slightly overlaps the northern edge of the current study area. The most recent geologic works are a series of maps prepared by Witkind (1972, 1975a, 1975b, 1976), and Witkind and Prostka (1980).

The gravity investigation of Burfeind (1967) extends south to the 45th latitude, and the U.S. Geological Survey aeromagnetic survey referred to by Witkind (1974) involved a section of the intermountain west as far north as the Centennial Valley. A compilation of gravity data for the Yellowstone and Island Park regions (Blank and others, 1974) contains a single string of stations through the study area. Bailey (1977) completed a study of the seismicity of the Hebgen Lake-Centennial Valley area. The Centennial Valley area was included when Reilinger and others (1977) reported the results of an investigation of the rate of regional uplift. Summary papers dealing with the geology (Sonderegger, 1981) and the geophysics (Schofield, 1981) of the area have been recently published.

Geologic framework

Introduction

This investigation was conducted in the basement province of McMannis (1965), which is typified by the absence of Belt strata; a generally thin Paleozoic and Mesozoic sequence; abundant Cenozoic basin deposits and volcanic ejecta; and extensive exposures of pre-Belt metamorphic rock. The study area is bounded on the west by the Tendoy Mountains and the Snowcrest Range, which is the easternmost range with proven thrust faulting as the dominant Laramide structural mechanism; late Cenozoic mountain formation is the result of block faulting. To the north, south and east, the Gravelly Range and Centennial and Henrys Lake mountains, respectively, are believed to have been formed by block faulting. Most of the structural grain has been previously ascribed to the Laramide orogeny. The Laramide stresses from the west deformed the Paleozoic and Mesozoic rocks into broad, gentle folds in most areas. Valleys between the mountains have been infilled with Tertiary sediments and sedimentary rocks, which normally appear undeformed, although highly deformed slump structures can be found in the Madison Valley. The thickness of valley fill varies, but up to 6,600 feet are believed to be present in the Centennial Valley; possibly 6,000 feet may exist under the mouth of the West Fork of the Madison River (Gary, personal commun., 1979). Schofield (1981) indicated 15,000 feet of fill between Moose and Papoose creeks in the Madison Valley. Volcanic ash-flow tuffs of early Pleistocene age cover much of the Centennial Valley margins. Glacial outwash and tills are preserved along the valley margins, while in the lower portions of the valleys, lacustrine deposits, sand dunes and alluvium are found.

Precambrian geology

Crystalline basement rocks of Precambrian age are exposed in the mountain ranges of southwestern Montana and in Wyoming. The major ranges of southwestern Montana in which these rocks are exposed are the Highland Mountains, Tobacco Root Mountains, Greenhorn Range, Gravelly Range, Madison Range, Gallatin Range, Beartooth Mountains and the Henrys Lake Mountains in both Montana and Idaho. In Wyoming there are large areas of Precambrian metamorphic rocks in the Big Horn and Wind River mountains. Although metamorphic rocks are exposed in central Idaho, it is not certain whether these rocks are equivalent to the basement rocks exposed in Montana and Wyoming or are younger.

Precambrian metamorphic rocks in southwestern Montana are mainly exposed in mountain ranges that have been formed by Laramide block faulting. Erosion has removed the overlying Phanerozoic cover to expose the Precambrian. Except for the Precambrian rocks exposed in the Black Hills of South Dakota, knowledge of these rocks in the Great Plains comes from the few holes drilled for oil or gas that penetrated the basement.

The major rock types in the Precambrian sequence are quartzofeldspathic gneiss, migmatite, hornblende gneiss, amphibolite, marble, quartzite, banded iron formation, various types of schist, diabase dikes, ultramafic bodies and pegmatites. In the areas of lower-grade rocks, phyllite, fine-grained marble, greenstone, mica schist, gabbro and metagranodiorite are recognized. The protoliths of the quartzofeldspathic gneiss, amphibolite and hornblende gneiss are far from established.

In most areas these rocks contain mineral assemblages of the amphibolite grade of metamorphism. Local occurrences of hypersthene-bearing assemblages have been considered an indication of a metamorphic episode under granulite conditions that was followed by amphibolite-grade metamorphism. Mueller and Cordua (1976, p. 33-36) suggested that a Rb-Sr whole-rock date of $2{,}267 \pm 66$ m.y., determined on gneiss from the Tobacco Root Mountains, indicates the time of amphibolite-grade metamorphism. Additional Rb-Sr dates reported by James and Hedge (1980), indicate a slightly older age of metamorphism of 2,750 m.y. A retrograde assemblage of minerals such as talc, serpentine, chlorite and sericite is thought to be correlative with K-Ar and Rb-Sr dates of about 1,700 m.y. from southwestern Montana (Giletti, 1966). This 1,700-m.y. retrograde overprint is limited to the Precambrian rocks northwest of a NE-trending boundary that crosses the Gravelly Range in the vicinity of Horse Creek.

Precambrian rocks of greenschist facies are exposed in some parts of the Gravelly Range, in the southern Madison Range and in the Henrys Lake Mountains. In these areas it is thought that metamorphic conditions never reached the amphibolite or granulite facies, and that this is a prograde assemblage.

Intense Precambrian folding produced isoclinal folds and a penetrative fabric in the metamorphic rocks. Evidence for as many as five Precambrian tectonic events has been recognized in southwestern Montana (Reid and others, 1975). In the southern Greenhorn Range, there is evidence that Laramide folding also deformed the Precambrian rocks (Tilford, 1978, p. 127).

Because of Laramide movement of faults in southwestern Montana, it is very difficult to recognize possible Precambrian movement. Reid and others (1975, p. 122, 123) presented evidence for major Precambrian displacement on some of the faults in the northwestern part of the Beartooth Range. They correlated movement on these faults with a metamorphic event of 1,700 m.y. Precambrian movement has also been suggested for high-angle faults in the Tobacco Root Mountains and Ruby Range (Schmidt and Garihan, 1980, p. 303). A 3-km-thick mylonite zone in the southern Madison Range indicates substantial Precambrian displacement (Erslev, 1980, p. 422).

The LaHood Formation of the Belt Supergroup offers further evidence of Precambrian faulting in southwestern Montana. The distribution of this coarse-grained formation, which is well exposed in Jefferson Canyon east of Whitehall, implies rapid uplift of a source area to the south. It has been suggested that this uplift occurred along an east-west fault (McMannis, 1963, p. 433).

Gravelly Range

Precambrian metamorphic rocks are exposed along the east limb of the large SW-plunging syncline of the Gravelly Range. Phanerozoic sedimentary rocks in this syncline separate exposures of Precambrian rocks in the Gravelly Range from those on the west limb of the syncline in the Greenhorn Range.

The geology of the northern part of the Gravelly Range has been mapped by Hadley (1969a, 1969b); Heinrich and Rabbitt (1960) mapped and studied the geology of part of this same area. Information on the Precambrian rocks of the southern part of the range is less complete. A U.S. Geological Survey open-file map by Wier (1965) shows the geology of an area of approximately 18 square miles (47 sq. km) surrounding the Black Butte iron deposit in T. 11 S., R. 1 W. Mann (1954) described the geology of a large part of the Gravelly Range, but concentrated on the Phanerozoic rocks and did not describe the Precambrian rocks in detail. The Precambrian geology of the Greenhorn Range is described by Berg (1979).

Both amphibolite-grade and greenschist-grade rocks are exposed in the Gravelly Range. The metamorphic rocks of the northern part of the range are of amphibolite grade similar to those in the Tobacco Root Mountains to the north and in the Greenhorn Range to the west. Southwest of Cameron, there is an abrupt change from these amphibolite-grade rocks on the north to greenschist-grade rocks on the south (Millholland, 1976). A detailed study by Millholland showed that these greenschist facies rocks are the result of prograde metamorphism. This east-west band of low-grade rocks is flanked on the south by amphibolite-grade rocks, which in turn are bounded by greenschist-facies rocks farther to the south at Horse Creek (Jahn, 1967). Jahn dated biotite and muscovite from rocks across this transition zone and found that biotite from rocks north of Horse Creek gave a K-Ar age of 1,600 m.y., whereas biotite from rocks south of the transition gave an age of 2,600 m.y. Both Millholland and Jahn mentioned evidence of cataclasis in some of the rocks from their respective areas. Perhaps these abrupt changes in metamorphic grade are the result of faulting that juxtaposed blocks of different metamorphic grade.

Henrys Lake Mountains

The pre-Belt metamorphic rocks in the southern part of the Madison Range have not been studied in as much detail as the pre-Belt rocks exposed in the other mountain ranges of southwestern Montana. Witkind's maps of the Henrys Lake quadrangle (1972) and of the southern part of the Upper Red Rock Lake quadrangle (1976) show the distribution and lithology of pre-Belt rocks at the east end of the Centennial Valley. Erslev (1980) made a detailed study of the Precambrian rocks in the Madison Range. The following synopsis of the Precambrian history of the area at the east end of the Centennial Valley is based on Witkind (1972, 1976). Further investigations will require major revisions in the inferred geologic history.

The oldest exposed Precambrian rocks in the Henrys Lake Mountains are metasedimentary. Metamorphism has obliterated those sedimentary structures that could be used to determine the direction of younging and thus help in reconstructing the sedimentary history, and folding has further complicated the interpretation of the stratigraphic sequence. The earliest recognized event was the deposition of mud, quartz sand and a carbonate sediment. The carbonate sediment contained thin layers of silica, either of detrital origin or chemically precipitated. Amphibolite is interlayered with the metasedimentary units and represents basalt flows, or graywacke beds, or possibly both.

After compaction and lithification of the sedimentary units, granodioritic magma was intruded to form generally concordant bodies. Metamorphism followed, accompanied by deformation that produced a penetrative fabric. The age of metamorphism has not been dated in this area but may be contemporaneous with the event of 2,600 m.y. dated to the north in the Gravelly Range (Jahn, 1967). This metamorphic event produced the rock types now exposed, namely quartzite, mica schist, greenstone, marble and metagranodiorite. The mineral assemblages in these rocks indicate metamorphism of greenschist grade. In addition to quartz and dolomite, the metamorphic minerals are muscovite, chlorite, biotite, actinolite, hornblende(?), clinozoisite and albite. Probably during the waning stage of this metamorphic episode, small concordant gabbro bodies and diabase dikes were intruded. Both gabbro and diabase were partly altered to a greenschist assemblage but do not have a penetrative metamorphic fabric. Scapolite veins occur in the gabbro, and minor talc is disseminated in the marble in proximity to some diabase dikes.

Phanerozoic geology

The Paleozoic section in the upper Centennial Valley begins with Middle Cambrian units and consists of carbonate interbedded with minor shale through the Mississippian Period, after which the rocks become more clastic, containing nearly equal amounts of sandstone, shale and carbonate (including chert).The Mesozoic strata are nearly all clastic, with minor carbonate and a preponderance of shale, siltstone and sandstone, and with minor coal reported. The exposed Tertiary strata consist of basalt, limestone and sandstone in that order of abundance. The Quaternary section includes 16 units, seven of which are volcanic or volcanic related.

Paleozoic geology

The Paleozoic strata do not constitute a complete stratigraphic section; faulting near Landon camp (along the West Fork of the Madison River) has juxtaposed the Devonian upper part of the Jefferson Formation against the Precambrian. A small block of what is believed to be Meagher Limestone is preserved along the creek bottom, just west of the fault controlling the West Fork drainage, but the intervening strata, if present, are covered by Pliocene basalt and Pleistocene ash-flow tuff. The description below uses the nomenclature of Sloss (1966) for the major depositional cycles of the Paleozoic and Mesozoic eras.

The upper Centennial Valley is east of the Greenhorn fault, and while it is part of the "stable" craton, the area has a thinner preserved Cambrian section and no identifiable Ordovician age rocks; this suggests that the southern Gravelly Range was to some extent a positive element during Sauk sequence deposition. There are no preserved rocks from the Tippecanoe sequence, and Kaskaskia deposition starts with the Jefferson Formation of Middle(?) and Late Devonian age. The area received most of the sediments found in southwestern Montana during the Kaskaskia sequence. Remnants of the Big Snowy Group may be present in the map area denoted as Mississippian undifferentiated. Red soil and chalcedony float characterize this area and may be (1) weathered Mission Canyon or Lodgepole; (2) remnants of a thin basalt flow now chemically weathered on top of the limestone; or (3) weathered Big Snowy Group, too thin and poorly exposed to be identifiable. Erosion at the end of the Kaskaskia sequence was severe to the north, where Christie (1961, p. 40) noted erosional channels 200 feet deep in the Mission Canyon Limestone.

The fine, clastic Amsden Formation, which unconformably overlies the Mississippian limestones, is succeeded by the lime-rich Quadrant Formation and the Phosphoria Formation of the Absaroka sequence.

Absaroka sequence rocks are thicker in this part of southwestern Montana owing to crustal downwarping; Armstrong and Oriel (1965) attributed the basin development southwest of the study area to the migration of the miogeosyncline toward the eastern portion of Idaho. The study area was on the shoulder or flexure of this structure and contains only two recognized unconformities within this sequence. Erosion following the deposition of the Kaskaskia sequence was accompanied by southward tilting of the strata by the time the Ellis Group was deposited at the start of the Zuni sequence (McMannis, 1965).

Mesozoic geology

The Dinwoody, Woodside and Thaynes formations of Triassic age are predominantly fine-grained clastic sequences that complete the depositional cycle of the Absaroka sequence.

The Zuni sequence begins with 300 to 400 feet of preserved Jurassic rocks, mainly the Morrison Formation, unconformably overlain by the Kootenai Formation. A local angularity of this unconformity suggests additional tilting of the strata (Christie, 1961, p.89). Deposition following the Kootenai includes a thin shale and the Aspen Formation, a predominantly nonmarine unit.

The Laramide orogeny deformed the sedimentary rocks, with the major structure in the area being the Metzel Creek anticline (Honkala, 1949). The northwesterly trend and asymmetry of the anticline suggest that compressional force was directed from the west-southwest **(cross section A-A', Sheet 1, back pocket)**. This is compatible with the overview of Scholten (1967) of the Laramide tectonics of this region. Culmination of this orogenic event included thrust faulting to the west in which Scholten (1967) and Ryder (1967) attributed the origin of much of the Beaverhead Formation, which unconformably overlies pre-Laramide rocks in the western part of the study area. Scholten (1967) believed, however, that the Blacktail-Snowcrest and Gravelly "arches" contributed significant volumes of clastics to the Centennial Valley. Thus, as the compressional forces of the Laramide were waning, renewed(?) block fault uplifting must have occurred to the north, probably in Paleocene time, to provide the quartzite clasts for the Beaverhead Formation.

The Odell Creek fault separates the Centennial Mountains into two major blocks. The western downdropped block has exposures of Upper Cretaceous and younger rocks, with the exception of a small area in section 30, T. 14 S., R. 1 W., where the Meagher Limestone is mapped (Witkind and Prostka, 1980), while the uplifted eastern block consists of rocks ranging in age from Lower Cretaceous to Precambrian, with a veneer of Quaternary volcanics. Based upon the ages of the rocks involved, the Odell Creek fault is believed to be of Laramide age; however, the authors believe the N30°E orientation is inherited from Precambrian structural patterns and zones of weakness.

Cenozoic geology

From the end of Beaverhead deposition, presumably in late Paleocene or early Eocene time, there is no exposed rock record until the extrusion of older Tertiary basalt near the beginning of late Oligocene. The basalt is overlain by lower(?) Miocene sandstone and limestone believed to be of continental origin. These units dip gently into the Centennial Valley from the north at about 3 degrees. Erosion followed, and only remnants of these strata are preserved.

Outside of the study area, the Cenozoic history may be summarized as follows:

Paleocene

Laramide deformation continued into the early Tertiary. Thrusting and folding in the Blacktail and Greenhorn ranges occurred as extensions of the Bighorn uplift to the northwest. This activity west and northwest of the Centennial region resulted in the deposition of the Beaverhead Formation (Eardley, 1960; Honkala, 1960). Conglomerate of the Beaverhead Formation was derived from Paleozoic limestones and quartzitic sandstones, while farther west it contains Belt clasts.

Eocene

The Eocene was a time of uplift, volcanism and erosion in southwestern Montana. Laramide deformation continued into the early Eocene; the Beaverhead Formation was cut by a thrust fault west of the study area (Honkala, 1960). Early Eocene deformation resulted in two systems of folds and thrust faults, one trending northeast and a later system trending northwest (Eardley, 1960). Thrusts and folds to the northeast in the Madison Range and the broad gentle arch of the ancestral Teton Range, to the southeast, originated during that time. Compressional deformation ceased before middle Eocene time (McMannis, 1965; Mann, 1960). Volcanism followed Laramide deformation in the middle Eocene with the eruption of the Gallatin and Absaroka volcanics beginning east of the Centennial Valley area (Eardley, 1960).

Oligocene

Erosion and volcanism continued into the early Oligocene. Andesitic volcanism reached a climax during the late Eocene-early Oligocene east of the Centennial region in the southern portion of the Absaroka Range (Chadwick, 1978).

The Snake River basin appears to have begun to form in the Oligocene. The Snake River Plain is thought to be a continental, tensional-rift feature. Two mechanisms have been proposed for the formation of this structure—one theory is that the Snake River Plain, with associated volcanic rocks, is the result of the movement of the North American plate across a mantle plume or hot spot (Morgan, 1972; Atwater, 1970); the other theory relates the rifting to the Idaho batholith (Hamilton and Meyers, 1966). The rift appears to be oldest to the west and becomes younger to the east, where it terminates with the Island Park and Yellowstone calderas (present apex of rift).

Block faulting of the Madison Range began in Oligocene time; faults such as the Gravelly range-front fault and Cliff Lake fault broke parallel to the previously established Laramide structures.

Miocene

Erosion and volcanism dominated the Miocene history of southwestern Montana. The Black Butte basalt was extruded in the Gravelly Range north of the Centennial Valley during late Oligocene-early Miocene time. The downwarping of the Snake River Plain and subsequent volcanism were probably active southwest of the Centennial region at that time. The Snake River Plain activity does not appear, however, to have influenced the Centennial region during the Miocene.

Pliocene

The Snake River Plain structure had migrated far enough to the east by late Pliocene time to begin to influence the Centennial and Teton regions (Hamilton, 1965). The Centennial and Teton range-front faults were probably created at that time. Assuming the rift theory to be correct, the tensional forces would have created crustal thinning and downwarping in the Snake River Plain region followed by faulting and volcanism. As the rift migrated (opened) eastward, the southern flank of the Centennial Mountains and northwestern flank of the Teton Range began to dip into the downwarp. As these blocks continued to tilt into the downwarp during late Pliocene-early Pleistocene time, the uplifted edges broke to form range-front faults, which now mark the north side of the Centennial Mountains and southeast side of the Teton Range. The exposed offset along the Teton faults is approximately twice that of the Centennial fault; this difference appears to be the result of the orientation of the faults. The Centennial fault cut across preexisting Laramide structures, while the Teton fault was enhanced by cutting along them (Love, 1968).

Within the study area, the Cenozoic history may be interpreted as follows:

(1) Doming occurred along a N-S trend on the east, as evidenced by the erosion of pre-Pleistocene rocks in the eastern portion of the area. Faulting along the trend appears to be fairly local, as evidenced by the juxtaposition of Devonian and Precambrian strata south of Landon camp **(Sheet 1)** and the normal transgressive sequence that is preserved along the West Fork of the Madison River, approximately 3 kilometers to the northwest (Christie, 1961).

(2) Fluvial sedimentation and erosion were ongoing processes during Paleocene time. Sedimentation may have continued until the early Eocene, with the major unit deposited being the Beaverhead Formation. An ancestral Centennial range probably existed further south (Ryder and Scholten, 1973, fig. 10, and discussion of Pinyon-Harebell Conglomerate in Wyoming), which later foundered into the Snake River trough. The Gravelly Range was probably the major source of sediment at that time. The near absence of limestone clasts in what is mapped as the Beaverhead suggests that drainage off the Gravelly block was to the west, or that the materials from the ancestral Centennial range uplift were the volumetrically dominant influx. Drainage may have been to the east-southeast as depicted by Ryder and Scholten (1973, fig. 10).

(3) Erosion locally reduced the land surface to a virtual plain during the remainder of Eocene and lower and middle Oligocene time, as evidenced by the flat basal nature of the Oligocene basalt flow near Cayuse Spring. The area now included on the Lower Red Rock Lake quadrangle was probably a pediment. To the north a structurally active, upland erosional area must have been maintained, based upon the description by Christie (1961) of the contact between the Tertiary basalt and underlying pre-Tertiary formations and preexisting normal faults.

(4) Fluvial deposition of the unnamed Tertiary sandstone in the Centennial Valley probably occurred during the lower Miocene. The influx of clastics was sufficient to disrupt existing drainage as this unit is overlain by a limestone believed to be of freshwater origin. Consequently, the authors envision moderate uplift of the Gravelly block during the late Oligocene-early Miocene interval followed by a period of tectonic quiescence.

(5) Pliocene sedimentary units have not been recognized in outcrop within the study area. It is believed that tectonic quiescence continued throughout most of that time. Renewed uplift and

erosion probably began in mid-to-late Pliocene as evidenced by volcanic flows and reworked Beaverhead Formation to the south (Witkind and Prostka, 1980); scattered patches of reworked gravels in the vicinity of Two Drink Springs; and the topographic relief formed on the Tertiary limestone and sandstone units before deposition of the overlying early Pleistocene ash-flow tuffs.

(6) During the early Pleistocene, a moderate amount of relief did exist, based upon the outcrop pattern of the oldest Pleistocene volcanic unit compared to underlying older strata (Witkind, 1976; Witkind and Prostka, 1980). It is believed that the Huckleberry Ridge ash-flow tuff erupted from the Island Park caldera and was deposited over the low ancestral Centennial Mountains, across the Centennial Valley and up against the southern margin of the Gravelly Range (Christiansen, 1979; Mannick, 1980). On the northwest side of Deer Mountain, a minimum of 500 feet of preexisting relief is indicated by the Huckleberry Ridge-Precambrian contact **(Figures 2, 3; and Sheet 1, back pocket)**. Movement along the Centennial fault continued throughout the Quaternary, displacing the Huckleberry Ridge Tuff a minimum of 1,525 to 1,830 meters (5,000 to 6,000 feet) in the last 2.0 million years. The Centennials were sufficiently high (1.2 million years ago) to block the passage of the Mesa Falls ash flow into the upper Centennial Valley from the Island Park caldera. The only portions of the ash flow to enter the valley crossed the divide through low passes such as one near Hell

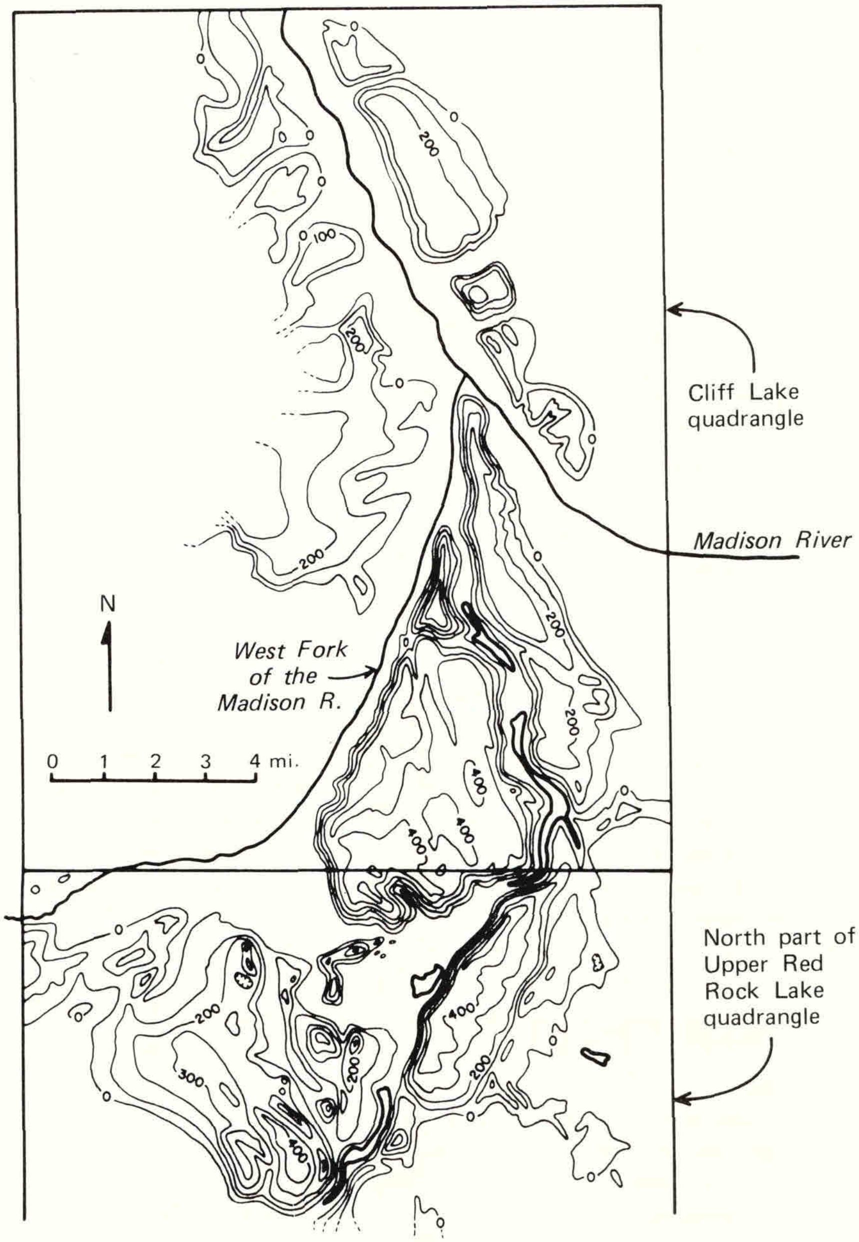


Figure 2—Isopach map of the 2.0-m.y.-old Huckleberry Ridge Tuff. Contour interval is 100 feet.

