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STRUCTURAL CONTROLS OF HOT-SPRING SYSTEMS IN SOUTHWESTERN MONTANA

By Robert A. Chadwick and Robert B. Leonard

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ABSTRACT

Thermal waters that issue as hot (more than 38°C) springs in southwestern Montana appear to circulate to depth along Cenozoic block faults, deep fractures penetrating the dominantly crystalline rock crust, or major structural lineaments. At individual hot springs, rising thermal waters are transmitted along conduits formed by the intersection of a major fault with other faults, fracture zones, anticlinal axes (which may be faulted or fractured), or sedimentary aquifers. Step faults and other intravalley faults may influence circulation at some springs. At others, fracture zones alone may provide the necessary vertical permeability.

Normal regional heat apparently is sufficient to maintain the hydrothermal systems without enhancement from cooling igneous bodies. The thermal gradient normally is higher in low thermal conductivity sediments of the block-fault valleys than the 30°C per kilometer average for crystalline rock. To attain reservoir temperatures of 60° to 120°C indicated by chemical geothermometers, waters would have to circulate to depths of about 2 to 4 kilometers in crystalline rock and about 1 to 2 kilometers in valley sediments.

INTRODUCTION

Southwestern Montana, defined herein as the area south of 47° N. latitude and west of 110° W. longitude, is a region of numerous but scattered thermal springs. Within this region the temperatures of 25 "hot" springs (fig. 1), including several having multiple vents, equal or exceed 38°C (100°F). The temperatures of most of the other thermal springs generally range from 20° to 30°C. The well known thermal region of Yellowstone National Park adjoins this region on the southeast.

The purpose of this report is to discuss the regional and local geologic controls that govern the occurrence of the hot springs and their related hydrothermal systems. The report summarizes the results of surface geologic mapping and shallow direct-current resistivity, seismic, and soil-temperature surveys conducted by Montana State University (Chadwick and others, 1978) and coordinated with related hydrogeologic and geochemical investigations conducted by the U.S. Geological Survey (Leonard and others, 1978) during 1975-77. The information is useful for the appraisal and development of geothermal resources in Montana and may be applicable to similar geologic settings elsewhere. Montana State University participation was made possible by Grants Nos. 14-08-0001-G-238 and 334 from the U.S. Geological Survey under the Extramural Geothermal Research Program.

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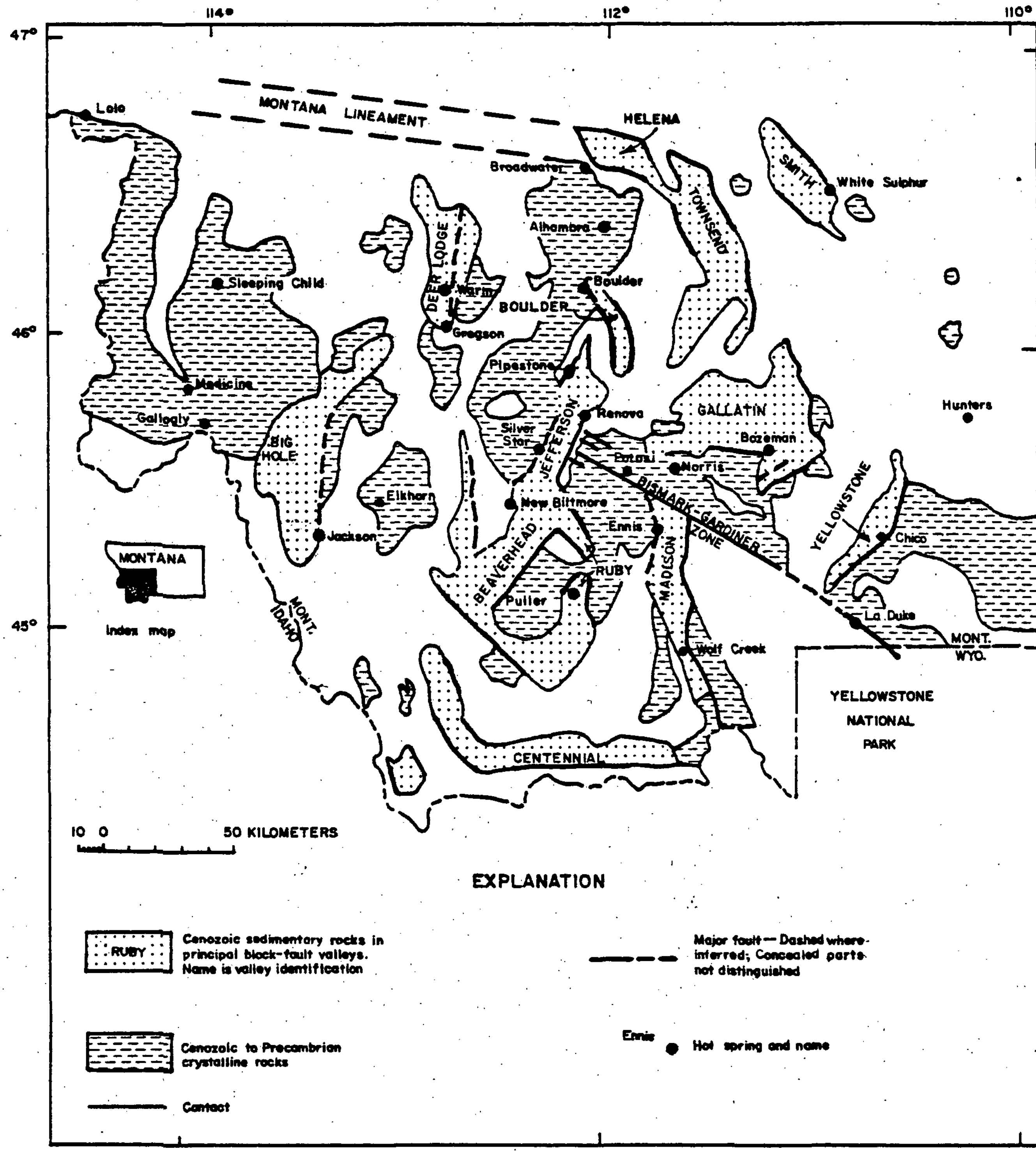