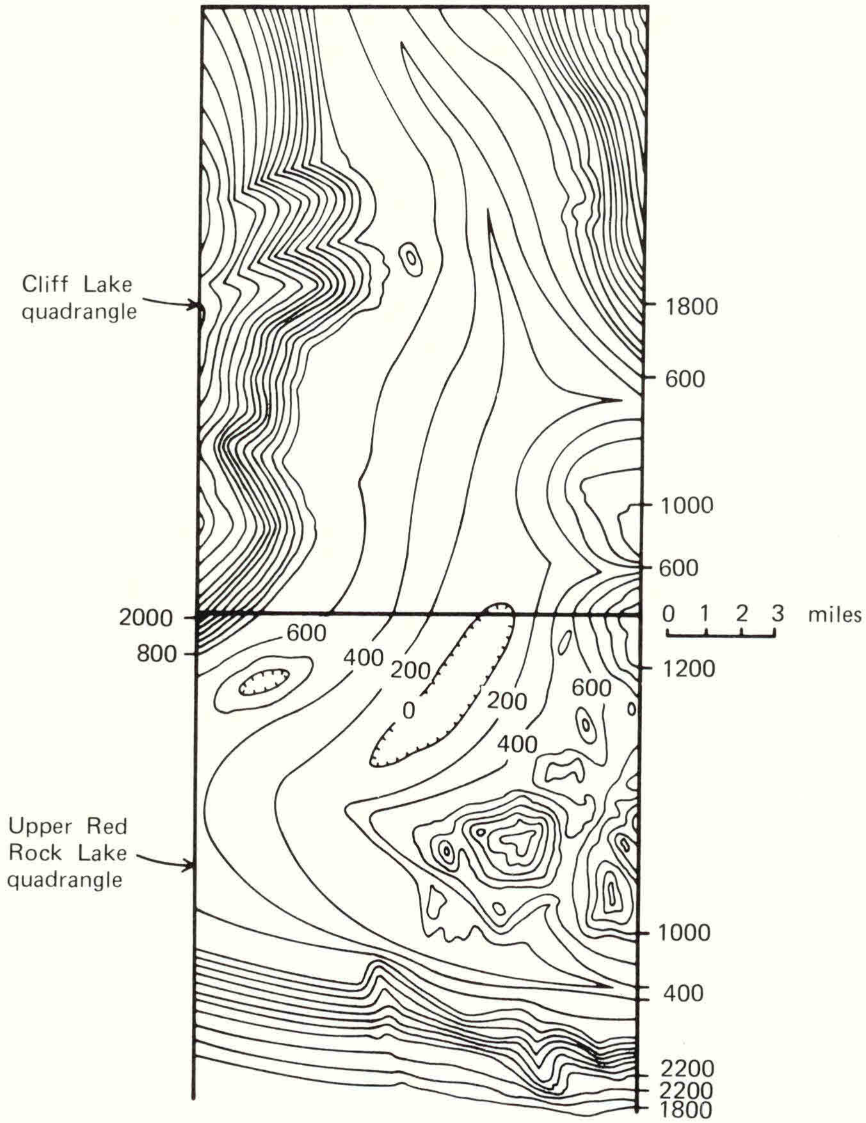


Figure 3—2.0-m.y.-old paleotopography of the Upper Red Rock Lake and Cliff Lake quadrangles. Contour interval is 200 feet. Cliff Lake quadrangle data are modified from Weinheimer (1979). The paleotopography was derived using an arbitrary datum (0 elevation) from relative thicknesses of the Huckleberry Ridge Tuff accounting for post-tuff structure.

Roaring Canyon (Witkind, 1976); the Mesa Falls ash flow is present in the Madison Valley, southeast of Cliff Lake. Major volcanism occurred in the Yellowstone caldera region 0.6 million years ago, producing the Lava Creek Tuff. This ash flow was entirely blocked from the upper Centennial Valley and is exposed along the lower southeastern slopes of the Centennial Mountains.

(7) Alpine glaciers eroded the mountain ranges surrounding the Centennial Valley at different times throughout the Pleistocene. Deposits of pre-Bull Lake, Bull Lake and Pinedale glaciations indicate that the ice originated in high cirques and traveled downward, cutting U-shaped valleys which, in many instances, reached the larger valley floors such as the Madison Valley (Weinheimer, 1979). Glacial meltwater accumulated in the Centennial Valley to form a large lake which drained at several locations, including a stream situated in the Cliff Lake fault trench. The West Fork of the Madison River began to cut a valley approximately along the Huckleberry Ridge Tuff-Precambrian contact (Gravelly fault) during late Pleistocene-early Holocene time. The Centennial Valley glacial lake eventually receded to form the Upper and Lower Red Rock lakes and associated ponds and marshes. As the lake receded, the drainage along the Cliff Lake fault trench was dammed by mass wasting to create Elk, Hidden, Goose, Otter, Cliff and Wade lakes.

(8) Based upon the altitude of features believed to be glacial lake shorelines, it appears that the rate of movement along the Centennial fault has in-

creased since Pinedale time to over an inch per year. The Centennial, Cliff Lake and southern Madison faults continue to be seismically active today.

The lithologic descriptions of the geologic units are presented on the map. Petrographic descriptions and discussion of the Precambrian and volcanic units are presented in **Appendix A** and **Appendix B**, respectively.

Thickness of volcanics and valley-fill materials

The upper Centennial Valley graben was selected for geothermal evaluation in 1976 because the area contains the most extensive exposure of Quaternary igneous rock known within the State. The open-file report by Witkind (1975b) on known and suspected active faults in western Montana provided documentation that the southern side of the valley contained an active fault system. With rocks that were believed to be correlative to the Yellowstone Group volcanics exposed on both the north and south sides of the graben, it was logical to speculate on the thickness of young volcanic rocks within the graben itself. The absence of detailed geologic mapping north of the lakes made it impossible to estimate the thickness of volcanics which might be preserved within the graben.

Figures 2 and 3 from Mannick (1980) are an isopach map of the Huckleberry Ridge Tuff and a paleotopographic map of the surface upon which the Huckleberry Ridge Tuff was deposited. The tuff thins to the west; the last recognized tuff exposure on the north side of the Centennial Valley occurs 11 miles west of the Upper Red Rock Lake quadrangle. The estimated maximum thickness of tuff within the graben is 700 to 800 feet. This is based upon preserved thickness trends that are projected back into the graben, allowing for additional filling of unknown topographic low areas and considering the mode of emplacement of an ash-flow sheet (Smith, 1960) with its consequent limitations upon the cumulative thickness of the unit.

The isopach and paleotopographic maps support the hypothesis of a pre-tuff drainage system that flowed eastward, which was originally based upon the frequency of eastward-draining streams with westward-draining mouths in the Lower Red Rock Lake quadrangle. Further support for this position is developed in the geophysical section that follows.

GEOPHYSICAL INVESTIGATION

Introduction

A geophysical study was included with the work on the upper Centennial Valley in an attempt to provide information to constrain and clarify the final geologic model. To achieve this end, a gravity and magnetics investigation was undertaken (Schofield, 1980), as well as a review of the seismicity of the region. The gravity investigation was the focus for the geophysical study because by modeling the gravitational anomaly over the Centennial Valley, an estimate of the thickness of the Cenozoic valley fill (the primary reason for using geophysics) could be made. Additionally, geophysical data may indicate which of several possible sources actually produced the warm water identified in the Upper Red Rock Lake.

Data acquisition

The gravity and magnetics data were collected in three phases: March 1978, July and August 1978, and September 1979; 193 stations were occupied during the three phases **(Sheet 2, back pocket)**. The data on the Upper and Lower Red Rock lakes were collected in March, while the lakes were still frozen. Stations not on main roads were reached on foot.

A Worden Gravity Meter (Educator model 181) was used to measure the vertical component of gravity. The meter has a scale constant of 0.4017 milligals per dial division, and the meter can be read to a tenth of a dial division. Measurements of the total intensity of the magnetic field were made with a Geometrics Portable Proton Magnetometer (Model G-816) with an advertised accuracy of ± 1 gamma. A hand compass and a 1,000-foot wire were used to establish station locations on the ice during the winter. Station interval on the lakes was 1,000 feet. On land, distinctive features (road intersections, fence corners, stream confluences) that could be precisely located on the map and easily found in the field were designated as station sites. Such sites could be located well within the desired 400-foot N-S limit, i.e., a latitude correction of less than 0.1 milligal.

Data reduction

The gravity data were reduced in an orthodox manner. Base stations were reoccupied approximately every 3 hours so the variations caused by drift could be recorded. Linear changes were assumed when making drift corrections. After the drift corrections, latitude, free-air and Bouguer corrections were made. The readings were reduced to a datum surface at 6,600 feet. Terrain corrections based on the Hammer charts (Hammer, 1939; Bible, 1962), using a bedrock density of 2.67 grams per centimeter (g/cc), were applied. Corrections for zones E through M were made, based on U.S. Geological Survey topographic maps. Zones B through D were calculated for stations in rough topography by assuming a plane with the slope estimated from the topographic maps. In terrain correction of the data, the lakes were assigned a density of 1.0 g/cc. Terrain corrections ranged from 5.8 to 0.5 milligals; most corrections were less than 1.5 milligals.

The relative importance of various factors on the accuracy of the gravity data can be estimated. The uncertainty in the reading at a typical station was 0.2 of a dial division, or ± 0.08 milligal. The error caused by imprecise north-south location is approximately ± 0.05 milligal. In areas of low relief, the error caused by mistaken elevation is expected to be approximately ± 0.20 milligal, whereas the error in mountainous terrain could be six times as great. Hand terrain corrections are usually within 10 percent; therefore, the error in terrain corrections probably ranges from ± 0.05 to ± 0.6 milligal, with the larger values associated with stations in the mountains.

Presentation

After the data were reduced, they were plotted and contoured **(Sheet 3, back pocket)**. All values were referenced to Bench Mark 1 (B.M. 1934 [W137]) south of the Upper Red Rock Lake, in section 28, T. 14 S., R. 1 W. A value of -179.0 milligals was assigned to B.M. 1 to conform to the value listed in the DOD Gravity Library. The magnetic values of **Sheet 3** are relative values only.

The interpretation of the gravity data is based on 2-D modeling, using a Talwani-type program (Talwani and others, 1959). Ten profiles **(Appendix C)** were modeled in detail (Schofield, 1980). The models consisted of a simplified geologic cross section having a single density contrast, with the observed and calculated gravity for the profile plotted above **(Figure 4)**. The basement interpretation of **Sheet 4 (back pocket)** was constructed by plotting the structures portrayed on the 2-D models in their proper location, and then using the gravity map to guide the interpolation between cross sections.

Interpretation

In general, gravity highs are associated with mountainous areas and gravity lows with valleys. Large gravity gradients are associated with major faults that juxtapose materials with different densities. Small highs within the low of the valley may be

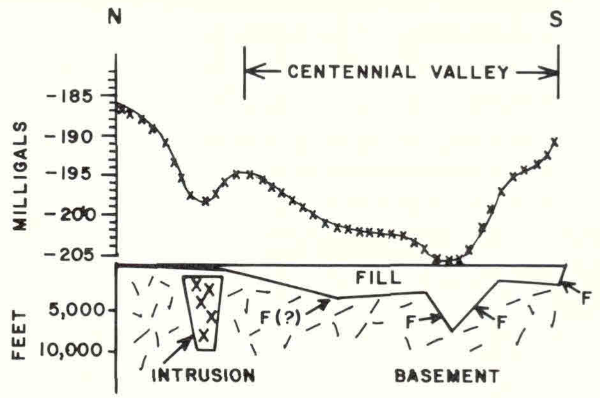


Figure 4—A 2-D gravity model of a north-south profile across the Lower Red Rock Lake, showing the Murphy Creek stock, the fill in Centennial Valley, the faults (F) controlling the shape of the valley and the N-dipping block associated with anomaly H on Sheet 3. The density contrast between both the intrusion and the fill with the basement is -0.564 g/cc. The X's represent the observed gravity; the solid line, the calculated gravity.

interpreted as uplifted basement blocks, but there is no unique structural interpretation for any given gravity anomaly. The expression of a particular body will broaden as it is buried deeper and deeper; thus, gravity anomalies of small areal extent are assumed to be near surface. Gravity anomalies within the valley can also be caused by lateral and vertical variations in density of the Cenozoic sediments.

Centennial Valley

Sheet 3 is the complete Bouguer gravity map of the eastern Centennial Valley. There is a broad, E-trending gravity low in conjunction with the Centennial Valley. The gravity anomaly across the valley is 20 milligals. The complete gravitational expression of the Centennial Valley is not known because very few gravity stations are set up in the mountains. The map of the basement of the valley shows a complex structure that is concealed beneath the nearly flat surface.

The smaller anomalous gravity features are marked with the following designations:

- (A) The steep gravity gradient with variable contour spacing between the Centennial Mountains and Lower Red Rock Lake.
- (B) The south-pointing indentation of the gravity contours where Odell Creek emerges from the Centennial Mountains.
- (C) The steep gradient between the Centennial Mountains and Upper Red Rock Lake.
- (D) The nose in the gravity contours at Upper Red Rock Lake.
- (E) The closed gravity low at Alaska Basin.
- (F) The narrow gravity low paralleling the trend of Elk and Hidden lakes.

(G) The south-pointing nose at Tepee Creek.

(H) The bend in the gravity contours north of Lower Red Rock Lake.

(I) The closed gravity low near Murphy Creek.

Centennial fault

Modeling of the gravity data suggests that the Centennial fault should actually be regarded as a fault system rather than a single fault. The Centennial fault is broken into a number of segments, and in the eastern half of the valley, there are two major branches to the fault—a branch beneath the sediments, as well as the high-angle, normal fault mapped on the surface. South of Lower Red Rock Lake, the displacement of the Centennial fault has taken place in two large steps and perhaps several smaller steps below the resolution of the data. The basement elevations south of the lower lake increase from sea level to +5,000 feet at the top of a fault block, which dips south into the surface branch of the Centennial fault **(Sheet 4)**. The S-dipping blocks are the cause of the widening of the coutours at gravity anomaly A on **Sheet 3**. A dip to the south would conform in direction with the dip of the Mesozoic formations (20 to 30° S) south of the lower lake (Pardee, 1950). Pardee stated that the Tertiary lavas on the summits also dip to the south, though less steeply. It should be noted that Witkind (1975a) mapped the Centennial Mountains east of Odell Creek as the southwest limb of a SE-plunging anticline. At Odell Creek, the Centennial fault appears to be a single unit that abuts the Odell Creek fault (anomaly B).

The Centennial fault bifurcated east of the Odell Creek fault and forms an upraised platform beneath Upper Red Rock Lake (nose D). The north branch of the Centennial fault lifts the bedrock more than 4,000 feet. The surface of the platform is inclined to the northwest. The unexpected existence of a shallow basement below Upper Red Rock Lake explains the presence of warm water discovered in the lake. Meteoric water probably circulates along the north Centennial fault, which forms the north face of the platform. The path of the warm water for the final few hundred feet to the surface is not known, but the most probable route is the suspected fault along the northern border of the upper lake. The fault was not detected on the gravity map; however, small vertical offsets would not be detected, and the fault need not be large to trigger depositional infilling of the older lake, of which Swan Lake, Upper Red Rock Lake and the surrounding swamps are but remnants. Such a fault could easily serve as a conduit for the warm water.

The imposing front of the Centennial Mountains, where the elevation rises 3,000 feet in one mile, coincides with the large gravity gradient marked C. At least 10,000 feet of vertical displacement has occurred along the Centennial fault system, and the movement has probably taken place in the last 10 m.y. An average rate of displacement of 0.3 millimeters per year (mm/yr) is easily accepted in view of present rates of uplift and subsidence. Recent releveling has revealed that the eastern Centennial Valley is being uplifted by 5 mm/yr (Reilinger and others, 1977), and rates of uplift as great as 14 mm/yr have been measured at the center of the Yellowstone caldera (Smith and Christiansen, 1980). Curiously, the Centennial fault has an obvious scarp in the western half of the valley where the rate of uplift is only a fifth of that on the eastern end, yet the eastern scarp is lost beneath till, landslides and alluvial fans. That the Pleistocene volcanics at the southern rim of the Centennial Valley were probably once continuous with the volcanics near the Continental Divide in the Centennial Mountains (Witkind, 1975a) illustrates the magnitude of movement of the bounding fault of the valley.

Odell Creek fault

The most important conclusion formed about the Centennial Valley is that the gravity low trending northeast from the eastern end of the valley (gravity anomaly F) is caused by the Cenozoic sediment fill of a valley that has been concealed by a layer of volcanics several hundred feet thick. This buried depression is a direct connection between the eastern Centennial Valley and the southern Madison Valley. Before the valley was covered by rhyolitic flows, it probably served as the drainage for the eastern Centennial Valley. The approximate width of the valley hidden by the volcanics is from Tepee Creek to Elk Lake. It would seem the western border is a N-trending fault east of Tepee Creek (anomaly G). The eastern boundary of the valley is more clearly defined. Gravity anomaly F is interpreted as a segment of the Odell Creek fault; this segment is a bounding fault of the buried valley. The mapped portion of Odell Creek fault has a strike of N30°E, which is similar to the western flank of gravity anomaly D. The Elk Lake segment **(Sheet 4, back pocket)** has the same N30°E strike as the series of lakes and ponds connecting the Centennial and Madison valleys. High seismicity was noted along a N35°E trend from Cliff Lake into the Madison Valley during a recent investigation (Bailey, 1977).

The Odell Creek fault has an estimated stratigraphic throw of 4,500 feet and a displacement of 3,000 feet in the Centennial Mountains (Honkala, 1949, p. 104). This high-angle, normal fault has a general strike of N30°E. The western edge of the platform of bedrock below Upper Red Rock Lake is formed by a segment of the Odell Creek fault, which has a vertical displacement in excess of 2,000 feet **(Figure 5)**. The Odell Creek fault has been offset by movement on the younger Centennial fault. The basement may have been dropped as much as 6,000 feet on the west side **(Figure 5)** of the Elk Lake segment of the Odell Creek fault. The movements indicated by gravity modeling are comparable to the measured movement of the fault.

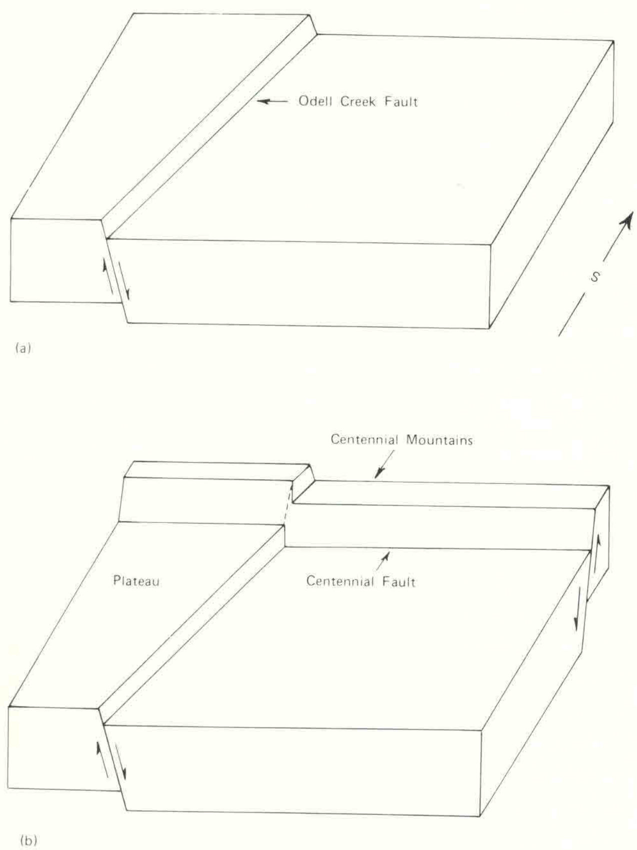


Figure 5—Block diagram showing the development of the eastern end of the Centennial Valley: (a) Odell Creek fault breaking pre-Laramide topography; (b) present configuration of the Centennial Valley (Tertiary sediments not shown).

Alaska Basin

A downdropped block, bounded on one side by a single, major, normal fault and on the other side by a series of normal faults of small displacement, with a valley floor that dips toward the large boundary fault, produces an asymmetric anomaly called the Alaska Basin (anomaly E, **Sheet 3**). On the gravity map, Centennial Valley and Alaska Basin appear as separate structures. The gravitational high between the valley and the basin coincides with a mapped horst. The horst is breached at the surface by Red Rock Creek. The small structural depression of Alaska Basin has more than 3,000 feet of relief.

Murphy Creek stock

The source of the gravity low near Murphy Creek (anomaly I, **Sheet 3)** not only has a low density but also a high susceptibility, as demonstrated by the results of a ground magnetic survey shown in Inset A, **Sheet 3** (Schofield, 1980). The dipolar magnetic anomaly suggests that the source is an intrusion. In southwestern Montana, gravity lows are often associated with intrusions. The gravity map of Montana (Bonini and other, 1973) has a low in conjunction with the Boulder batholith and another over the Tobacco Root batholith.

A 7-milligal residual anomaly was modeled as a sphere **(Figures 6, 7)**. The calculations reveal a depth of 6,000 feet to the center of the anomalous mass, which is 4.3 billion tons lighter than a uniform bedrock section would be. Such a small intrusion would cool quickly, and indeed the strong magnetic anomaly **(Figure 8)** indicates that the body is below its Curie point (Curie point for magnetite is 578°C). A steeply dipping cylinder is a good approximation of the shape of the stock.

Several locations had high magnetic gradients associated with them. In most cases volcanic rock, either basalt or rhyolite, is found on the surface. The exception not discussed previously is the N-trending low at the Lower Red Rock Lake **(Figure 9)**.

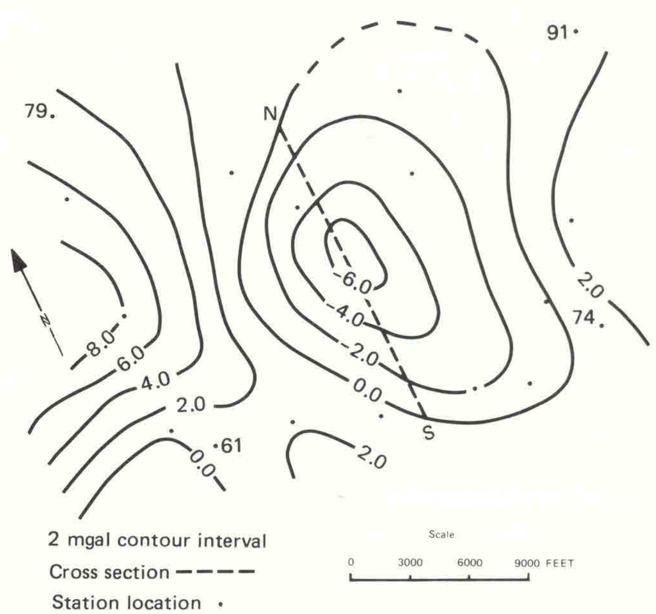


Figure 6—Residual gravity map of the Murphy Creek anomaly, created by subtracting a plane striking N35°W and dipping southwest at 3.16 milligals/mile.

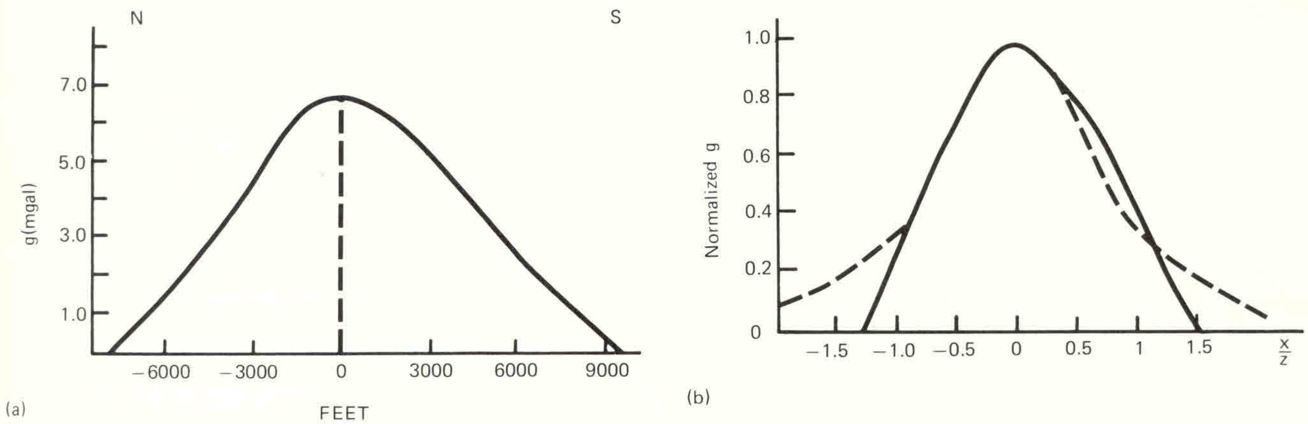


Figure 7—Profile of the Murphy Creek gravity residual: (a) cross section of the residual gravity anomaly shown on the previous figure; (b) normalized residual profile (solid line) and normalized gravity effect of a sphere (dotted line).

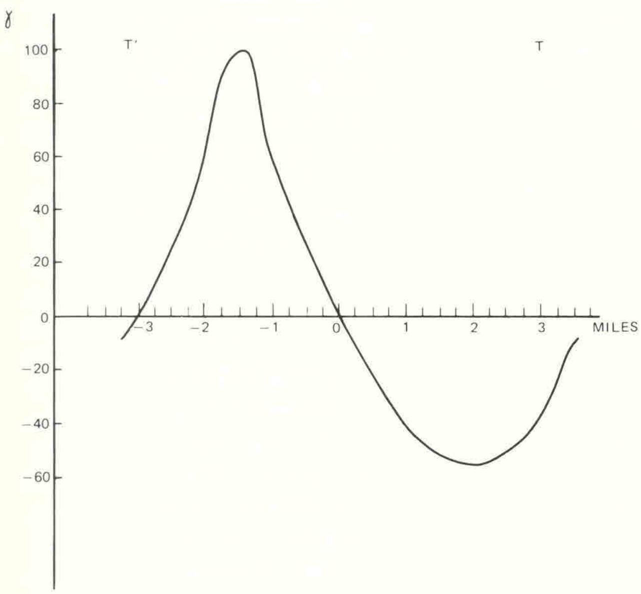


Figure 8—East-west profile of the Murphy Creek magnetic anomaly.

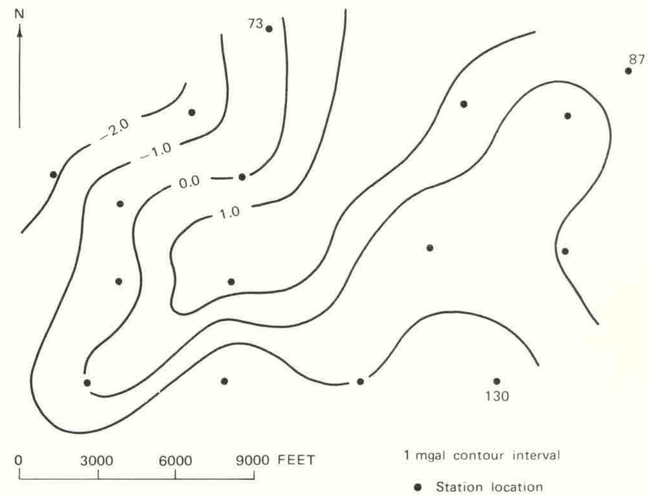


Figure 9—Residual gravity map of anomaly H, a possible rhyolite flow north of Lower Red Rock Lake; plot created by subtracting a plane striking N40°W and dipping southwest at 4.21 milligals/mile.

Conclusions

The following list summarizes the results of the Centennial Valley gravity and magnetic investigation:

(1) The Odell Creek fault is a major structure, at least of Laramide age, running from the Snake River Plain to the Madison Valley and having a displacement of as much as 6,000 feet; it has hitherto been overlooked because much of the fault is hidden by sediments in the Centennial Valley or volcanics from Elk Lake to Cliff Lake.

(2) The Centennial fault is actually a fault system generally thought of as two young, high-angle, normal faults—one forming an impressive scarp at the surface and the other resting beneath the valley surface.

(3) The fracture zone known as the Centennial fault has a total vertical displacement of at least 10,000 feet.

(4) A large block of basement material, referred to as the Upper Red Rock Lake plateau, exists at shallow depth beneath the valley floor where the younger Centennial fault intersects the Odell Creek fault.

(5) The Centennial Valley, which has 7,000 feet of valley fill, is a complex structure belied by a surface of low relief.

(6) The warm water discovered in Upper Red Rock Lake is probably meteoric water heated during

circulation in the north Centennial fault and conducted from the plateau to the surface along smaller faults.

(7) A large structural feature connecting the Centennial Valley to the Madison Valley, herein termed the Hidden valley, is covered by a layer of volcanics several hundred feet thick.

(8) The Alaska Basin is a narrow graben, with a maximum depth of approximately 3,000 feet, that is seemingly isolated from the Centennial Valley.

(9) An intrusion best described as a stock was discovered near Murphy Creek.

(10) A rhyolite(?) flow that cuts across the gravity contour trend is believed to be buried below the dunes north of Lower Red Rock Lake.

HYDROCHEMICAL STUDIES

Introduction

The Centennial Valley is located in a precipitation shadow, being on the leeward side of the Continental Divide. The weather station at Lakeview, on the south margin of the valley, had a 30-year average annual precipitation of 20.52 inches, while the main part of the valley averages 14 to 18 inches. The calendar year of 1978 was relatively dry, with Lakeview receiving only 15.67 inches of precipitation. The average temperature during 1978 was 34.7° F (1.5° C).

A spring and well inventory was conducted; springs outnumber wells in the study area. Most of the springs are currently used for agricultural application, either for stock watering or to provide ditch irrigation. Wells have been drilled for stock and domestic water. The only known irrigation well is north of the road in the SW¼ section 22, T. 14 S., R. 3 W., in the southwest corner of the area; the well was completed with a pitless adapter, and the landowner would not permit pumping and sampling of the well.

Water chemistry and temperature were the major factors in the hydrochemical survey. To evaluate mixing ratios, surface-water discharge measurements were also needed. **Appendix D** contains the hydrochemical data. **Table D-1** is a listing of springs and streams inventoried or measured; **Table D-2** contains the well-inventory information; **Table D-3** contains most of the chemical data from the water analyses. Sixty springs and nine streams or reservoirs were investigated; field determination of silica and fluoride was used to reduce the number of sites to be sampled; 37 wells were inventoried. Water samples for chemical analysis were collected from all wells for which permission could be obtained to sample and which were capable of being pumped. The data are from both the Centennial Valley and the Madison Valley.

Aquifer yields

The hydrologic interpretation of valley-fill sequences requires detailed stratigraphic knowledge of the zone(s) from which water enters a well. Comparison of water levels in wells on the north side of the Centennial Valley confirmed that different aquifer units separated by less permeable units must exist; the Anderson wells in section 20, T. 13 S., R. 2 W., are an example. The northern stock well is at a land surface altitude of 6,713.4 feet, and the 1979 water level was at an altitude of 6,649.2 feet; the southern domestic and stock well, about 1,000 feet from the first well, has a land surface 23.3 feet lower and a water level 0.9 foot higher than the first well. The wells are aligned normal to the axis of the valley; consequently, if both wells were open to the same water-bearing zone, one would expect the altitude of the water level in the northern well to be slightly greater than that of the southern well. Elevations of wells on the north side of the valley were surveyed to the nearest 0.1 foot; water-level measurements were made with a steel tape to the nearest 0.01 foot; and the northern well had not been pumped for at least one week. Drillers' logs were not filed for these wells. Reported depths are about 170 feet for the northern well and 120 feet for the southern well, suggesting that either (1) the potentiometric surface in one aquifer unit is very flat, or (2) multiple aquifer units with lower heads at greater depth exist in this area. Based upon the available information, neither case can be established; however, the two wells are believed to tap two different water-bearing zones. Their positions are such that it is inferred that the northern well intersects a stratigraphically deeper aquifer and that vertical movement of water between these strata is from overlying to underlying units.

A similar situation exists further east, where the Smith domestic well has a higher water-level altitude than the Kessler stock well in section 20, T. 13 S., R. 1 W. Also, the Smith stock well, an 11-foot piece of 4-foot-diameter galvanized culvert set in a bulldozed or backhoed excavation, definitely receives water from the shallowest sand unit.

Most of the wells on the north side of the Centennial Valley are stock wells with jackleg piston pumps, converted from hand operated to windmill operated, set on top of the casing. Consequently, no attempt was made to conduct pump tests, but access holes were drilled, tapped and plugged to permit water-level measurements.

Drillers' logs do provide static and pumping water levels for three of the stock wells on the north side of the valley. These wells range from 86 to 105 feet in depth and have specific capacities (S.C.) of 2, 2 and 3 gallons per minute (gpm) per foot of drawdown. The wells with S.C. values of 2 were completed by open-end methods, while the well with an S.C. of 3 has 7 feet of torch-cut slots. Neither completion method adequately permits the testing of the entire aquifer thickness; however, using the approximation of D. L. Coffin (personal commum., 1979)

$$T \cong 2{,}000 \times S.C.$$

a minimum transmissibility (T) for the aquifer (in gallons per day per foot) can be estimated. Thus, for an average S.C. of 2.3, the transmissibility of the Quaternary sands from which these stock wells produce water is at least 4,600 gallons per day per foot. Alternatively, using the Johnson manual (1966, p. 206) estimate of open-end completion being equivalent to 6 inches of well screen, and assuming vertical anisotropy restricts flow to the well to a 1-foot vertical section, the hydraulic conductivity, K, can be estimated to be 4,000 gallons per day per square foot. This value is undoubtedly a maximum value for the aquifer because of the assumption employed.

On the south side of the valley, the geologic units are less regular. Wells generally are developed in alluvial-fan materials but probably pass through Holocene earthflow deposits (Montgomery well). Data from drillers' logs permit estimates of the transmissibility ranging from 2,700 to 4,000 gallons per day per foot in the alluvial-fan deposits. One would expect that considerably higher T values would result from screened and gravel-packed wells.

To date, irrigation water has come from springs and creeks with one minor exception—the Breneman well, section 22, T. 14 S., R. 3 W., provides some ditch irrigation when pumped to provide stock water. The upper 1,000 feet of valley-fill materials probably have an effective porosity of at least 15 percent; thus, the ground water in storage in this zone is about 150 acre-feet per acre. Extensive development of the ground water, however, would be expected to dry up some of the valley-margin springs. Extensive development of ground water in the vicinity of the Red Rock lakes would increase the natural rate of lake-area reduction and reduce the amount of shoreline habitat for waterfowl.

The lake(s) filled most of the valley 10 to 12 thousand years ago at the end of the last glacial period. They have retreated to their present form during the intervening period of time. Any modification of the present ecological balance that increases evaporation or plant transpiration will reduce the amount of water available to maintain the lakes. Increased irrigation, particularly of the sprinkler type, using ground or surface water which normally feeds the lakes, will hasten the rate of lake-area reduction.

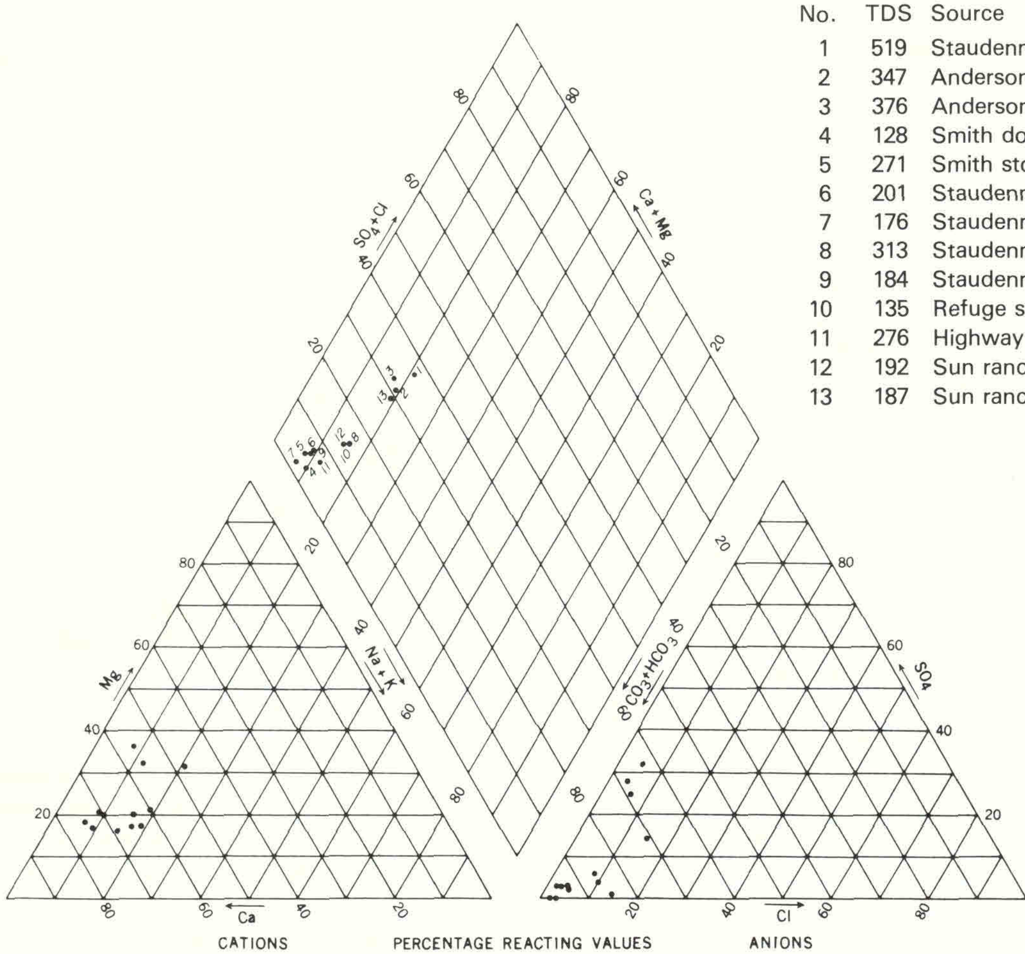
In the Madison Valley, wells in the Quaternary deposits may have fairly good yields, as evidenced by a highway department well at the rest area (T= 6,800 gal/ft x day). Wells in the Huckleberry Ridge Tuff or underlying terrace deposits generally do not have high yields but are adequate for domestic or stock use.

Figure 10 is a trilinear plot of selected well-water chemical data; numbers 1 through 10 are from the north side of the Centennial Valley; numbers 11, 12 and 13 are from the Madison Valley. These well waters commonly range from 180 to 300 milligrams per liter (mg/L) dissolved-solids content and are slightly alkaline; they are calcium bicarbonate type. Well waters 1-3, with sulfate between 25 and 32 percent and higher-than-average dissolved solids, are believed to contain a significant fraction of water leaking from a pre-Quaternary limestone, presumably the Madison Group. Well waters 4-10 are typical of waters in the Quaternary sand units. Well waters 11, 12 and 13 show the range of values for samples collected in the Madison Valley.

Springs

Spring yields are highly variable. By far the largest discharge encountered was at the Wade Lake Spring in the Madison Valley. The discharge could not be adequately measured and was visually estimated at 45 cubic feet per second (cfs). The second-largest spring volume was found at Culver Spring. **Figure 11** is a trilinear diagram; the dots numbered 11 and 12 represent the Wade and Culver springs, respectively. The total dissolved solids and carbonate-bicarbonate contents are nearly identical. The only significant difference in the water chemistry is in the cation ratios, as the Wade Lake Spring has a higher Mg/Ca ratio. These waters should be representative of the cold-spring waters in the area.

The Anderson and Staudenmeyer springs and the Staudenmeyer ranch well, located in sections 17 and 18, have a very similar water chemistry, as denoted by numbers 8, 9 and 10, respectively, on the trilinear diagram. The well water is about 30 percent higher in total dissolved solids; however, the ratios of chemical constituents remain virtually unchanged. Wood-chip float times and visual estimates of flow from the five Staudenmeyer springs resulted in a combined flow of 4.2 to 4.8 cfs (sums of lowest and highest values for springs with an uncertainty factor), while a gaged value for the combined dis-



No.	TDS	Source
1	519	Staudenmeyer house well
2	347	Anderson stock well
3	376	Anderson house well
4	128	Smith domestic well
5	271	Smith stock well
6	201	Staudenmeyer stock well NE
7	176	Staudenmeyer stock well SE
8	313	Staudenmeyer stock well SW
9	184	Staudenmeyer stock well NW
10	135	Refuge stock well
11	276	Highway rest area well
12	192	Sun ranch well
13	187	Sun ranch well

Figure 10—Trilinear diagram of water chemistry from wells.

charge at the road was 3.8 cfs. This suggests that some of the spring discharge is entering the alluvial materials along the irrigation ditch route. The ranch-house well, located in the basement, is 38 feet deep, and the owner reported a 10-foot depth to water from the top of the casing in 1963. The springs discharge above the ranch where the temperatures range from 28 to 31 °C, while the temperature of the ranch well water (passing through a pressure tank) at an outside hydrant, after 15 minutes of maximum discharge, was 10.2°C. Three nearby wells have temperatures ranging from 6.9 to 7.7°C, suggesting that the Staudenmeyer ranch well is warmer than average. Thus, both the well water-level and temperature information permissively supports some thermal water leakage into the surficial aquifer.

An estimate of the change in water composition caused by heating and rock-water interaction for the Centennial Valley warm springs is depicted in **Figure 11** by the solid arrows. Cayuse Spring (no. 13), a cold spring, issues from a Tertiary limestone at a temperature of 5.9°C; its water composition should be similar to shallow water recharging the geothermal flow system. The thermal waters (no. 8, 9 and probably 10) have essentially the same amount of calcium and increased amounts of magnesium, sodium and potassium. Consequently, the position of the thermal waters shifts because of the increased percentages of the latter constituents. The major change in anion chemistry is a fifteen-fold increase in sulfate content. Silica content is about the same for both the warm springs and the cold spring. The dashed arrow in the triangle in **Figure 11** shows the general trend that could be expected for dilution of thermal waters, based upon cation data, for the Madison Valley. The anion data show a wide scatter, owing mainly to variations in sulfate content. The Wall Canyon Spring (no. 4) has the highest dissolved-solids content; if this is assumed to be the least diluted of the thermal waters, a mixing line (which requires no precipitation or dissolution reactions) could be postulated from point 4 to point 3, as shown on the diamond by the dashed line.

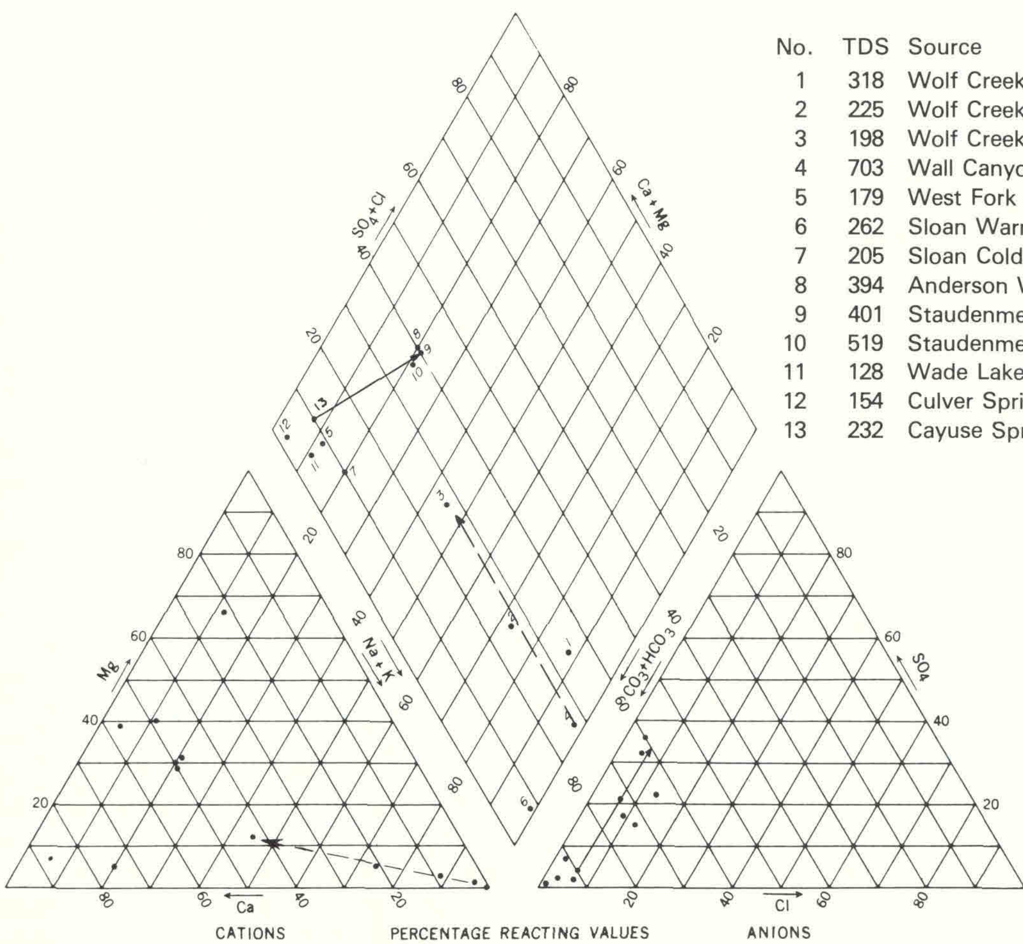
Chemical geothermometers

The water chemistry can be used with chemical geothermometers to estimate the maximum temperature at which the water has reacted with rock, without reequilibration; these are commonly known as *reservoir temperatures*. The methodology varies; the silica geothermometers that some investigators have used compare the silica (SiO_2) content (mg/kg or ppm) of the thermal water with that believed to be in equilibrium (laboratory data) for solid phases such as quartz, chalcedony, α or β cristobalite or amorphous silica, while the cation geothermometers use ppm ratios of the elements Na, K, Ca and Mg (Fournier, 1981, p. 114).

In the Centennial Valley, the thermal springs have water chemistry with a definite gypsiferous limestone signature. Quartz and chalcedony geothermometers yield reservoir temperatures of 66 and 34°C, respectively. The Na-K-Ca geothermometer yields a temperature of 52°C, owing mainly to the Na/K ratio 6.4. The high Mg and Ca content of the water suggests that temperatures were mild (Mg/Ca molar ratio of 0.59). By comparison, Madison Group water from a deep water well at Colstrip, Montana, with a 97°C *in situ* temperature, yielded 124 and 96°C with the quartz and chalcedony geothermometers, respectively, and had a Na-K-Ca temperature of 115°C. The Na/K ratio of this hot water was 3.15, and the Mg/Ca ratio was 0.16. The magnesium correction (Fournier and Potter, 1979) is 18°C, giving a Mg-corrected Na-K-Ca temperature of 97°C, which agrees with the chalcedony geothermometer.

The magnesium correction is not applied to waters that yield a Na-K-Ca temperature of less than 70°C (Fournier and Potter, 1979, p. 1547), so this correction does not apply to the Centennial Valley thermal waters. Based upon the data available, and assuming no dilution, the reservoir temperature is believed to be 45°C or less, with 40°C probably being a reasonable estimate of the maximum reservoir temperature. The highest chalcedony temperature calculated from these springs is 37°C. Chloride content varies from 9.0 to 10.0 mg/L, and silica from 20.1 to 23.3 mg/L; these data prohibit the use of most dilution models for the spring data.

If the water from the Staudenmeyer ranch-house well (T = 10.2°C) is considered, owing to its high dissolved-solids content, to represent the best water sample, chalcedony and Na-K-Ca temperatures of 48 and 62°C, respectively, are calculated. The silica content may be elevated owing to contact with detrital volcanic material, and the maximum



No.	TDS	Source
1	318	Wolf Creek Hot Spring
2	225	Wolf Creek Warm Spring
3	198	Wolf Creek Cold Spring
4	703	Wall Canyon Warm Spring
5	179	West Fork Swimming Hole
6	262	Sloan Warm Spring
7	205	Sloan Cold Spring
8	394	Anderson Warm Spring
9	401	Staudenmeyer Warm Spring
10	519	Staudenmeyer house well
11	128	Wade Lake Spring
12	154	Culver Spring
13	232	Cayuse Spring

Figure 11—Trilinear diagram of water chemistry from springs and an associated well.

temperature for the thermal system should be about 45°C if significant mixing or precipitation have not occurred.

The silica geothermometry becomes almost useless when water interacts with glass-rich acidic rocks, such as the Huckleberry Ridge Tuff. Samples of this rock and the Mesa Falls Tuff were ground to medium- to fine-sand size and leached with distilled water in the laboratory at different temperatures. Within three months, the silica content in the water plotted between the α and β cristobalite curves for laboratory temperatures of 27 and 57°C.

This difficulty can be best discussed in terms of the Sloan warm and cold springs, which occur in section 19, T. 12 S., R. 1 E., along the West Fork of the Madison River. The springs have very different water chemistry. The cold spring (analysis 78M0396; also **Figure 11**, no. 7) discharges a calcium bicarbonate water with a silica content of 67.2 mg/L, while the warm spring (analysis 78M0397; **Figure 11**, no. 6) discharges a sodium bicarbonate water with a silica content of 50.9 mg/L.

For the warm spring, quartz and chalcedony geothermometers give calculated reservoir temperatures of 103 and 73°C, respectively, while the Na-K-Ca geothermometer calculated value is 98°C. The β value in Na-K-Ca calculation (Fournier, 1977) has little effect upon the calculated temperature (if $\beta = 1/3$, T = 102°C), and the magnesium correction is negative and should be ignored (Fournier and Potter, 1979). In this instance, the quartz temperature more closely approximates the Na-K-Ca temperature.

By comparison, the cold-spring water yields calculated reservoir temperatures of 87, 66 and 18°C for the chalcedony, α cristobalite and β cristobalite geothermometers, respectively. The Na-K-Ca temperature is 47°C, which also appears to be unrealistically high. The two springs are no more than 10 feet apart, and it is difficult to envision the water from the cold spring, with its radically different chemistry, as ever having been heated significantly.

The other warm spring known along the West Fork of the Madison River also has an unusual water chemistry, as depicted in **Figure 11**, no. 5. This is a magnesium bicarbonate water with quartz and chalcedony temperatures of 50 and 17°C, respectively; a Na-K-Ca temperature of 23°C; and an observed temperature of 28°C near the bottom of the pond and 26°C at the pond outflow. The water (analysis 78MO-395) is quite dilute, with a dissolved-solids content of only 179 mg/L; however, magnesium constitutes ⅔ of the cation equivalents in the water. This water composition has most likely resulted from mixing.

The physical model presented by Weinheimer (1979) in the following section appears feasible, with a deep, thermal component mixing with dilute, magnesium-rich, shallow ground water. The cause of the high magnesium content could be the leaching of the basalt. An exposure of basalt is mapped less than one mile to the north (Weinheimer, 1979). The rising thermal water (or vapor) would have to be extremely dilute. Constituents not included in the tabled analysis include: I <0.01; Sb <0.2; NH_4 (as N) <0.03; Br <0.1; Sr $=0.12$; and Hg <0.0003, all in mg/L. If a dilution of 10:1 was assumed, and all chloride and sodium came from the thermal component, the maximum values for Cl and Na in the deep thermal water would be 27.5 and 48 mg/L, respectively. This presents the dilemma of having either (a) a considerably more dilute source water than seen elsewhere (compare Wall Canyon Spring and Wolf Creek Hot Spring), or (b) a vapor phase, condensing at shallow depth and probably mixing with shallow ground water, in an area devoid of any other manifestations of a vapor-dominated system.

To date, the authors have not resolved this dilemma. The Cl/SO_4 ratio differs from other springs, and the presence of H_2S precluded resolving this problem with oxygen isotopic data. The simplest interpretation is that there is an unmapped basalt dike in the vicinity, which permits the circulation of water to a depth of about one kilometer, and that the lateral movement of water is stopped by a (buried) range-front fault, thereby causing upward movement of the heated water. This hypothesis is not very satisfactory in terms of magnesium reactivity, but is violates none of the observed facts from this or adjacent areas.

In the Madison Valley, the two major spring areas are near the mouth of Wall Canyon and along Wolf Creek. The Wall Canyon Spring yields the highest dissolved solids (analysis 78M0293; **Figure 11**, no. 4). Chalcedony and the Mg-corrected Na-K-Ca geothermometers yield temperatures of 63 and 76°C, respectively. The observed temperature of this water (24°C) is probably caused by conductive cooling as alkalis make up 96 percent of the major cations, thus precluding the possibility of significant dilution. A small addition of shallow ground water could substantially alter the magnesium correction, however, without significantly altering the Na/K ratio (Na/K temperature is 117°C; Na-K-Ca temperature is 126°C). The low silica content of the water (41.7 mg/L) suggests that minor dilution is not the case, and the estimated reservoir temperature is 70°C.

The Wolf Creek Hot Spring is apparently not related to the Wall Canyon Spring. Chalcedony and Na-K-Ca geothermometers yield maximum temperatures of 77 and 78°C, respectively. The maximum observed temperature is 68°C, suggesting that equilibration has occurred at a relatively shallow depth.

Varying amounts of dilution are shown on **Figure 11** as no. 2, 3; these points do not project back the spring itself (no. 1) in the diamond portion of the diagram, mainly because of the greater sulfate content of water from the hot spring. Three analyses of the spring (78M0292, 78M0425 and 79M3773) in 1977 and 1978 show little change. An earlier sample by the U.S. Geological Survey (Leonard and others, 1978) has somewhat higher sodium, sulfate and silica, and lower calcium and magnesium. It was this earlier sample that produced the highest calculated reservoir temperatures; the reservoir temperature is believed to be 77°C.

The Wolf Creek Hot Spring discharged about 50 gpm prior to the use of heavy equipment intended to increase the flow. The 9°C drop from calculated reservoir temperature to surface temperature can be used to provide a crude estimate of the depth of equilibration. The Precambrian metamorphic bedrock should have a thermal conductivity similar to that of granite or granodiorite. Considering the flow to be adequate for relatively rapid ascent, the depth to the zone of equilibration (presumably the reservoir) is probably 1,500 to 2,000 feet.

Centennial Valley interpretation

A lake-temperature survey showed warmer water along the northern side of Upper Red Rock Lake. Heat-sensing imagery was flown over the northern side of the valley during the fall of 1977. An intense rainstorm the afternoon before the flight resulted in ponding of water in almost all areas except the sand dunes. Air temperature was slightly below freezing between 3:00 and 5:30 a.m., while ground-truth data were collected and the overflight occurred. Water temperatures on the lakes and river were considered to be the only reliable values under these conditions. Temperatures ranged from 8.5 to 15.5°C. A series of warm zones was identified from the computer printout of the digital data, extending from the NW¼ sec. 15, T. 14 S., R. 1 W., to the SW¼ sec. 35, T. 13 S., R. 2 W.

The alignment of this thermal-water zone southeast of the thermal springs is subparallel to Quaternary age faults identified by low-altitude aerial reconnaissance. These fractures are interpreted to be small-scale step faults on the southernmost margin of the Gravelly Range. These faults are believed to aid in localizing the ascent of warm water from the Madison Group limestones. The heat source could be either a postulated intrusive or deep circulation; based upon the small size of the postulated body and the rapid cooling it would undergo, it appears more probable to assign the source of the heat to the deep circulation of meteoric water.

MADISON VALLEY THERMAL SPRINGS

by

Gerald J. Weinheimer

Introduction

Six known thermal springs occur in the upper Madison Valley **(Figure 1)**; one is located 2 kilometers north of Ennis and five are 45 to 60 kilometers south of Ennis. These five will be discussed individually in this section.

Wolf Creek Hot Spring was the only thermal spring in the valley to receive more than cursory attention prior to 1976. During the summer of 1975, Michael Galloway and John Goering of Montana State University conducted a soil-temperature survey in the Wolf Creek area and found the warmest soil around the spring and along the Wolf Creek Hot Spring fault (Galloway, 1977). A preliminary base map was constructed for the project and was used in subsequent investigations at the spring.

Extensive field work by the author in 1976 at Wolf Creek included resistivity and seismic surveys, as well as geologic mapping of the area at a scale of 1:24,000. Field work at the other springs described below consisted of temperature measurements and mapping of the local geology.

In 1977, extensive sampling and analysis of thermal springs and cold waters in the valley were conducted by the Montana Bureau of Mines and Geology as partial support of this study (Weinheimer, 1979). In addition, a deeper seismic survey was run at Wolf Creek by the author and a field crew, and mapping of the spring areas and geology continued.

Wolf Creek Hot Spring

Wolf Creek Hot Spring is located on the Sun ranch (NW¼ sec. 9, T. 10 S., R. 1 E.) in the upper Madison Valley, 45 kilometers south of Ennis, Montana.

A shallow resistivity survey conducted during the summer of 1976, using the Wenner array, began with four resistivity depth profiles. Electrodes were spaced to provide measurements at depth intervals of 5 meters, to a maximum apparent depth of 35 meters. Two stations were located on each side of the Wolf Creek Hot Spring fault. These data showed no distinct change in resistivity across the fault. A resistivity grid system consisting of 120 stations, 30 meters apart, was established in the Wolf Creek Hot Spring-Warm Pond area in order to locate conductive zones within the outwash. The resistance was measured at apparent depths of 10 and 20 meters.

A seismic survey, using a Bison signal-enhancer seismograph with an 8-pound hammer, was conducted across the Wolf Creek Hot Spring fault **(Figure 12)** in 1976. The survey encountered only outwash to the presumed maximum depth of penetration of 28 meters. In the summer of 1977, a seismic survey was run using an improved system, the Bison seismograph, which was equipped with an 18-pound hammer, improved cable and geophones, and was run by an experienced operator (Clyde Boyer, personal commun., 1978). This survey did not encounter rhyolite tuff to its assumed maximum penetration depth of 70 meters.

The highest spring discharge temperatures were recorded during the summer of 1976, when the main spring was measured as 68°C and the adjacent warm pond as 44°C. During the winter of 1976, Sun ranch personnel dug a ditch across the hot spring and warm pond in an attempt to increase the flow of water for irrigation. The heavy equipment broke the calcite-cemented conduit of the main hot spring, allowing near-surface ground water to mix with the thermal water, which lowered the temperature to 56°C. The water in the warm-pond area, at different locations along the trench, ranges from 11 to 33°C.

The springs issue through the Pinedale outwash plain of the Wolf Creek glacier. The outwash contains Precambrian gneiss and granite cobbles from the Madison Range. Under this glacial deposit, 30 meters of Huckleberry Ridge Tuff are underlain by pre-Pleistocene alluvial gravel that lies on the Precambrian metamorphic basement complex **(Figures 12, 13)**.

Two exposed faults were mapped in the area **(Figure 12)**. The N5°W-trending Wolf Creek Hot Spring fault cut Pinedale outwash and dropped the land surface 1 to 1.5 meters on the east side, as shown by the fault scarp. The second fault, the Sun Ranch fault, has a W-NW trend along Wolf Creek and has displaced the regionally extensive Huckleberry Ridge ash-flow tuff downward on the north side. This fault lacks a scarp but is evidenced by the disappearance of the tuff, which is not found in a 30-meter-high hill about 200 meters north of the creek. Also, the fault plane is visible along the west bank of the Madison River where it dips 50 to 70°N. An outcrop on the down-dropped side of the fault on the east bank of the Madison River near the Palisades campground can be projected under the outwash 2

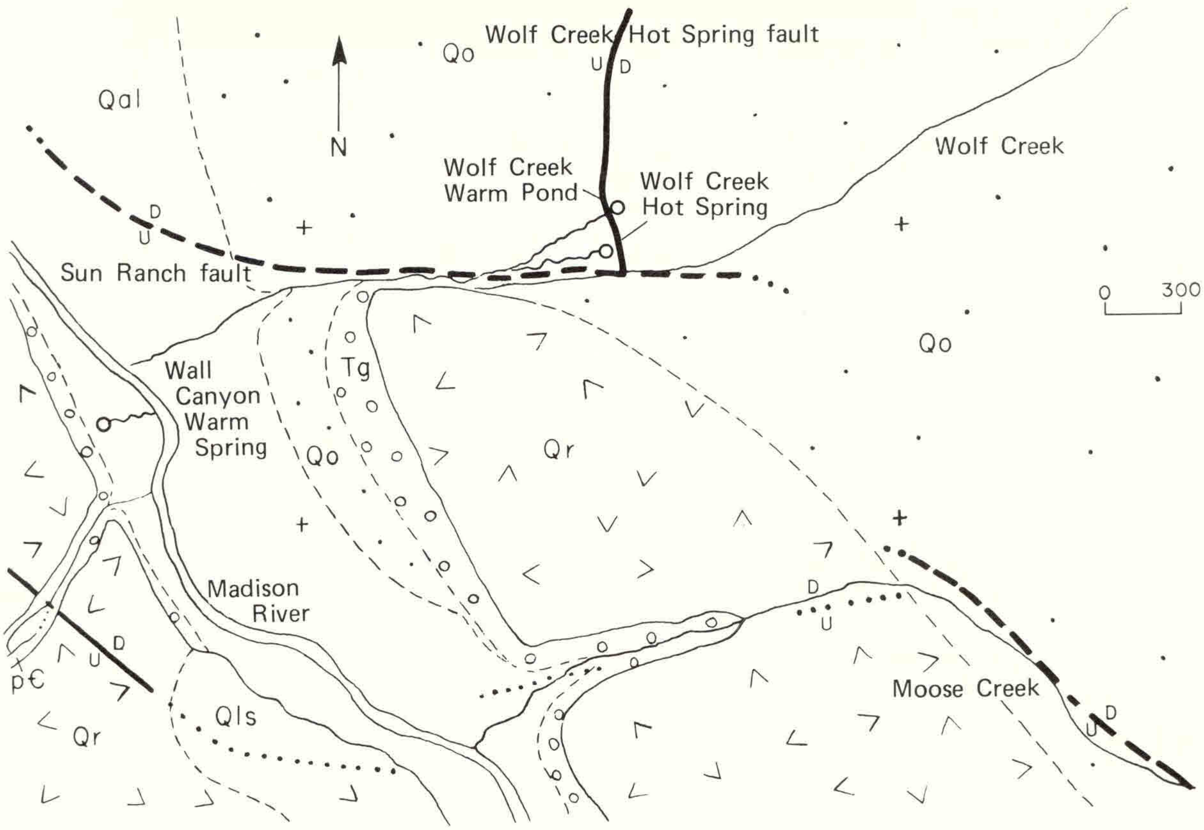


Figure 12—Wolf Creek Hot Spring geologic map sketched in the field; sections 4, 5, 8 and 9, T. 10 S., R. 1 E. Map units are (pЄ), Precambrian metamorphic rocks; (Tg), Tertiary gravel; (Qr), Huckleberry Ridge Tuff; (Qo), Quaternary glacial outwash; (Qls), landslide; and (Qal), alluvium.

kilometers southeast to the hot-spring area, where the projection indicates that the rhyolite tuff should be at a depth of 100 to 150 meters.

The resistivity-survey lateral profiles at 10 and 20 meters apparent depths are shown in **Figures 14 and 15**. The lowest resistivity, slightly under 100 ohm-m at both depths, is located about 30 meters southwest of the hot spring; another resistivity low lies along the Wolf Creek Hot Spring fault beneath the warm pond at a depth of 10 meters **(Figure 14)**. This low broadens and trends northwest at a depth of 20 meters **(Figure 15)**. Results suggest that thermal water rises along this north-south fault from fractured and faulted Precambrian gneissic bedrock on the floor of the Madison Valley graben, and upward consecutively through pre-rhyolite gravel, fractured Huckleberry Ridge Tuff and approximately 100 meters of permeable Wisconsin (Pinedale) glacial outwash. When the water reaches the permeable glacial outwash, which is composed of Precambrian cobbles and gravel from the Madison Range, much of it drains westward through the outwash toward the Madison River. This produces the westward extension of the resistivity low along the Wolf Creek Hot Spring fault near the Wolf Creek Warm Spring shown in **Figures 14 and 15** (Chadwick and others, 1977). Only a small fraction of the thermal water may reach the surface. According to White and others (1971), thermal springs that emerge on an alluvial surface above the water table have a major portion of the flow within the alluvium—provided the spring lacks a sealed calcite- or silica-cemented conduit for the thermal water through the alluvium. At Wolf Creek, frequent earthquakes may break the cemented conduit and allow cold water contamination of the thermal spring and leakage of thermal water through the conduit into the undersaturated surface alluvium.

Microseismic activity in the Wolf Creek-Wall Canyon area was monitored by the University of Utah in 1976. A concentration of microearthquakes was located about 8 kilometers northwest of the Wall Canyon Warm Spring, and several microearthquakes were recorded in the vicinity of the Wolf Creek Hot Spring. The highest activity was located, however, in the Missouri Flats basin 20 kilometers to the south. This basin lies along a southwest line representing a seismically active belt extending from the West Yellowstone basin into the Centennial Valley (Bailey, 1977).

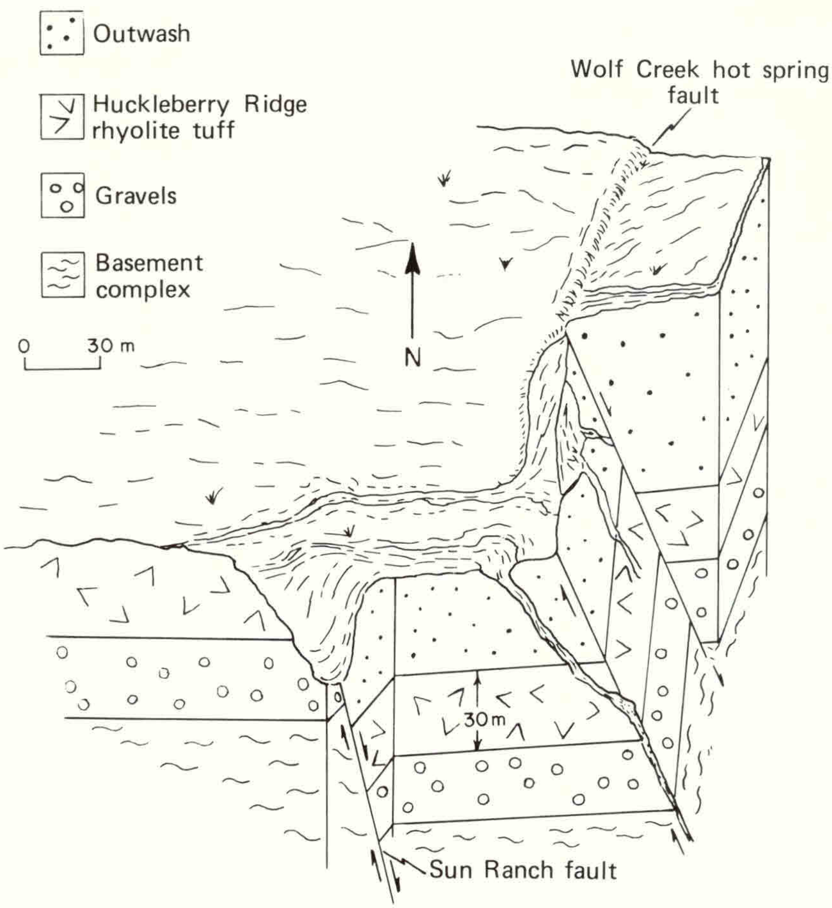


Figure 13—Wolf Creek Hot Spring block diagram showing the proposed intersection of the W-bearing Sun Ranch fault and the N-trending Wolf Creek Hot Spring fault. The intersection provides thermal water access to the surface (vertical scale diagrammatic).

Wall Canyon Warm Spring

Wall Canyon Warm Spring lies 0.5 kilometer north of Wall Canyon on the alluvial plain of the Madison River (NW¼ sec. 7, T. 10 S., R. 1 E.).

The spring emerges from the base of a cliff of Huckleberry Ridge Tuff and flows onto a flood plain along the west bank of the Madison River **(Figure 16)**. The spring feeds a 3- to 7-hectare marsh; therefore, the total flow of the spring is difficult to measure. A 1-meter-square test pit previously dug near the base of the cliff was used to measure the spring, and here the water is 25°C and flows from the pit at a visually estimated rate of 8 L/min.

Scattered Precambrian gneiss outcrops are found from about 1 kilometer westward in Wall Canyon and continue east to a point near the top of the bench where gneiss and tuff are nearly in contact. A thin, pre-Pleistocene gravel is mantled by a 1- to 2-meter-thick sand and ash layer emplaced prior to the tuff or possibly during the early stages of the eruption.

A major cross-valley, NW-trending fault (Wall Canyon fault) passes through the rhyolite 2 kilometers west of the cliff and has dropped the tuff about 30 meters on the northeast. The fault trace is seen in Precambrian gneiss bedrock and continues northwestward from Wall Canyon into the Gravelly Range, where the displacement is greater than the 30 meters representing the most recent Pleistocene movement (Hadley, 1969a).

A thermal-water circulation model proposed here for the Wall Canyon Warm Spring uses the Wall Canyon fault to circulate water to a sufficient depth for heating. The heated water rises up the NE-dipping fault and at some point leaves the confining gneiss far enough north of Wall Canyon so that the water does not flow into the canyon but flows upward through the alluvial gravel layer until dammed by the rhyolite tuff. There, it flows laterally about 1 to 2 kilometers beneath the tuff to the edge of the tuff outcrop (cliff face) where it emerges as the warm spring **(Figure 16)**. During its course, it may mix with

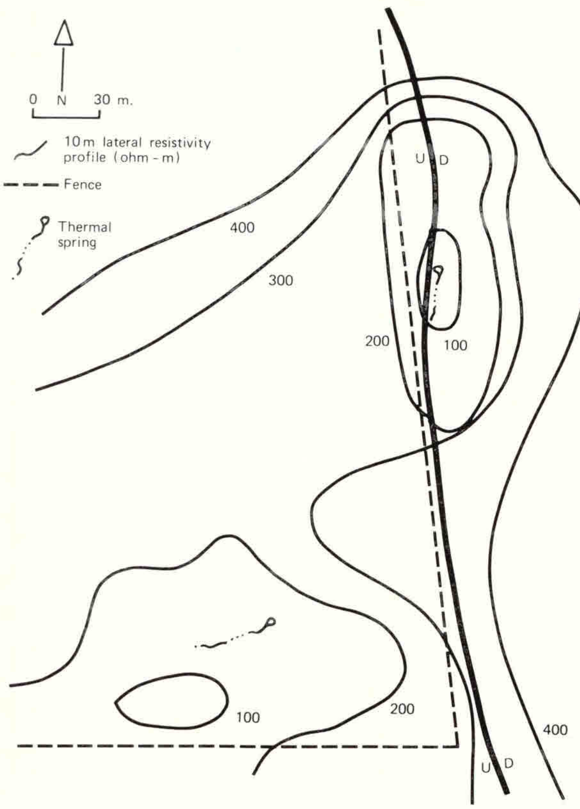


Figure 14—10-m depth lateral resistivity profile showing low resistivity values along the Wolf Creek Hot Spring fault in the vicinity of the warm pond and near the main spring.

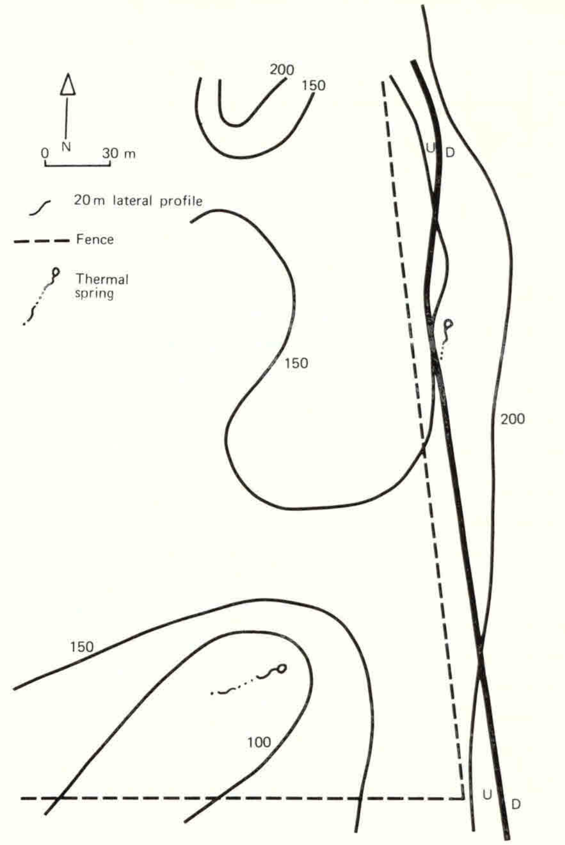


Figure 15—20-m depth lateral resistivity profile (in Ohm-m) showing low resistivity values along the Wolf Creek Hot Spring fault and near the main spring.

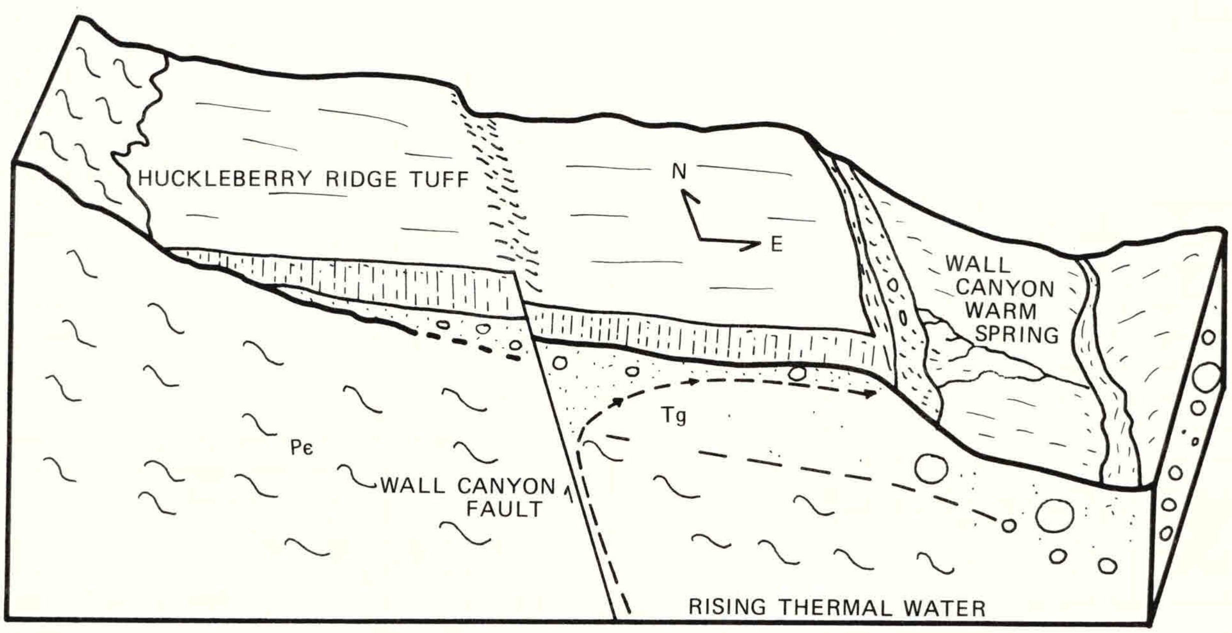


Figure 16—Diagrammatic representation of possible fault control on the Wall Canyon Warm Spring thermal water. Legend: (pЄ), Precambrian metamorphic rock; (Tg), Tertiary gravel.

cold ground water and pick up dissolved solids from the alluvium and ash deposits.

Curlew Creek Warm Spring

Curlew Creek Warm Spring emerges from a marsh at the junction of the north and south forks of Curlew Creek (SE¼ sec. 13, T. 11 S., R. 1 E.), 1 kilometer northeast of the Madison River.

Spring flow was visually estimated at 2 L/sec. The spring lies in the center of a triangular marsh that separated the 10°C water of the two branches of Curlew Creek **(Figure 17)**. The warmest temperature measured at the spring was 23°C in July 1976; in February 1977, the temperature was 21°C. Cold surface waters and ground water may infiltrate through the marsh and contaminate the thermal spring.

At Curlew Creek Warm Spring, the Huckleberry Ridge Tuff overlies pre-Pleistocene alluvial gravel, and Precambrian bedrock lies at an unknown depth beneath the gravel. The warm spring issues from the alluvium at the bottom of the steep-sided, narrow Curlew Creek valley. It issues about 15 meters below the base of the Huckleberry Ridge Tuff, probably from the pre-Pleistocene gravel unit. It is inferred that a NW-trending fault which cut the rhyolite **(Figure 17)** would have provided a zone of weakness, allowing Curlew Creek to cut its channel and remove the tuff from the fault zone.

Two principal lines of evidence suggest the presence of a fault along Curlew Creek: (1) The steplike nature of the Huckleberry Ridge Tuff outcrop suggests that there are pre-tuff, Tertiary-age faults that extend to about 3 kilometers northward from the north edge of Missouri Flats; these faults can be correlated with several southwest-bearing stream valleys and draws. (2) A 3- to 5-meter N-NE-trending straight scarp offsets Quaternary alluvium along the south side of Curlew Creek. The W-dipping scarp is visible from 0.2 kilometer downstream from the warm spring to 0.1 kilometer upstream along the North Fork of Curlew Creek. At this point, a recent scarp is superimposed on the older scarp; the recent scarp is inferred to be related to fault movement or slumping during the 1959 earthquake. If a fault exists along Curlew Creek, it could act as a conduit for circulation of ground water within the pre-Pleistocene gravel and/or the Precambrian bedrock.

Sloan Cow Camp Warm Spring

Sloan Cow Camp Warm Spring is located about 3 kilometers southwest of the junction of the West Fork of the Madison River and Elk River (SW¼ sec. 19, T. 12 S., R. 1 E.).

The 30°C spring flows from an orifice 1 meter in diameter, at approximately 1 cubic foot per second.

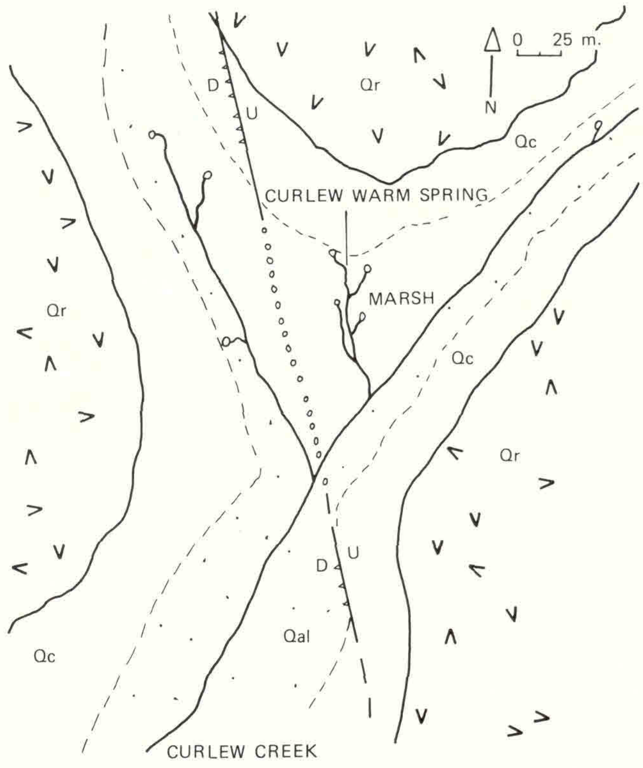


Figure 17—Geologic map sketched in the field showing Curlew Creek Warm Spring area, section 13, T. 12 S., R. 1 E.; (Qr), Huckleberry Ridge Tuff; (Qc), Quaternary colluvium; (Qal), Quaternary alluvium. Sawtooth line indicates recent scarps.

(Note: A Montana Bureau of Mines and Geology flow measurement for the combined warm and cold springs gives an approximate 350-gpm value for the warm spring.) A 10°C cold spring, located about 5 meters southeast of the warm spring, flows directly into the warm spring discharge water 1 meter beyond the orifice. Spring temperatures were measured several times during the summer and fall of 1976 and 1977. The warm spring varied from 28.5 to 31.5°C, and the cold spring from 8 to 10°C. No seasonal relationship was noted in the temperature fluctuations.

Quaternary alluvium and colluvium, and volcanic units in the valley bottom, overlie older alluvium; these units rest on Precambrian gneiss. The 2-kilometer-wide valley was produced by the ancestral West Fork drainage, which appears to have drained the Centennial Valley, and which is now occupied by the underfit West Fork. The valley walls consist of Huckleberry Ridge Tuff to the south and Precambrian gneiss of the Gravelly Range to the north.

The conduit allowing deep circulation of the thermal water may be the postulated range-front fault at the southeastern border of the Gravelly Range.

West Fork Swimming Hole Warm Spring

West Fork Swimming Hole Warm Spring is located 1 kilometer west-southwest of the junction of the West Fork of the Madison River and Elk River in the valley of the West Fork of the Madison River (SE¼ sec. 18, T. 12 S., R. 1 E.).

The 28°C spring forms a pool approximately 10 meters in diameter and 3 meters deep. It emerges from an alluvial slump block 25 meters higher and about 50 meters north of the West Fork **(Figure 18)**. A flow visually estimated at 1.5 cubic feet per second cascades into the West Fork. (Note: A Montana Bureau of Mines and Geology flow measurement was 480 gpm.)

A model for spring circulation uses thermal waters that have circulated along the inferred Gravelly Range front fault, which is postulated to extend northward from Sloan Cow Camp Warm Spring. This fault would allow circulation and heating of meteoric water within the Precambrian bedrock. The rising thermal water would mix with cold, shallow ground water penetrating the overlying alluvium.

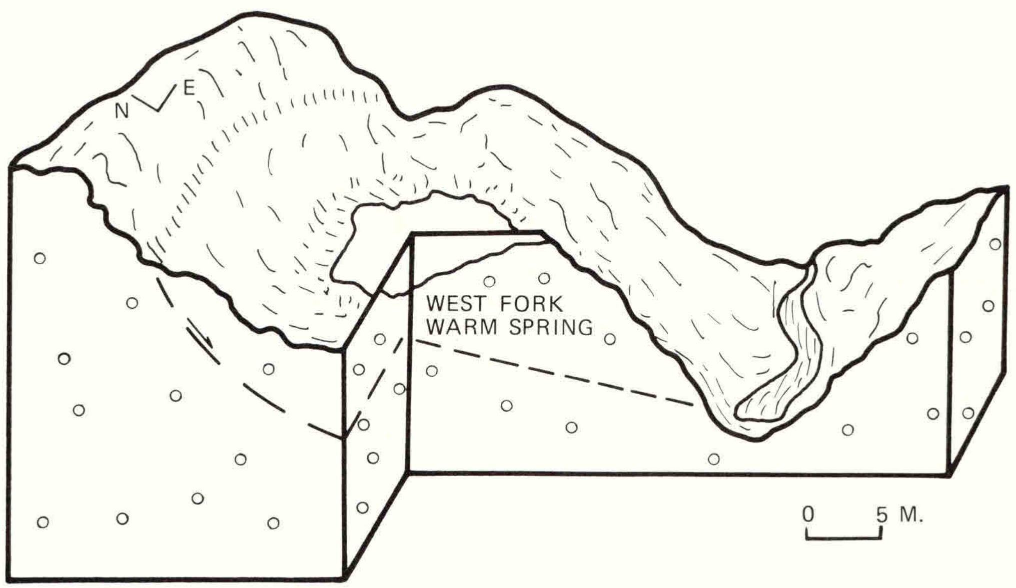


Figure 18—Diagrammatic representation of the West Fork Warm Spring emerging from alluvial slump block near West Fork River.

Summary

Study results for the upper Centennial and Madison valleys can be summarized as follows:

1. The geothermal resource in the eastern Centennial Valley is of low temperature, suitable mainly for direct-use applications.

2. The geothermal resources along the West Fork and Madison River are warmer, but still in the low-temperature category. The Wolf Creek Hot Spring area may have a considerably larger subsurface discharge component; it constitutes the best exploration target for water warmer than 50°C.

3. Ground-water resources within the upper Centennial Valley are substantial. Water-level data suggest that vertical movement of water is downward; consequently, flowing artesian wells probably cannot be developed in the deeper Quaternary sands.

4. The Centennial fault system is more complex than was previously believed.

5. The upper Centennial Valley drained to the northeast, into the paleo-Madison River, during late Cenozoic time.

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Appendices

APPENDIX A
Precambrian rocks

Mica schist

The largest area of mica schist is in the northeastern corner of the Upper Red Rock Lake quadrangle. Exposures of schist continue north of this quadrangle to the low hill called The Horn. Mica schist weathers dark brown to gray, typically forming slabby outcrops. Quartz, feldspar and biotite are megascopically recognizable in schist in the northwestern part of this large area underlain by schist. Across strike to the southeast, grain size decreases, and flexural slip folds with wavelength and amplitude of approximately 1 cm are abundant. Layers of quartzite and quartzitic schist occur within the schist north of the Upper Red Rock Lake quadrangle. To the north of Antelope basin, a layer of mica schist within the marble is folded into an antiform with a NE-plunging axis.

Quartz and feldspar porphyroblasts generally range from 0.2 to 0.8 mm in size and are surrounded by a fine-grained matrix of quartz, feldspar and biotite grains ranging in size from 0.02 to 0.05 mm. Foliation is caused by the alignment of biotite and also by the concentration of biotite in thin layers (slip surfaces). These biotite layers wrap around the quartz and feldspar porphyroblasts and are folded into open folds with 1- to 3-mm amplitude and 1-cm wavelength. In the more intensely deformed schist, folds with an amplitude-to-wavelength ratio of 1:1 are abundant and some kink bands occur in the biotite. Syntectonic recrystallization of some biotite is indicated by the "piling up" of biotite flakes like shingles at the hinge lines of folds.

Quartz is the most abundant constituent of the biotite-feldspar-quartz schist. Albite and K-feldspar are the other major constituents. Pale yellowish green to dark olive drab biotite with pleochroism is a minor constituent. Epidote and clinozoisite are associated with some of the biotite. Minor constituents are magnetite (octahedrons 0.4 mm across), muscovite, apatite, zircon and hematite.

With increasing quartz and muscovite concentration, the biotite-quartz-feldspar schist grades into micaceous quartzite, which, with further increase in quartz content, grades into quartzite. The micaceous quartzite is a minor rock type that occurs as thin layers in both the quartzite and biotite-quartz-feldspar schist.

Quartzite

Although large areas are underlain by quartzite, as indicated by float, this rock type is not well exposed. Exposures of quartzite are typically low outcrops, and the color on weathered surfaces ranges from light tan to brown and sometimes reddish brown.

Typically, quartz grains exhibiting undulose extinction and ranging in size from 0.2 to 1 mm are surrounded by a fine-grained (0.02 to 0.05 mm) mixture of quartz with lesser muscovite and rarely chlorite. The larger quartz grains are aligned in those specimens where they are elongate. Protomylonite is exposed in the SW¼ SW¼ sec. 25, T. 13 S., R. 1 E., and in the SE¼ sec. 19, T. 13 S., R. 2 E. In outcrop, this rock is recognized by its glassy appearance, bluish cast and a faint lineation. In thin section, the protomylonite is recognized to be spectacular ribbon mylonite with ribbons of quartz ranging in thickness from 0.06 to 0.4 mm and up to 6 mm long.

Quartz accounts for more than 90 percent of the quartzite with the remainder muscovite and chlorite. Biotite is a minor constituent in one specimen from the south flank of Deer Mountain. Trace constituents are K-feldspar, tourmaline, rutile, pyrite (partly altered to hematite and limonite), and small grains several μm across of an unidentified opaque mineral.

Dolomitic marble

The dolomitic marble forms blocky outcrops and weathers tan, gray or, less commonly, reddish brown. On a fresh surface, this fine-grained marble is light gray, and at many localities has a flinty appearance and breaks with a conchoidal fracture. Abundant quartz layers from 1 mm to 30 cm in thickness stand out in relief on a weathered surface. The rustic and irregular surface makes this rock desirable as decorative stone, and loose blocks are obtained for this purpose from Saddle Mountain. The quartz layers also make it possible to recognize folds in the marble. Folds range in geometry from isoclinal (limb separation less than 10°) to tight folds (limb separation from 10 to 30°), and from 1 cm to 2 meters in amplitude.

The major constituent of the marble is dolomite, which usually exhibits a mosaic texture with individual grains between 0.03 and 0.08 mm in size. Irregular patches and veinlets of coarse-grained dolomite with grains up to 1 mm across occur in some of the marble. Quartz layers are composed of equidimensional grains usually the same size as the surrounding dolomite grains. Some specimens contain quartz scattered throughout the dolomite. The quartz content of this rock is estimated to range from 5 to 50 percent. Small phlogopite grains 0.06 mm across are a trace constituent of much of the marble. Veinlets 1 mm thick and consisting of a mixture of talc and chlorite (identified by x-ray diffraction analysis) occur rarely in the marble. Contact metamorphism adjacent to a metagabbro dike has produced the assemblage tremolite, serpentine, calcite and dolomite.

At another locality talc, calcite, quartz and minor dolomite occur next to a metagabbro dike. In the SE¼ SE¼ sec. 6, T. 14 S., R. 2 E., chrysotile veinlets 1 cm thick are exposed in marble adjacent to metagabbro.

Rare layers of calcitic marble occur in the dolomitic marble and also in the amphibolite.

Amphibolite

The amphibolite shows considerable variation in the development of foliation from a schist to a massive rock. It weathers green to greenish brown and is generally not as well exposed as the other Precambrian units. Scattered exposures of metagabbro and metadiabase occur within the area mapped as amphibolite; only the larger or more continuous of these bodies are shown on the geologic map. A short distance southeast of Saddle Mountain, a concordant layer of dolomitic marble approximately 30 m thick occurs within the amphibolite. Layers of micaceous quartzite and graphitic schist up to tens of meters thick are surrounded by amphibolite in the vicinity of Elk and Deer mountains.

The massive amphibolite is fine grained, with an average grain size of 0.1 mm. Where this rock is foliated, actinolite and, less commonly, albite porphyroblasts up to 2.5 mm in size are surrounded by a fine-grained (0.1 mm) matrix of quartz and epidote. Actinolite is the dominant mineral in most of the amphibolite, accounting for 50 to 90 percent of the rock. It is pleochroic from pale yellowish green to pale bluish green. Some grains are rimmed by bluish-green hornblende. Tremolite instead of actinolite is the amphibole in two specimens. Albite, epidote, quartz, chlorite and magnetite are generally minor constituents, although one specimen is estimated to contain 30 percent epidote and 30 percent chlorite. The chlorite, which is a major constituent of the fine-grained groundmass, shows blue to purple interference colors. Apatite and sphene are generally trace constituents, although one specimen contains 3 percent sphene. Biotite, reddish-brown in this section, is a minor constituent of some specimens.

Graphitic schist

Graphitic schist was found only in scattered outcrops surrounded by amphibolite in the SW¼ sec. 35, T. 13 S., R. 1 E. It is a distinctive unit that is very fine grained, thinly foliated and shiny black in outcrop. Average grain size of this schist is 0.04 mm, and it is characterized by small flexural slip folds with wavelength and amplitude of 0.2 mm. Quartz, chlorite and muscovite are the major constituents of this schist. Rutile and graphite are minor constituents.

Metagabbro

The numerous metagabbro bodies in this area range in size from dikes and sills a few meters thick to the large, generally concordant bodies south of Antelope basin. Although generally concordant, the metagabbro body northwest of Saddle Mountain is discordant to the surrounding amphibolite at its southwestern end. This discordance can be seen locally in outcrop, as well as from the configuration of the contact shown on the geologic map.

The metagabbro tends to be well exposed and is brown to greenish gray on a weathered surface. Grain size ranges from 0.3 to 3 mm. Much of the finer-grained metagabbro exhibits a subophitic to ophitic texture, which is preserved in some specimens where actinolite has entirely replaced the pyroxene. Although not recognizable on a fresh surface, the ophitic texture is easily recognizable on a weathered surface. Within the larger of the two metagabbro bodies south of Antelope basin, there is a small exposure of metapyroxenite in which the pyroxene has been entirely replaced by tremolite. In the same metagabbro body in the NE¼ NW¼ sec. 29, T. 13 S., R. 2 E., an outcrop of anorthosite, in which the plagioclase has been completely replaced by fine-grained epidote or clinozoisite, is exposed.

The igneous minerals partly preserved in some of these gabbroic rocks are mainly sodic labradorite and clinopyroxene. The average composition of this plagioclase is estimated to be An_{55}, but in one specimen is An_{78}. Both normal and oscillatory zoning was observed. The estimated mode of the metagabbro is 35 to 50 percent plagioclase, with the remainder of the rock mainly clinopyroxene (probably augite). Most specimens contain several percent each of quartz, magnetite and micrographic intergrowths of quartz and feldspar (presumably K-feldspar). Biotite is a minor constituent in some specimens.

All of the 18 specimens of gabbroic rocks examined in thin section show the effects of greenschist-facies metamorphism to varying degree. In most specimens, the plagioclase is unaltered, but the pyroxene has been replaced by actinolite or, less commonly, by tremolite or hornblende. Actinolite is pleochroic from pale yellowish green to light green, and the rare hornblende that occurs as a minor constituent in some of the actinolite-bearing specimens is pleochroic from pale brown to dark greenish brown. Some actinolite grains are pleochroic from pale yellowish green to bright green with a faint bluish cast. Clinozoisite, penninite and sphene are minor constituents in the partly altered metagabbro. The metagabbro exposed between Deer Mountain and Alaska Basin, although not foliated, is extensively altered. Most of the pyroxene has been replaced by actinolite-tremolite, and almost all of the plagioclase has been replaced by clinozoisite or epidote with minor, fine-grained, white mica.

Scapolite veinlets a few millimeters thick occur in intensely altered metagabbro exposed on the north side of Elk Mountain in the NE¼ sec. 33, T. 13 S., R. 1 E., and also in metagabbro exposed in a small quarry on the east side of The Horn, 3.2 km northeast from the northeast corner of the Upper Red Rock Lake quadrangle. Scapolite from The Horn locality has a refractive index of 1.552, indicating that it is in the dipyre range with an estimated composition of Me_{35}. The major constituents of this metagabbro are actinolite and clinozoisite accompanied by minor albite, muscovite, chlorite and magnetite.

APPENDIX B
Cenozoic volcanic rocks

The Cenozoic volcanic rocks of the upper Centennial Valley include localized basalts of both Tertiary and Pleistocene age, as well as Pleistocene welded tuffs that cover extensive portions of the north flank of the valley. **Figure B-1** shows the location of the measured sections described below.

Tertiary basalt

This is a dark-gray to black, dense, fine-grained basalt. Locally vesicular, it contains abundant magnetite with sparse olivine phenocrysts. It is exposed primarily as thin flows which commonly display columnar jointing. This rock is petrographically similar to the basalt near Elk Lake; it has a K-Ar age of 6.6 m.y., compared to a date of 1.95 m.y. for the Elk Lake basalt. It is possible that both units may be comagmatic.

Pleistocene basalt

This is a dark-gray to black, dense basalt. It contains sparse olivine phenocrysts and is locally vesicular. A whole-rock K-Ar date gives 1.95 ± 0.35 m.y. This unit is exposed only at the south end of Elk Lake where its depositional relationship with that of the Huckleberry Ridge Tuff is questionable. The basalt occurs as boulder piles and is not exposed where definitely in place. The unit appears either to have cut through and flowed over the Huckleberry Ridge Tuff or to have cut through and flowed over the first cooling unit of the tuff. The basalt appears to have risen from a magma source, possibly related to the Snake River Plain-Yellowstone volcanic system, via the deep Cliff Lake fault. A similar basalt, dated at slightly less than 2.4 m.y., is known to have originated from this volcanic system in Yellowstone National Park, 40 miles (64 km) to the east of Elk Lake (Christiansen and Blank, 1972).

Pleistocene Huckleberry Ridge Tuff

The Huckleberry Ridge Tuff, which originated from the Island Park caldera, is the oldest (2.0 m.y.) of the three units that comprise the Yellowstone Group volcanics of Christiansen and Blank (1972).

The tuff found in the upper Centennial Valley is petrographically similar to that of the type section of the Huckleberry Ridge Tuff, where it has been subdivided into three members, each a cooling unit of a compound ash-flow sheet (Christiansen and Blank, 1972). Two members are present in the upper Centennial Valley. The lower cooling unit in the Centennial Valley can be correlated with the lower member described at the type section, and the upper cooling unit with the middle member. The lower member at

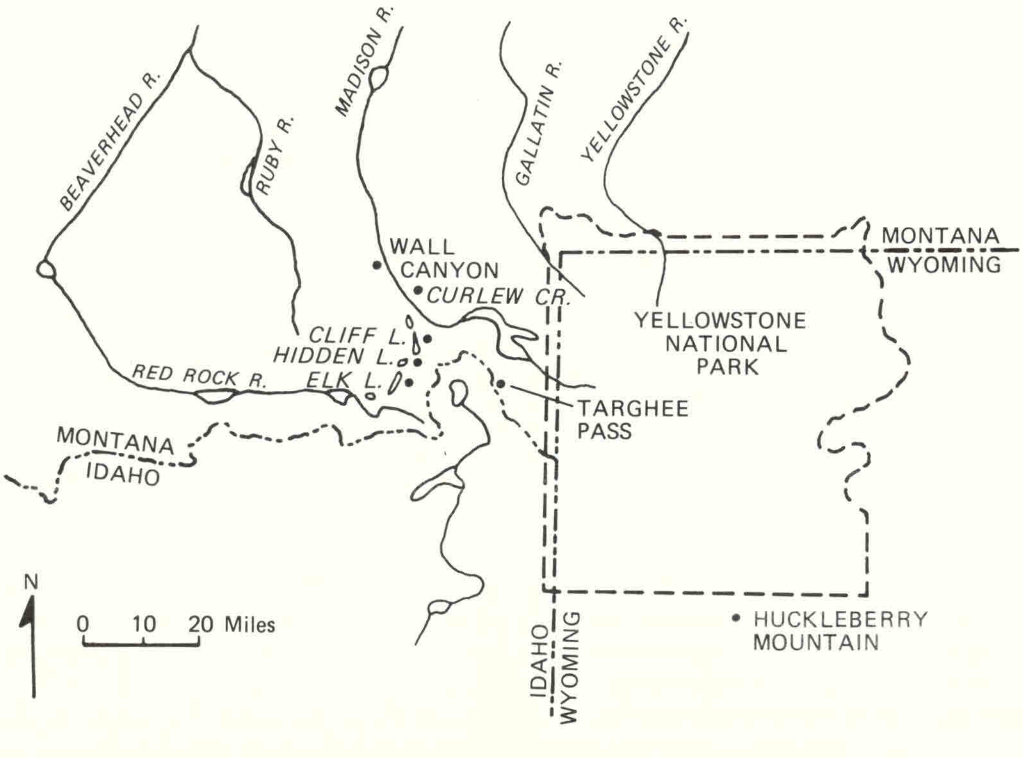

Figure B-1—Measured section location map. (See Figure B-4.)

the type section contains a thick zone of dense welding beneath a zone of partial welding. Phenocrysts decrease in number toward the top of this member. The lower cooling unit of the tuff in the Centennial area contains a smaller percentage of phenocrysts overall but a similar pattern of welding and phenocryst distribution. The upper cooling unit is similar to the middle member of the type section, with a phenocryst-poor zone near the base of the unit. Phenocrysts increase in abundance and size toward the top of the middle member of the tuff at both locations. The upper member of the tuff described at the type section occurs only south of central Yellowstone National Park (Christiansen and Blank, 1972).

K-Ar dates of approximately 2.0 m.y. have been obtained from the tuff at the following locations: (a) 1 mile north-northwest of Hidden Lake; (b) the south end of Elk Lake; (c) the cliff face on the west bank of the Madison River, 2 km north of the mouth of Wall Canyon; and (d) the northeast slope of Flatiron Mountain.

The following description represents the Huckleberry Ridge Tuff sequence located near Hidden Lake **(Figures B-2, B-3)**. A chill zone is present as float at the base of the outcrop. The tuff of the chill zone is vitric, partially welded and contains 3 to 5 percent sanidine, with trace amounts of quartz and olivine phenocrysts, all of which range up to 3 mm in diameter. In these pyroclastic rocks of rhyolitic composition, the olivine phenocrysts appear to be the result of the partial mixing of the bimodal magmatic system beneath the Yellowstone region at the time of eruption. The bimodal system consists of basaltic material beneath magmas of rhyolitic composition (Struhsacker, 1978). Above the basal chill zone is a zone of dense welding 75 meters thick, in which devitrified shards are extremely flattened and deformed. This portion of the tuff contains 3 to 5 percent sanidine phenocrysts at the base of the zone, which decrease to 1 to 2 percent near the top. Phenocryst size also varies vertically from an average diameter of 1.5 mm near the base to 0.5 mm at the top. The zone of dense welding grades upward into a zone of partial welding that is approximately 25 meters thick. The zone of partial welding contains 1 to 2 percent sanidine phenocrysts averaging 1 mm across in a groundmass of devitrified shards that are slightly deformed. Also present are trace amounts of quartz and olivine phenocrysts. This zone of partial welding coincides with the vapor phase of the ash. Vesicles filled with minute feldspar crystals, that readily weather to clay when exposed to surface conditions, are associated with the vapor phase of this cooling unit.

Immediately above the zone of partial welding is a compacted ash bed ½ meter thick. The ash bed separates the lower cooling unit from the upper unit. This vitric ash contains undeformed glass shards together with 2 percent sanidine and 1 percent quartz phenocrysts averaging 3 mm across. The presence of quartz phenocrysts indicates that the ash bed is related to the second eruptive pulse of the Huckleberry Ridge Tuff. Quartz phenocrysts are rare in the lower member but are relatively abundant in the second member of the tuff. The fact that the ash is unaltered and has not been removed by erosion is further evidence for association with the second member of the tuff.

Overlying the ash bed is the vitric chill zone of the upper cooling unit. The glass shards of this zone are moderately deformed and partially welded. This chill zone contains 2 to 3 percent sanidine phenocrysts up to 2 mm in diameter as well as trace amounts of quartz and olivine phenocrysts. The basal chill zone grades vertically into a zone of partial welding 9 meters thick, in which the shards are devitrified and moderately deformed. This portion of the tuff contains 10 to 15 percent sanidine with 1 percent quartz phenocrysts, all ranging up to 1.5 mm in diameter. The zone of partial welding grades upward into a zone of dense welding that has undergone devitrification. This zone is 27 meters thick and contains 10 percent sanidine phenocrysts at the base, increasing vertically to 25 percent near the top. The phenocrysts of this zone range up to 4 mm in diameter. Quartz and olivine phenocrysts averaging 1 percent of the rock are also present in this zone. The zone of dense welding grades upward into a zone of partial welding that is devitrified and contains 15 to 30 percent sanidine, 5 percent quartz and 1 percent olivine phenocrysts. The phenocrysts of the zone of partial welding average 3 mm in diameter. The upper portion of the zone has probably been removed by erosion, and the remaining 5 meters of the zone are badly weathered.

The upper cooling unit of the Huckleberry Ridge Tuff as seen at this location is thinner (42 m) than the lower cooling unit (100 m). The upper unit has a zone of dense welding, which is thinner than that of the lower cooling unit, sandwiched between two zones of partial welding **(Figure B-3)**. The lower unit has a very thick zone of dense welding with only a thin, partially welded zone beneath (chill zone). This indicates a much higher emplacement temperature for the lower cooling unit.

The timing of the Huckleberry Ridge Tuff emplacement into the upper Centennial Valley was such that the lower member was allowed to cool as a single, simple cooling unit before being covered by the second member. The emplacement of the second unit followed soon after the first episode, as shown by the lack of extensive erosion of the lower

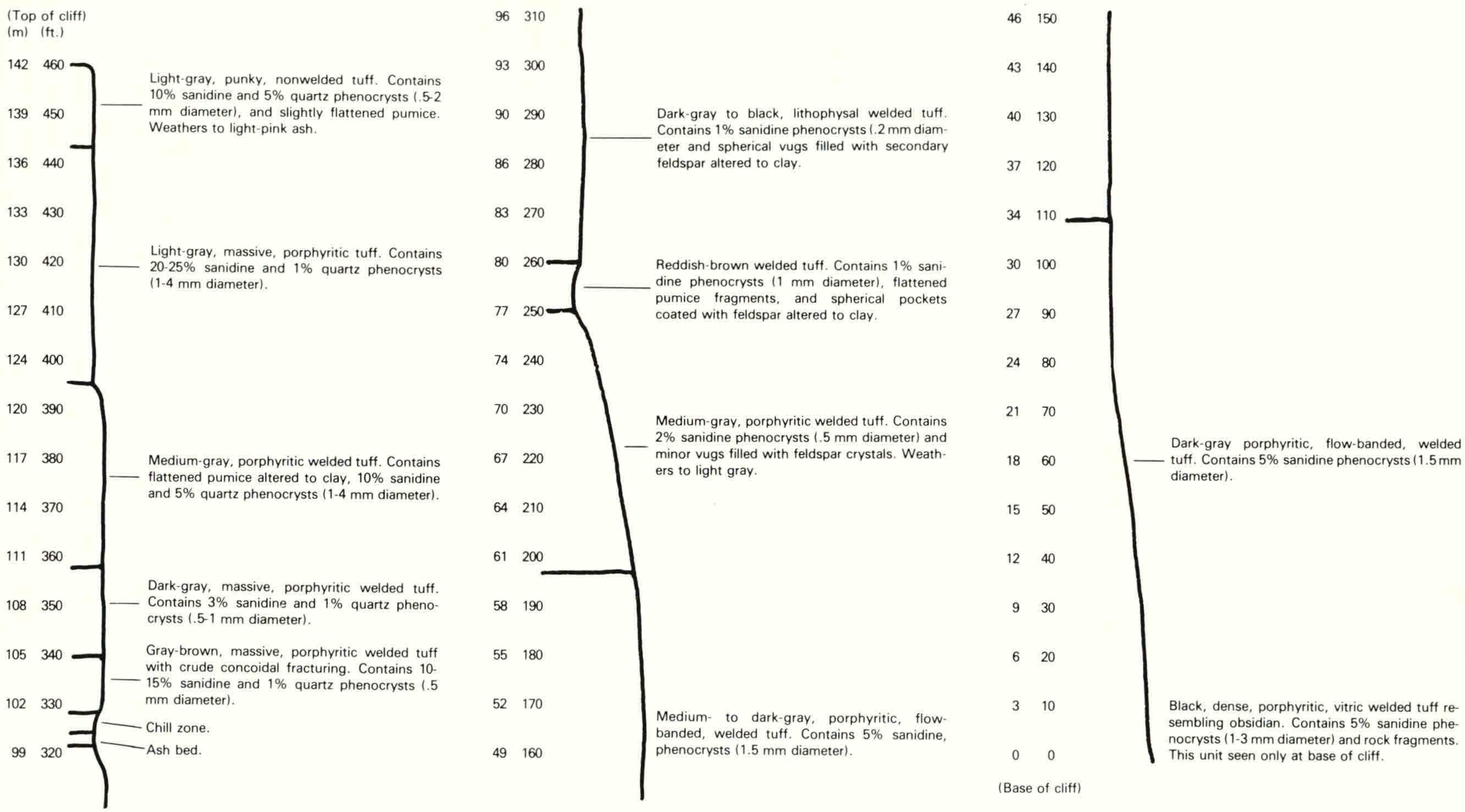

Figure B-2—Huckleberry Ridge Tuff measured section located 1 mile NNW of Hidden Lake—NW¼ NE¼ sec. 4, T. 13 S., R. 1 E. (*from* Mannick, 1980).

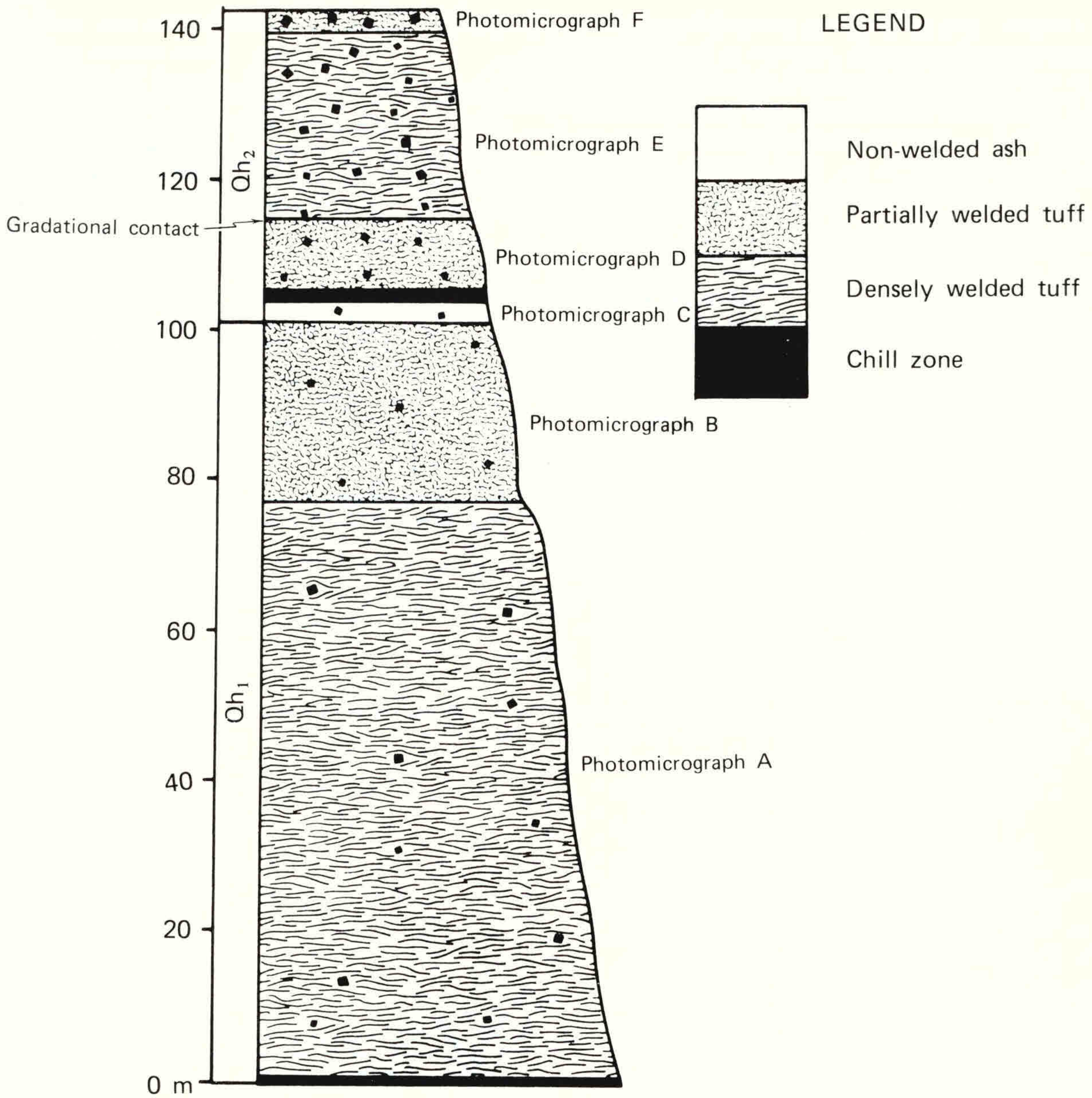

Figure B-3—Zonal variations in welding density of the Huckleberry Ridge Tuff—NW¼ NE¼ sec. 4, T. 13 S., R. 1 E. (*from* Mannick, 1980).

member surface. Potassium-argon dates of the upper and lower cooling units are identical within the resolution of the dating method.

The dimensions of the welding zones vary with depositional thickness of the tuff and distance from the source. The thickness of the Huckleberry Ridge Tuff can vary greatly over a short distance when deposited over uneven paleotopography. The stratigraphic correlations at the type section (Christiansen and Blank, 1972) and at Elk Lake, Hidden Lake, Cliff Lake, Curlew Creek and Wall Canyon display changes in the dimensions of the individual cooling units and welding zones within them **(Figure B-4)**.

The Huckleberry Ridge Tuff can be traced discontinuously from the caldera region into the Centennial area and northward into the upper Madison Valley, where the tuff thins and eventually terminates **(Figures B-4, B-5)**. The Huckleberry Ridge Tuff has been measured at over 300 meters in thickness along the western side of Yellowstone National Park (Christiansen and Blank, 1972) and is present at Targhee Pass, Idaho, in thicknesses up to 200 meters. The tuff is 123 meters thick to the west at Elk Lake and decreases in thickness northward to 91 meters at Hidden Lake, 86 meters at Cliff Lake (Weinheimer, 1979), 48 meters at Wall Canyon (Weinheimer,

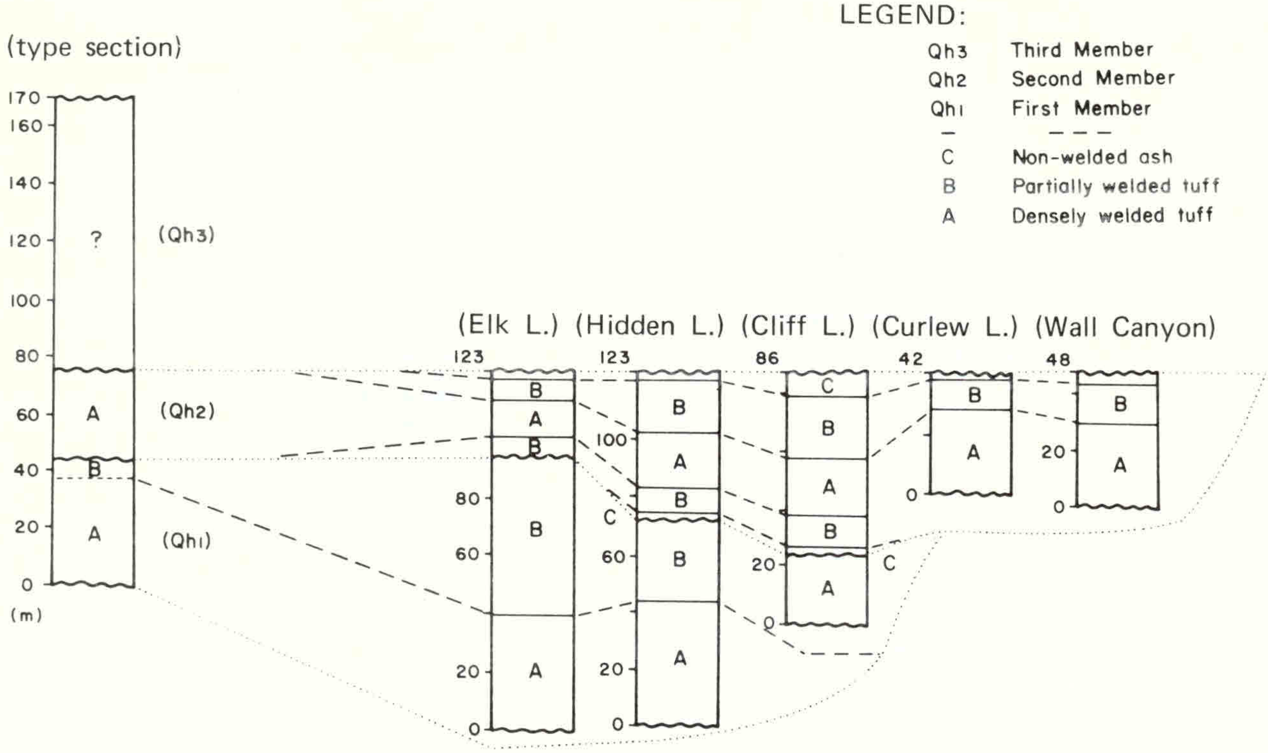

Figure B-4—Stratigraphic correlation of the Huckleberry Ridge Tuff. Measured sections are located in Figure B-1. Horizontal not to scale.

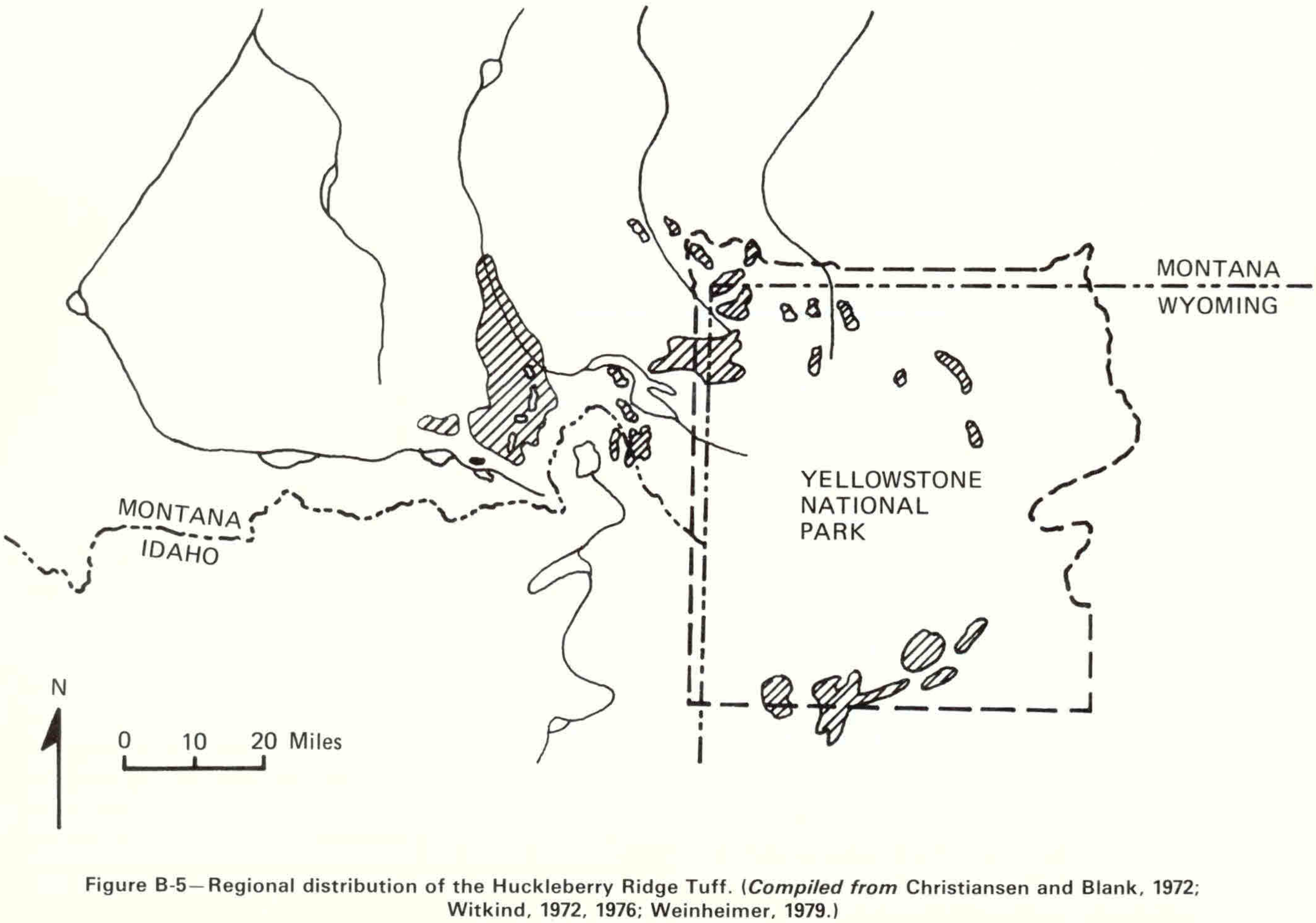

Figure B-5—Regional distribution of the Huckleberry Ridge Tuff. (*Compiled from* Christiansen and Blank, 1972; Witkind, 1972, 1976; Weinheimer, 1979.)

1979), and thins to zero near Ruby Creek in the Madison Valley.

The thinning of the tuff away from the Yellowstone-Island Park calderas, the correlation of cooling units from the Yellowstone area to the Centennial region, and potassium-argon dates strongly support the theory that the tuff traveled into the Centennial Valley from the caldera and not from a local vent.

Pleistocene Mesa Falls Tuff

The Mesa Falls Tuff is the second unit of the Yellowstone Group (Christiansen and Blank, 1972). The tuff originated from the Island Park caldera 1.2 m.y. ago.

The Mesa Falls Tuff of the Centennial region is a single ash-flow sheet. The tuff is brownish gray with phenocrysts of quartz and sanidine that comprise up to 45 percent of the rock, situated in a groundmass of devitrified shards. Phenocrysts of quartz and sanidine range up to 5 mm in diameter. This tuff is seen only locally in the Centennial Valley.

Pleistocene Lava Creek Tuff

The Lava Creek Tuff is the youngest unit of the Yellowstone Group (Christiansen and Blank, 1972). The tuff originated from the Yellowstone caldera 0.6 m.y. ago. The tuff is light gray, punky to dense, and has a fine-grained to aphanitic groundmass of devitrified shards. Phenocrysts of sanidine and quartz, averaging less than 1 mm across, make up 2 to 10 percent of the rock. Spherical vesicles mark the presence of a gas phase at the top of the ash-flow sheet. The Lava Creek Tuff is present to the east of the Centennial Valley but is absent in the study area.

APPENDIX C
2-D gravity models

(**Note**: The two-dimensional models used to construct **Sheet 4 (back pocket)** are included in this appendix.)

For all models, left is the primed end of the profile. There is no vertical exaggeration. The models were developed using an interactive Talwani-type program limited to a single-density contrast (Talwani and others, 1959). By manipulating the properties of line integrals, positive and negative multiples of the density contrast were incorporated into the models. Any piece of a model above the zero datum has a positive density contrast (with respect to the basement) of the same magnitude as the negative density contrast of the fill and low-density material. The volcanic material was assigned the same density as the basement, whereas the intrusion was given a density equal to the fill.

Sheet 4 is based on the models preferred by the authors given the available program, but they are by no means the only possible model (Schofield, 1980). A more realistic interpretation would include rapid lateral density variations near faults and Quaternary deposition. It would also allow for vertical change of density in the Tertiary sediments more than 0.5 g/cc.

The range of density contrasts used in the models is indicative of different basement materials and different average Tertiary densities. The density contrasts (in g/cc) used for the models are shown in **Figures C-1 through C-11.**

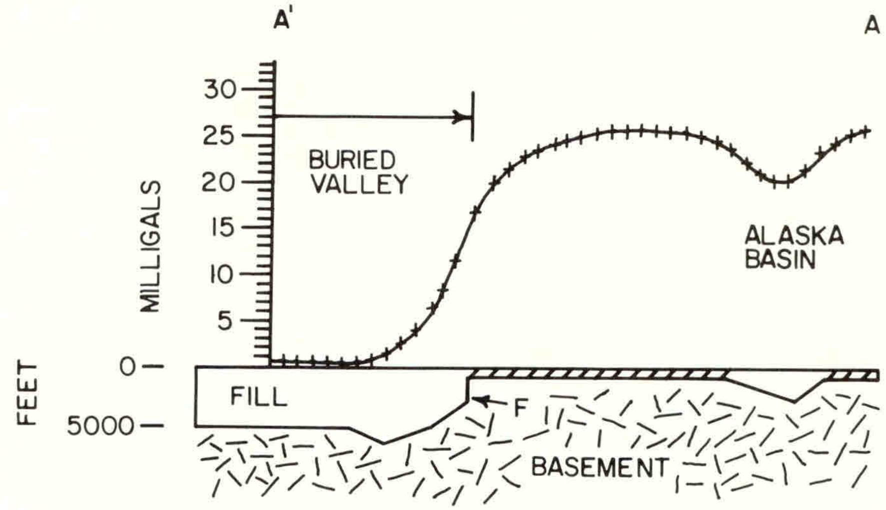

Figure C-1 — A'A (– 0.539).

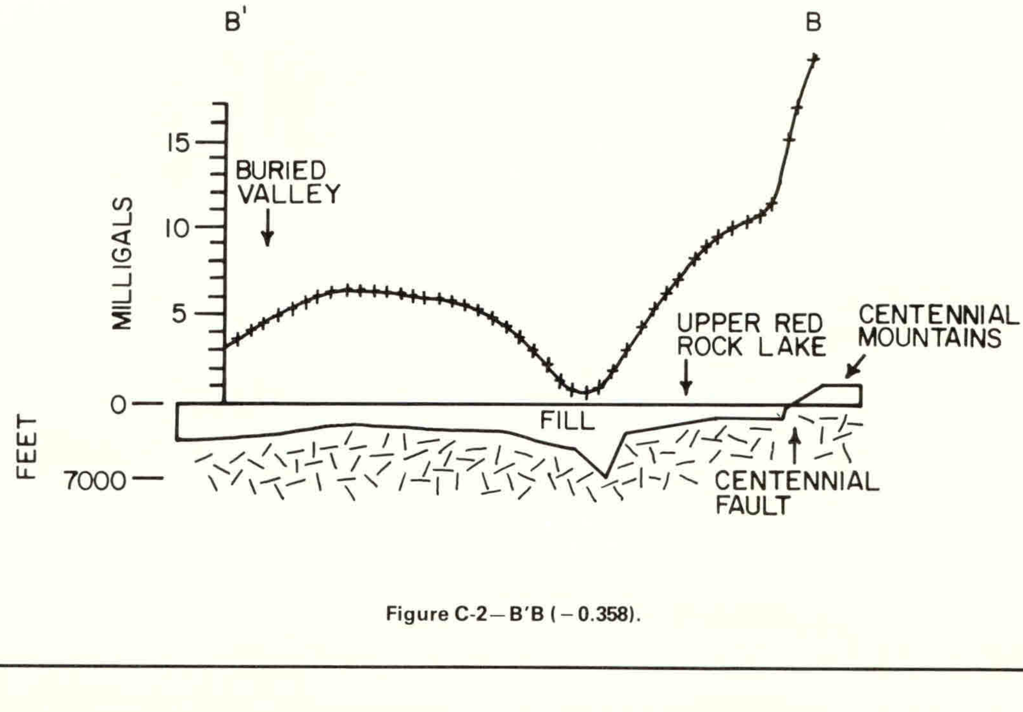

Figure C-2 — B'B (– 0.358).

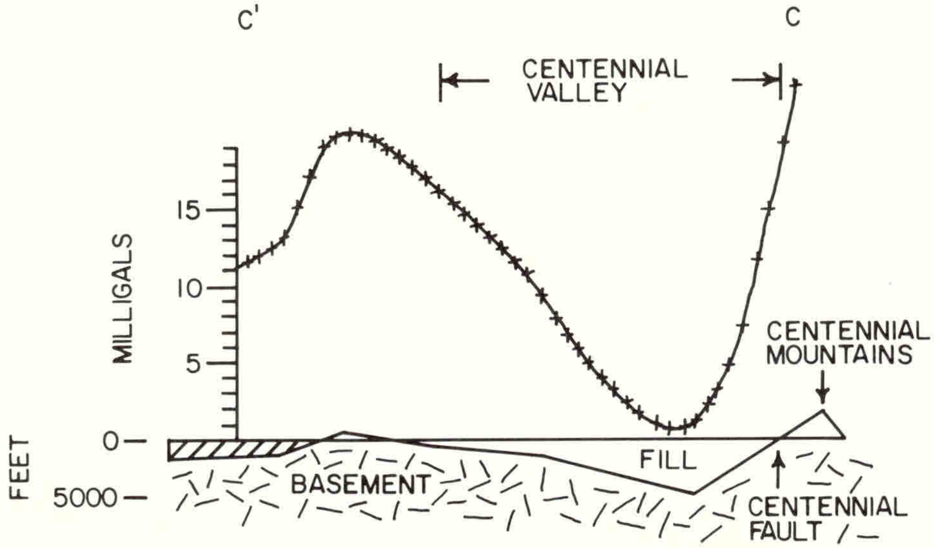

Figure C-3—C'C (−0.414).

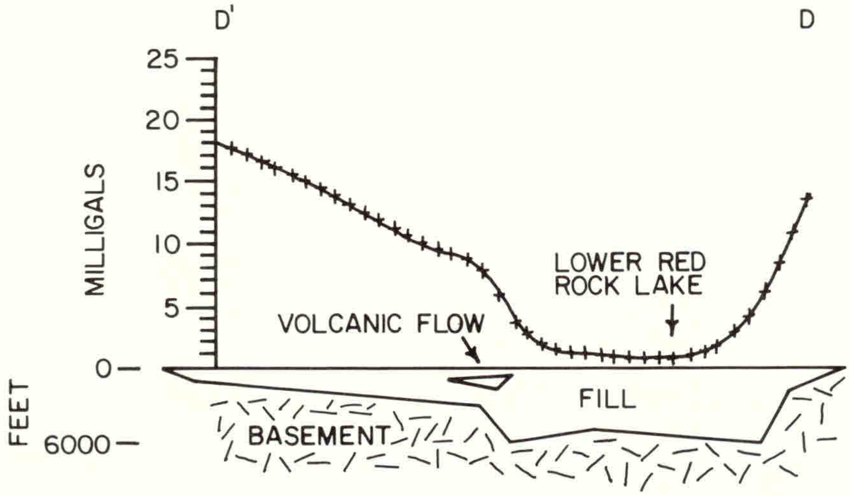

Figure C-4—D'D (−0.371).

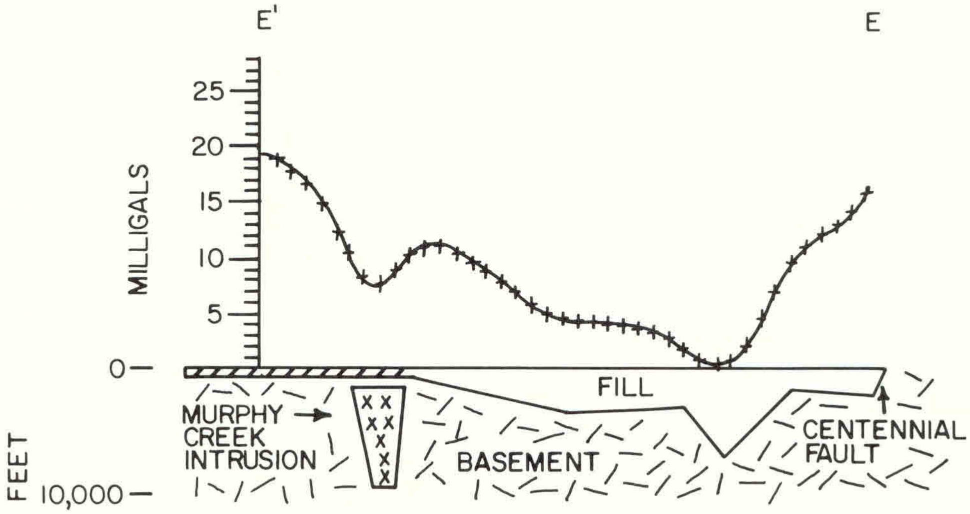

Figure C-5—E'E (−0.564).

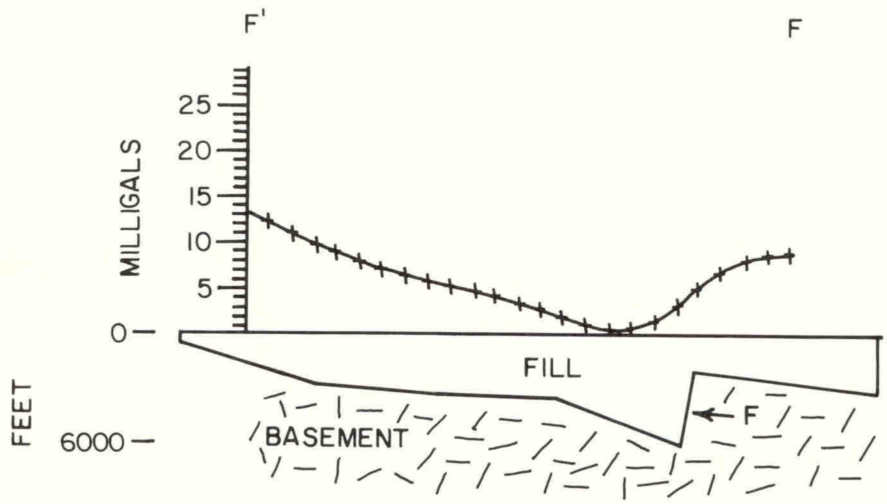

Figure C-6—F'F (−0.615).

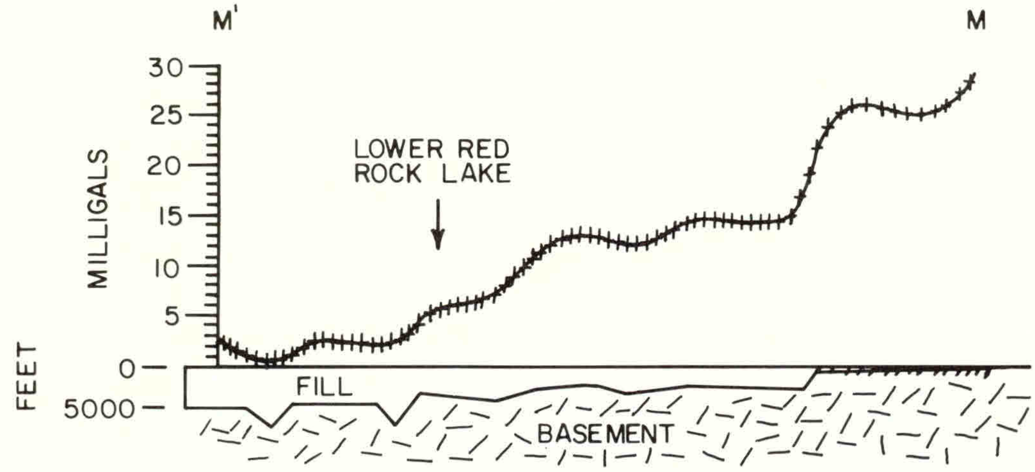

Figure C-7— M'M (– 0.478).

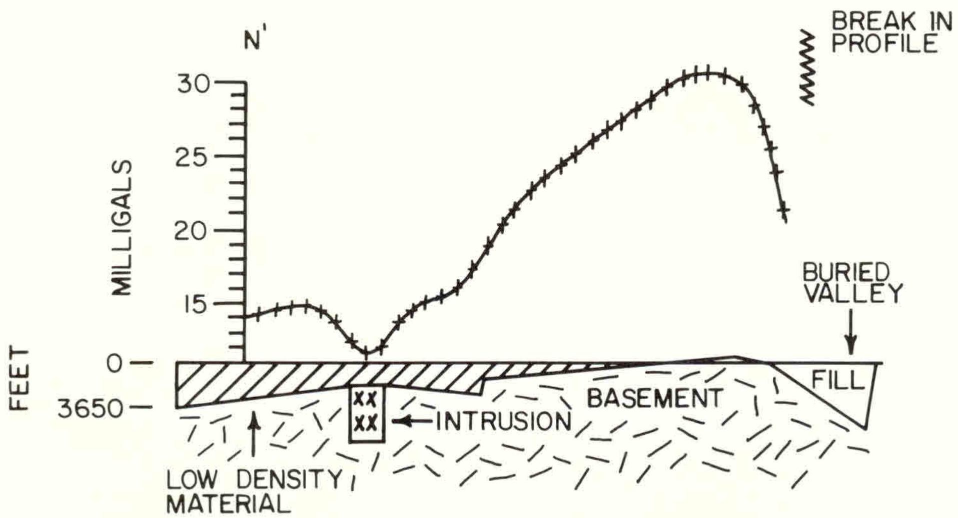

Figure C-8— N'-break (– 0.486).

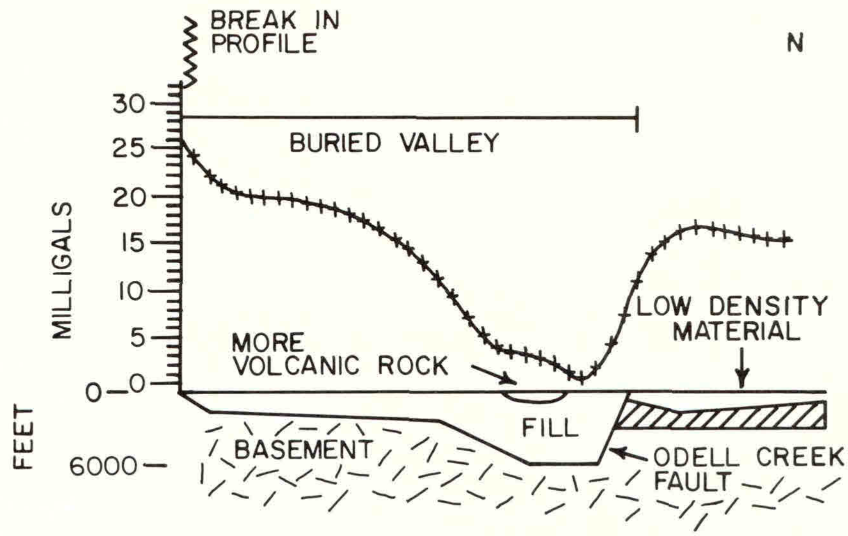

Figure C-9—Break-N (−0.556).

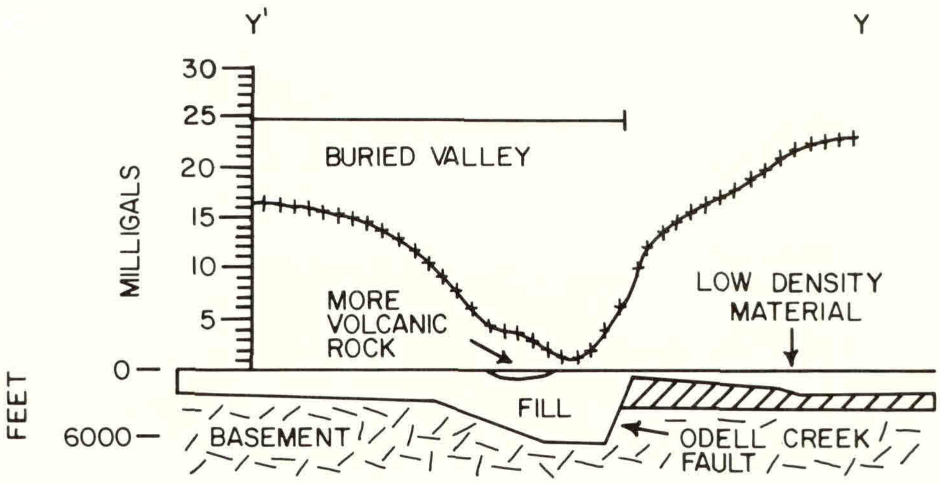

Figure C-10—Y'Y (−0.441).

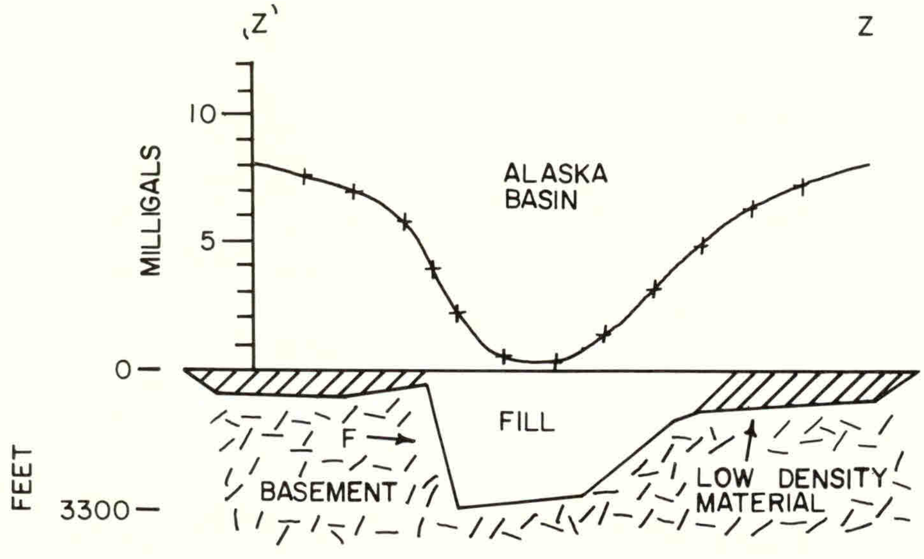

Figure C-11—Z'Z (−0.523).

APPENDIX D
Hydrologic data

[Note Tables D-1, D-2 and D-3 follow.]

Table D-1—Sampled springs and streams.

Location T	Location R	Location Section	Name	Altitude (ft)	Geologic source	Flow (gpm)	Field temp. °C	Field conductance µmho/cm @ 25°C	Water use	Lab no.	Remarks
10S	1E	4CCC	Wolf Creek Warm Spring	6,100	Terrace deposits	35	11.8	310	Stock	78M0291; 78M0424	
10S	1E	7CAB	Wall Canyon Warm Spring	5,800	Alluvium/Precambrian		24		Unused	78M0293	
10S	1E	7CAB	Wall Creek	5,800			15		Unused	78M0294	Stream
10S	1E	9BBBB	Wolf Creek Hot Spring	6,085	Terrace deposits/ Precambrian	53	54	494	Stock	78M0292; 79M3773; 78M0425 See text.	
10S	1E	28DDD	Snow Ball Cold Spring	6,000	Landslide and slump		10		Unused	78M0296	
11S	1E	13DBB	Curlew Creek Schuster's place	6,240			16		Stock	78M0289	Stream
11S	1E	13DBB	Schuster's place—Hillslope Spring	6,280	Terrace deposits	15	9.5	128	Stock	78M0430	
11S	1E	13DBC	Curlew Creek Warm Spring	6,220	Tuff	36	23		Unused	78M0290	
11S	1E	24BBD	Curlew Creek (at road)	5,950		1,300	10	154	Multiple use	78M0432	Stream
11S	1E	24BDB	Deadman Creek (at road)	5,930		1,400	7.5	96	Multiple use	78M0431	Stream
11S	1E	24DBD	Madison River Lodge (spring at highway)	6,000	Terrace deposits	450	8.2	96	Unused	78M0392	
11S	1E	27AAA	West Fork above Soap and Lake creeks	6,050		14,600	8.0	253	Multiple use	78M0394	Stream
11S	2E	32DCDC	Cliff Lake Town Spring	6,080	Terrace deposits	450	8.7	285	Multiple use	78M0435	
12S	1E	11ADD	Wade Lake Spring	6,280	Tuff	20,000*	11.8	204	Public supply	78M0436	
12S	1E	18DB	West Fork Swimming Hole	6,720	Basalt dike(?) terrace deposits	480	25.5	322	Multiple use	78M0295; 78M0395	
12S	1E	19CD	Sloan Cow Camp Cold Spring	6,540	Colluvium		7.4	254	Domestic	78M0396	
12S	1E	19CD	Sloan Cow Camp Warm Spring	6,540	Uncertain	350	29.8	410	Bathing	78M0397	
12S	2E	20DDD	Horn Creek Spring	6,450	Tuff	1,240	11.5	260	Domestic and stock	78M0437	
13S	1W	4DCC	Tepee Creek Spring (1.5 mi SW of Snowshoe Pass)	7,230	Tuff	200	5	79	Stock	78M0461	Discharge at tuff-basalt contact
13S	1W	10BBA	Tepee Creek Spring (1 mi S of Snowshoe Pass)	7,210	Tuff	2	7.4	61	Stock	78M0475	
13S	1W	12DAAC02	Lower spring (.75 mi from VABM Brimstone)	7,730	Tuff	0.08	6.8	92	Stock	78M0381	
13S	1W	12DAAC01	Springs (.75 mi S of Two Drink Springs)	7,710	Tuff	0.4	6.3	73	Stock	78M0382	
13S	2W	1BAA	Martinell stock tank	7,480	Volcanics	0.2	4.7	439	Stock	78M0455	Discharge at tuff-basalt contact
13S	2W	5CAAA	Cayuse Spring (2.5 mi N of Staudenmeyer's)	7,950	Tertiary limestone	0.05	5.9	336	Stock	78M0383	
13S	2W	17CBD	Uppermost spring—Staudenmeyer ranch	6,820	Madison	1,340	28	644	Multiple use	78M0440	Discharge from tuff
13S	2W	17CBD	Upper west spring—Staudenmeyer ranch	6,810	Madison	450	29	649	Multiple use	78M0441	Discharge from tuff
13S	2W	17CBD	Lower west spring—Staudenmeyer ranch	6,800	Madison	50	31	645	Multiple use	78M0442	Discharge from tuff
13S	2W	17CBD	Upper east spring—Staudenmeyer ranch	6,810	Madison	220	29	659	Multiple use	78M0444	Discharge from tuff
13S	2W	17CBD	Lower east spring—Staudenmeyer ranch	6,800	Madison	100	28	681	Multiple use	78M0445	Discharge from tuff
13S	2W	18ADC	Jimmy Anderson's Spring no. 1	6,840	Madison	900	28	614	Stock	78M0438	Discharge from Phosphoria
13S	2W	18BDAD	Jimmy Anderson's Spring no. 2	6,840	Madison	900	22	709	Stock	78M0439	Discharge from tuff and Phosphoria colluvium
13S	2W	18DB	Jimmy Anderson's Spring	6,750	Madison	25	24.5	614	Stock		
13S	3W	22CCBC	Lousy Spring (2.5 mi NE of Brundage Bridge)	6,830	Tuff	410	7.5	380	Stock	78M0459	

Table D-1 *(continued)*.

Location T	R	Section	Name	Altitude (ft)	Geologic source	Flow (gpm)	Field temp. °C	Field conductance µmho/cm @ 25°C	Water use	Lab no.	Remarks
13S	3W	22DAAB	Small spring (1 mi ENE of Lousy Spring)	6,770	Tuff	4	9.5	322	Stock	78M0460	
13S	3W	23ABD	Spring (1.5 mi W of Fish Creek road junction)	6,810	Colluvium	3	7	313	Stock	78M0398	
13S	1E	7AAD	Brimstone Creek Spring	7,700	Tuff	5	9.9	86	Wildlife		Bear prevented sampling
13S	1E	9DBBC	Hidden Lake Spring (SW of Hidden Lake)	6,715	Tuff	38	6.9	192	Multiple use	78M0380	
13S	1E	28AACD	Spring NW of Deer Mt. on U.S. Forest land	7,320	Tuff	2	9	322	Stock	78M0391	
13S	1E	29CAB	Elk Lake camp water supply	6,710	Tuff	30	9.0	114	Public	78M0476	Spring
13S	1E	31DCAB	Upper Elk Springs (1 mi SW of Elk Lake)	6,640	Alluvium	630	8.5	223	Multiple use	78M0386	
13S	1E	33AAC	Limestone Creek Spring	7,250	Amphibolite	2	11.75	235	Stock	78M0465	
13S	1E	36DDD	Spring (head of Lone Tree Creek)	7,410	Quartzite	0.5	18.9	150	Stock	78M0466	
13S	2E	19DADB	Spring for livestock	6,880	Amphibolite	10	17.8	168	Stock		
13S	2E	20CBB	1 mi SE of Conklin Lake (Orr ranch)	6,880	Amphibolite	18	10.8	189	Stock	78M0384	
13S	2E	31CADD	Spring for livestock	7,640	Quartzite	5	11.3	80	Stock	78M0390	
13S	2E	32CAB	Main spring headwater for Antelope Creek	7,680	Amphibolite	56	7.9	122	Multiple use	78M0467	
14S	1W	21CDD	U.S. Fish-Wildlife campground	6,620	Alluvial fan	4	7.5	353	Public	78M0399	
14S	1W	21DDD	Spring in marsh above lake	6,615	Alluvium	2	8.5	362			
14S	1W	22DAB	Spring in marsh above lake (1.25 mi E of Upper Red Rock Lake campground)	6,640	Alluvium	450	8	418	Multiple use	78M0468	
14S	1W	23DBA	Spring in marsh above lake (2 mi E of Upper Red Rock Lake campground)	6,635	Alluvium	1	6.1	355	Stock	78M0470	
14S	1W	23DBC	Spring in marsh above lake	6,620	Alluvium	1	6	356			
14S	2W	6BDD	Lower Red Rock Lake dam	6,605		10,900	6.2	187	Multiple use	78M0471	Reservoir
14S	2W	23BDD	Shambow Creek below septic tank – Lakeview	6,710		620	3.8	182	Multiple use	78M0473	Stream
14S	3W	23BBD	Huntsman ranch spring	6,630	Miocene(?) basalt	310	9.8	381	Domestic and stock	78M0402	
14S	1E	3CAD	Spring for livestock	7,520	Metagabbro	672	5.5	140	Stock		
14S	1E	3CBAD	Huntsman Spring (1.5 mi SE of Elk Mountain)	7,520	Quartzite	5	17.8	176	Stock	78M0474	
14S	1E	8DACC	Culver Spring	6,680	Alluvial fan	11,200*	7.5	294	Multiple use	78M0458	
14S	1E	13BDA	Huntsman Spring (Alaska basin)	6,800	Dolomite	815	8.1	283	Domestic and stock	78M0469	Discharges from colluvial or alluvial material
14S	1E	15CC	Fruin Spring	6,740	Tuff	18.7	8.2	290	Irrigation		
14S	1E	20CAB	Spring above Walsh ranch	7,040	Tuff	0.25	14	78	Domestic and stock	78M0477	
14S	1E	22ABCC	Corral Creek at Tobe Morton ranch	6,775		1,240	6.1	200	Multiple use	78M0464	Stream
14S	1E	23BCC	Small spring E of Tobe Morton ranch	6,800	Alluvial fan	59	7	102	Stock	78M0385	
14S	1E	23CDC	Picnic Creek Spring	7,080	Tuff	107	5.8	50			
14S	1E	29ACD	2.5 mi SW of Walsh ranch	7,120			6.5	261	Multiple use	78M0462	Stream
14S	1E	29DBAB	Spring at W fork of Antelope Creek source	7,140	Tuff	77	5.1	80	Stock	78M0463	
14S	2E	6CBD	Spring for stock (0.5 mi SE of Lone Tree Pass)	7,480	Quartzite	13	21.2	216	Stock	78M0389	
14S	2E	6CCCB	Spring (1 mi S of Lone Tree Pass)	7,360	Quartzite	29	19	230	Stock	78M0388	
14S	2E	7CBAD	Spring (2 mi N of Red Rock Pass)	7,420	Dolomite	450	8	297	Stock	78M0387	
14S	2E	7DDC	Spring for stock	7,360	Dolomite	0.5	16.25	327	Stock		

*Visual estimate of spring discharge; conditions did not permit discharge measurement.

Table D-2—Inventoried wells.

T	Location R	Section	Owner's name	Date drilled	Depth of well (ft)	Static water level (ft)	Perforated intervals
10S	1E	6CCB	Bureau of Land Management	7/27/1966	62.5	4	None
10S	1E	7BDD	Sun ranch	9/1950	33		None
10S	1E	7DDB	Sun ranch	8/10/1958	41.2	17	None
10S	1E	17DAB	Sun ranch	8/05/1959	75	15	None
10S	1E	17DAB	Sun ranch	9/02/1959	299	16	None
10S	1E	33DDB	Sun ranch	11/14/1960	79	14	None
11S	1E	3CA	Montana State Highway Dept.	8/26/1966	99	5.7	None
11S	1E	14BBBA	U.S. Forest Service	11/28/1967	101.5		None
11S	1E	35ACCD	Vujovich, J. J.	9/05/1969	595	545	None
12S	1E	1ABDA	Vujovich, J. J.			31.2	
12S	1E	12ABDB	Vujovich, J. J.			191.8	
13S	1W	19BADD	Centennial Livestock	8/09/1972	105	24.3	95-102
13S	1W	20DA	Kessler, Fred		60	42.9	
13S	1W	21DB	In litigation	8/25/1961	111	57.2	None
13S	1W	22CCD	Jones, Willis	6/27/1963	86	58.2	None
13S	1W	26DD	U.S. Dept. of Interior Red Rock National Wild Life Refuge	10/16/1953	85.5	60.1	None
13S	1W	28BCB	Smith, Stanley			10.9	None
13S	1W	28DDD	Smith, Stanley	1959	11	5.1	None
13S	1W	30CBA	Centennial Livestock	1960	92	20	None
13S	2W	12BCC	Martinell, Lee	8/08/1959	181	159	None
13S	2W	14BDC	Kessler, Fred	7/29/1963	92	65	None
13S	2W	17CCB	Centennial Livestock (Les Staudenmeyer)	1960	38	10	None
13S	2W	20AAA	Anderson, Jim			63.4	None
13S	2W	20ADB	Anderson, Jim			29.1	None
13S	2W	24DCCC	Centennial Livestock	1937	207(?)	9.2	None
13S	2W	26DDDB	Centennial Livestock		28	7.4	None
14S	1E	17BDB	Old Hanson ranch		29	10.2	None
14S	1E	17CB	Selby	1936	38	12.2	None
14S	2W	22AACD	Montgomery Children	8/21/1971	100	65	None
14S	2W	23BCA	Rush, Keith		38	7.3	None
14S	2W	23BDAA	Old Schoolhouse—Lakeview		52		
14S	2W	23CAA	U.S. Fish and Wildlife Refuge	7/18/1956	52	20	None
14S	3W	21D	Breneman, Sam	10/26/1966	68	6	63-68
14S	3W	21D	Breneman	9/15/1970	53	25	None
14S	3W	21DCDB	Breneman, Sam E.	6/1951	40	28	None
14S	3W	22CB	Breneman, Tom	12/24/1975	148	22.6	132-143

[Columns are continuous across both pages.]

Yield (gpm)	Drawdown (ft)	Geologic unit	Altitude of well (ft)	Water use	Field temp. °C	Field conductance µmho/cm @ 25°C	Lab no.
30	40	Quaternary terrace or alluvium	5650	Public supply	8.5	234	78M0434
		Quaternary terrace deposits	5715	Domestic	6.0	142	78M0427
30	13	Quaternary terrace deposits	5740	Stock	14.8	276	78M0426
25	30	Pleistocene volcanics	5840	Domestic	8.9	236	78M0428
20	59	Pleistocene volcanics	5800	Domestic	8.0	278	78M0429
30	44	Pleistocene volcanics	5880	Domestic	6.0	148	78M0433
30	8.8	Quaternary terrace deposits	5870	Public supply	11.3	417	78M0393
15	91	Terrace deposits	5925	Public supply			
12	50	Terrace deposits	6580	Stock			
			6260	Stock			
			6360	Stock			
12	4	Quaternary sands	6676.3	Stock	8.5	294	78M0451
		Quaternary sands	6693.3	Stock; secondary use-unused			
10		Quaternary sands	6709.3	Unused			
20	10	Quaternary sands	6707.9	Stock			
30		Quaternary sands	6713.3	Stock	8.3	221	78M0456
		Quatenary sands	6662.3	Domestic	6.5	108	78M0449
		Quaternary alluvium	6649.0	Unused; secondary use-stock	9.3	474	78M0450
20		Quaternary sands	6645	Stock	7.5	324	78M0452
15		Tertiary sediments	7020	Stock			
20	10	Quaternary sands	6710.7	Unused; secondary use-stock			
35		Holocene alluvial fan deposits	6760	Domestic	10.2	786	78M0446
		Holocene alluvial fan deposits	6713.0	Stock	7.4	528	78M0447
10		Quaternary alluvium	6680.1	Domestic and stock	7.5	543	78M0448
12		Quaternary sands	6660.4	Stock			78M0443
15		Quaternary alluvium	6635.8	Stock	7.0	452	78M0453
		Quaternary alluvial fan	6675	Unused; secondary use-domestic	8.0	290	78M0454
		Quaternary alluvial fan	6700	Domestic			
20	15	Alluvial fan (?)	6720	Domestic	7.5	300	78M0401
		Alluvial fan (?)	6765	Stock	4.5	517	78M0400
30		Quaternary alluvial fan deposits	6690	Domestic	7.7	404	78M0457
17 +	10	Quaternary alluvial fan	6715	Secondary use-public supply	7.7	404	78M0472
50		Quaternary alluvial fan	6660	Stock			
20		Quaternary alluvial fan	6680	Domestic			
25		Quaternary alluvial fan	6670	Domestic			
15	9(?)	Quaternary alluvial fan	6670	Stock			

Table D-3—Water-quality analyses (springs and wells).

Analysis number	Date	Ca	Mg	Na	K	Fe	Mn	SiO_2	CO_3	HCO_3	Cl	SO_4	NO_3 as N	F	As	B	H_2S	Li	U	Field pH	T.D.S.	S.C.
78M0289	09-09-77	16.5	3.5	5.3	1.7	.28	.12	19.	0	79.	1.55	2.8	.059	.3	--	-	-	-	-	7.68#	90.	131.
78M0290	09-09-77	12.5	1.3	33.	1.2	1.11	.01	19.7	.6	88.5	4.05	12.3	.377	6.4	--	-	-	-	-	8.35#	136.	217.
78M0291	09-13-77	14.7	2.	58.	1.4	.18	<.01	33.8	3.8	136.	10.10	26.2	.221	7.2	--	-	-	-	-	8.49#	225.	343.
78M0292	09-13-77	8.9	1.7	96.	1.8	.07	<.01	49.4	6.0	158.	19.3	43.9	.059	13.2	--	-	-	-	-	8.50#	318.	474.
78M0293	09-13-77	6.6	1.7	260.	6.	.08	.02	41.7	0	493	49.2	80.8	.052	14.4	--	-	-	-	-	8.06#	703.	1097.
78M0294	09-13-77	33.4	8.	8.2	2.1	.22	<.01	25.7	1.8	149.	2.3	4.6	.221	.3	--	-	-	-	-	8.37#	160.	247.
78M0295	09-13-77	18.6	28.8	4.9	1.9	.02	<.01	13.5	12.	177	1.9	11.2	.226	.4	--	-	-	-	-	8.71#	181.	308.
78M0296	09-14-77	31.4	4.9	18.8	3.9	.05	.02	50.9	0	138	10.2	7.3	.172	.5	--	-	-	-	-	6.99#	196.	268.
78M0380	09-29-77	17.2	2.4	3.9	3.7	.01	<.01	49.	0	61.5	3.85	4.6	.398	.2	<.002	.03	<.10	<.01	.0003	6.74#	116.	131.
78M0381	09-29-77	8.	1.6	3.4	10.4	1.63	.49	24.6	0	46.	4.	2.9	.115	<.1	<.002	.02	.17	<.01	.0002	5.75#	80.	99.3
78M0382	09-29-77	5.4	1.3	2.3	2.2	.75	.09	28.	0	24.2	1.45	3.2	.251	<.1	<.002	.02	<.10	<.01	<.0002	6.24#	57.	56.3
78M0383	09-29-77	70.5	4.6	3.4	2.5	.02	.01	22.9	0	224.	8.45	7.4	1.54	.5	.0025	.02	.12	<.01	.002	7.09#	232.	386.
78M0384	09-28-77	20.	6.4	5.3	.9	.10	.01	22.9	0	99.2	4.3	4.5	.178	.1	<.002	.01	.17	<.01	<.0002	7.64#	114.	168.
78M0385	09-28-77	4.4	.7	2.8	1.1	.06	.01	21.8	0	21.2	1.65	3.4	.097	<.1	<.002	.02	.21	<.01	<.0002	6.99#	47.	46.1
78M0386	09-28-77	23.6	4.2	4.8	4.	.01	<.01	34.4	0	98.	4.1	3.2	.404	.4	<.002	.03	<.10	<.01	.0006	7.44#	127.	178.
78M0387	09-27-77	30.4	18.2	2.	1.8	.04	<.01	11.1	0	186.	2.05	2.5	.210	<.1	--	-	-	-	-	7.86#	160.	284.
78M0388	09-27-77	23.9	10.	2.4	1.3	.02	<.01	15.2	0	126.	1.90	1.5	.490	<.1	--	-	-	-	-	7.57#	119.	204.
78M0389	09-27-77	23.	9.7	2.3	1.3	.02	<.01	13.9	0	120	2.45	3.0	.221	<.1	--	-	-	-	-	7.54#	115.	195.
78M0390	09-27-77	3.9	1.2	1.5	.3	.21	<.01	11.3	0	13.7	1.2	.5	.768	<.1	--	-	-	-	-	6.23#	28.	36.4
78M0391	09-27-77	35.4	12.	2.5	.4	.09	<.01	15.0	0	169.	1.95	3.5	.312	<.1	--	-	-	-	-	7.63#	155.	263.
78M0392	09-29-77	12.8	2.4	3.2	1.0	.01	<.01	15.8	0	54.2	.85	2.5	.208	.6	--	-	-	-	-	7.22#	66.	99.2
78M0393	09-29-77	51.4	6.8	28.	7.1	.02	<.01	43.2	0	251.	6.95	6.6	.249	1.6	--	-	-	-	-	7.28#	276.	416.
78M0394	09-29-77	36.	8.	5.1	2.2	.03	<.01	22.	2.4	149.	2.25	7.6	.023	.2	--	-	-	-	-	8.42#	159.	249.
78M0395	09-29-77	19.	29.	4.8	1.9	<.01	<.01	13.7	0	194.	2.75	11.8	.099	.4	.0028	.02	.17	<.01	.001	8.3	179.	321.
78M0396	09-29-77	38.8	1.6	7.6	6.8	<.01	<.01	67.2	0	144.	5.80	5.1	.411	.2	--	-	-	-	-	8.6	205.	254.
78M0397	09-29-77	.9	.1	88.	1.1	.17	<.01	50.9	74.4	64.2	7.65	3.7	.050	3.1	<.002	.16	.94	<.01	.0003	10.05	262.	396.
78M0398	09-28-77	52.6	8.	13.3	4.	.02	<.01	37.9	0	210.	8.55	10.4	.312	.3	<.002	.01	.25	<.01	.0036	7.8	239.	360.
78M0399	09-27-77	46.2	16.5	4.8	1.0	.03	<.01	10.7	0	231.	1.20	4.9	0.75	.2	--	-	-	-	-	7.82#	200.	352.
78M0400	09-27-77	82.5	13.4	12.	5.8	.17	.02	44.1	0	345.	3.85	5.2	.097	.2	--	-	-	-	-	7.23#	337.	531.
78M0401	09-27-77	42.8	8.4	8.5	3.5	--	.01	34.2	0	195.	1.60	2.9	.181	.1	<0.010	-	-	-	-	7.84#	198.	301.
78M0402	09-27-77	38.4	13.2	20.9	4.8	<.01	<.01	37.2	0	223.	12.3	8.1	.325	.2	<.002	.02	.09	<.01	.0028	7.31	245.	378.
78M0424	09-30-77	27.5	4.7	30.8	3.4	.24	.13	21.4	7.8	121.	6.	31.2	.380	4.3	<.002	.02	.08	.04	.0023	10.2	197.	300.
78M0425	09-30-77	8.6	1.5	97.	1.8	<.01	<.01	50.3	8.2	154.	20.8	42.6	.063	16.	.005	.03	.20	.07	.0006	11.03	307.	483.
78M0426	09-30-77	38.	7.9	12.8	1.5	.02	<.01	21.6	0	133.	15.9	23.1	.700	.3	--	-	-	-	-	7.40#	187.	305.
78M0427	09-30-77	18.9	3.8	7.0	1.0	.01	<.01	10.7	0	84.2	2.0	7.2	.131	.4	--	-	-	-	-	7.22#	93.	155.
78M0428	10-01-77	34.5	4.	10.8	4.2	.01	<.01	42.8	0	141.	5.60	7.	.633	.3	--	-	-	-	-	7.64#	179.	255.
78M0429	10-01-77	41.2	5.4	8.7	3.7	<.01	<.01	40.2	0	153.	8.05	8.3	.531	.3	<.002	.02	.28	<.01	.0031	8.57	192.	277.
78M0430	10-01-77	15.7	3.2	4.7	1.	.03	<.01	19.3	1.2	68.	1.40	4.1	.644	.4	<.002	.02	.17	<.01	.0006	8.10	85.	124.
78M0431	10-01-77	12.2	2.2	3.3	1.1	.02	<.01	13.5	0	52.5	.95	0	.242	.5	--	-	-	-	-	8.05#	60.	94.
78M0432	10-01-77	17.6	3.1	8.7	1.3	.02	<.01	18.4	21.8	38.2	2.50	3.7	.355	2.	--	-	-	-	-	9.72#	98.	148.
78M0433	10-05-77	23.5	4.	6.4	3.9	.38	.01	43.2	0	98.9	4.55	0	.291	.5	--	-	-	-	-	7.12#	135.	186.
78M0434	10-02-77	19.	5.9	21.2	1.8	.05	.01	17.8	0	121.	12.7	6.1	.194	1.1	--	-	-	-	-	7.88#	145.	236.
78M0435	10-27-77	29.2	17.2	4.6	1.7	.02	<.01	11.6	0	181.	1.75	4.3	.492	.1	--	-	-	-	-	8.30#	160.	283.
78M0436	10-02-77	23.3	11.5	5.2	1.7	<.01	<.01	13.3	0	139.	2.30	2.	.237	.4	--	-	-	-	-	7.93#	128.	223.

78M0437	10-02-77	27.	14.2	9.6	2.	<.01	<.01	17.3	0	170.	3.65	4.2	.565	.3	<.002	.03	.12	<.01	.0014	8.5	163.	276.
78M0438	10-03-77	66.5	24.	27.7	7.3	<.01	<.01	21.4	0	246.	9.7	114.	.036	1.7	.0136	.20	<.10	.05	.0024	7.4	394.	615.
78M0439	10-03-77	71.	24.	26.9	7.3	.01	<.01	21.0	0	247.	9.	118.	.066	1.8	.0152	.23	<.10	.05	.0025	7.4	401.	627.
78M0440	10-03-77	67.5	24.5	25.8	6.8	.01	<.01	20.1	0	244.	9.25	114.	.075	1.8	.016	.20	<.10	.05	.0022	7.57	390.	608.
78M0441	10-03-77	67.	24.	27.9	7.2	.02	<.01	20.8	0	249.	9.8	114.	.056	1.8	.0118	.20	.17	.05	.0019	7.54	395.	617.
78M0442	10-03-77	68.	24.	29.	7.7	<.01	<.01	21.4	0	251.	9.35	116.	.050	1.8	.0154	.23	<.10	.05	.0027	7.44	401.	626.
78M0443	10-03-77	44.	7.8	4.6	3.	<.01	<.01	30.8	0	170.	5.05	3.3	1.717	.2	--	-	-	-	-	7.91#	184.	291.
78M0444	10-04-77	69.	25.2	28.1	7.4	<.01	<.01	22.7	0	250.	10.	115.	.038	1.8	--	-	-	-	-	7.52#	402.	629.
78M0445	10-04-77	68.	24.6	27.8	7.4	<.01	<.01	23.3	0	251.	9.80	114.	.027	1.8	--	-	-	-	-	7.48#	400.	628.
78M0446	10-04-77	87.5	34.	36.2	12.1	.02	<.01	30.	0	342.	14.2	134.	.488	2.	--	-	-	-	-	7.55#	519.	799.
78M0447	10-04-77	72.5	11.5	22.7	5.	<.01	<.01	36.4	0	237.	12.3	69.	.813	.2	<.002	.05	<.10	.01	.0039	8.16	347.	523.
78M0448	10-04-77	82.	13.1	21.1	5.	.02	.01	34.7	0	258.	8.	84.6	.380	.2	--	-	-	-	-	7.98#	376.	569.
78M0449	10-04-77	33.	5.7	3.6	2.9	.07	.21	6.8	0	147.	3.	.2	<.023	.1	--	-	-	-	-	7.8	128.	234.
78M0450	10-04-77	59.	23.	7.5	4.4	.02	.74	15.	0	303.	5.70	4.2	1.717	.7	--	-	-	-	-	7.88#	271.	472.
78M0451	10-04-77	50.	7.1	5.1	3.	.01	<.01	31.9	0	192.	4.10	4.1	.971	.2	--	-	-	-	-	7.99#	201.	310.
78M0452	10-04-77	45.5	6.7	3.6	1.7	.01	<.01	25.7	0	183.	2.25	0	.678	.1	--	-	-	-	-	7.93#	176.	294.
78M0453	10-04-77	65.5	12.4	13.8	7.3	1.22	.54	51.8	0	263.	25.9	3.6	1.536	.3	--	-	-	-	-	7.47#	313.	487.
78M0454	10-05-77	39.2	15.	2.3	1.4	1.25	.25	12.2	0	192.	1.10	3.4	.147	.1	--	-	-	-	-	7.40#	171.	314.
78M0455	10-05-77	58.2	16.3	6.6	3.6	.02	.02	31.	0	251.	11.7	5.9	.858	.3	--	-	-	-	-	7.86#	258.	417.
78M0456	10-05-77	27.	3.9	5.3	2.6	.13	.03	37.	0	96.	6.65	3.2	1.9	.4	--	-	-	-	-	7.25#	135.	198.
78M0457	10-05-77	60.	9.7	8.7	5.1	.02	.10	30.4	0	247.	4.85	6.5	.384	<.1	--	-	-	-	-	7.42#	248.	401.
78M0458	09-28-77	33.	13.6	2.3	1.1	<.01	<.01	16.3	0	171.	1.20	1.9	.167	<.1	--	-	-	-	-	7.92#	154.	263.
78M0459	09-30-77	48.	10.3	8.4	1.1	<.01	<.01	19.3	0	204.	5.35	3.2	.655	.1	--	-	-	-	-	7.58#	197.	334.
78M0460	09-29-77	50.8	12.3	12.7	1.8	.04	<.01	21.	0	212.	6.4	5.2	.587	.2	--	-	-	-	-	7.76#	215.	318.
78M0461	09-30-77	17.4	2.9	4.3	5.3	.19	.04	35.5	0	70.5	5.60	1.2	.971	.2	--	-	-	-	-	6.91#	108.	146.
78M0462	10-01-77	42.2	13.	1.4	.8	.02	<.01	9.	3.8	184.	.80	5.2	.043	<.1	--	-	-	-	-	8.47#	167.	293.
78M0463	10-01-77	13.2	2.4	2.3	1.1	.02	<.01	13.	0	54.9	.60	3.4	.056	<.1	--	-	-	-	-	7.48#	63.2	98.6
78M0464	10-01-77	27.6	12.4	1.3	.5	<.01	<.01	10.9	5.	136.	.40	6.	.023	<.1	--	-	-	-	-	8.48#	131.	223.
78M0465	10-02-77	23.6	11.	2.4	1.4	<.01	.01	17.5	0	126.	3.	6.	.373	<.1	.002	.01	.17	<.01	.0007	7.71#	128.	209.
78M0466	10-02-77	5.9	1.8	2.8	.6	.24	.09	9.8	0	29.9	2.25	.2	<.023	<.1	--	-	-	-	-	7.18#	38.5	59.6
78M0467	10-02-77	5.7	3.	1.9	.3	.03	<.01	16.7	0	34.5	1.15	1.3	.300	<.1	--	-	-	-	-	7.15#	47.5	64.
78M0468	10-03-77	48.8	16.9	1.4	.9	.03	.02	6.4	0	228.	.40	5.	.183	<.1	--	-	-	-	-	7.95#	192.	347.
78M0469	10-03-77	30.	18.7	2.4	1.1	<.01	<.01	13.9	0	187.	1.45	.9	.255	.1	--	-	-	-	-	7.99#	161.	284.
78M0470	10-03-77	62.	18.3	2.8	.6	<.01	<.01	10.3	0	276.	1.50	5.3	.269	.1	--	-	-	-	-	8.12#	237.	413.
78M0471	10-04-77	21.1	13.5	6.1	1.8	.10	<.01	10.1	0	139.	2.90	5.3	.077	.2	--	-	-	-	-	7.93#	130.	225.
78M0472	10-04-77	42.2	6.7	4.6	2.3	.01	<.01	24.4	0	173.	.95	4.4	.450	.1	--	-	-	-	-	7.97#	171.	268.
78M0473	10-04-77	30.6	4.6	4.2	2.1	.01	<.01	24.4	0	126.	.85	3.9	.108	<.1	--	-	-	-	-	8.20#	133.	200.
78M0474	10-04-77	21.1	8.7	2.7	1.6	.05	<.01	10.7	5.5	95.2	2.25	8.	.036	.2	--	-	-	-	-	8.52	108.	184.
78M0475	10-05-77	21.6	3.3	4.3	3.4	.04	.01	44.1	0	88.2	4.05	2.5	.309	.2	--	-	-	-	-	7.69#	127.	158.
78M0476	10-05-77	20.	3.	4.2	3.6	.03	<.01	48.3	0	79.5	3.65	2.2	.456	.3	--	-	-	-	-	7.89#	125.	147.
78M0477	10-06-77	10.3	2.6	2.7	.8	.16	.01	16.3	0	46.	2.75	3.8	.079	<.1	<.002	.02	.25	<.01	<.0002	7.31#	62.	82.
79M3773	09-28-78	8.0	1.4	104.	1.8	<.01	<.01	50.7	8.4	157.	19.	43.	.029	18.	.0073	-	-	.07	-	8.25	332.	493.

All water-quality information is in milligrams per liter (mg/L).
Includes wells that were discussed in table.
Explanation of symbols:
(#), Laboratory rather than field pH.
(-), Not determined.

Back Pocket

Sheet 1— Geologic map of the northern part of the Upper and Lower Red Rock Lake quadrangles, Beaverhead and Madison counties, Montana.

Sheet 2— Geophysical station and cross section location map.

Sheet 3— Complete Bouguer gravity map.

Sheet 4— Basement interpretation map.

Production Information

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