Figure 1.—Geologic setting of hot springs in southwestern Montana.

REGIONAL STRUCTURAL SETTING

Southwestern Montana is underlain primarily by crystalline rocks--Precambrian metamorphic rocks and Cretaceous to Tertiary batholiths. It contains the only hot (above 38°C) springs in the State, with the exception of two (Quinn and Camas) associated with crystalline rock in northwestern Montana. Therefore, knowledge of the structural conditions of this southwestern Montana crystalline province is essential to an understanding of the hot-spring systems.

The crust of the region is considered to be a basement complex of Precambrian W metamorphic rocks intruded in part by Precambrian X gneiss (King, 1976). The basement is overlain in places by infolded Precambrian Y metasedimentary rocks of the Belt Supergroup and Paleozoic to Mesozoic sedimentary and volcanic rocks. During Cretaceous and Tertiary time the basement was intruded by major granitic batholiths (Boulder, Idaho, Philipsburg, Pioneer, Tobacco Root) and smaller intrusives.

The crystalline rock complex is broken by numerous fractures that appear to control occurrences of thermal springs. Halbouty (1976) described two dominant fracture directions in Montana, northwest and northeast, that probably reflect trends of weakness in the basement.

Thrust faulting during Cretaceous to early Tertiary time followed by block faulting further altered the regional fracture pattern. Cenozoic volcanics and sedimentary rocks drape the crystalline-sedimentary crustal blocks or fill the valleys formed within them. Present-day earthquake zones (Smith and Sbar, 1974) tend to follow the north- to northwest-trending pattern of thrust and block faulting, indicating that some faults are still active. Figure 1 illustrates the distribution of crystalline rock, the boundaries of major block-fault valleys as indicated by surface geology and by gravity surveys (Burfeind, 1967), and selected major structural elements in the region.

The predominantly metamorphic-granitic province is bounded on the north by the Montana lineament, a major crustal discontinuity (Weidman, 1965), which separates it from the Precambrian Y folded sedimentary rocks of northwestern Montana. Reynolds (1977) extends the Montana lineament eastward to the White Sulphur Springs area and postulates that late Cenozoic movement along this zone stretched the crust to allow development of the Smith, Townsend, and Helena block-fault valleys. Rhyolitic volcanism took place along or near the lineament intermittently during the period 44.5 to 19.6 m.y. (million years) ago (Williams, 1975; Chadwick, 1978). Thus, the Montana lineament is a major crustal discontinuity that has persisted as a zone of structural adjustment and igneous activity for an extended period.

A major northwest-trending basement zone of weakness, the Bismark-Gardiner fault zone (fig. 1), has been intermittently active during geologic time (Reid, 1957). This zone forms one of the major linear features of the northwestern United States and is at various places along its length a steep reverse fault, a wide crush zone, and a belt of igneous intrusive and volcanic centers. In Montana, the Bismark fault extends southeastward from the Jefferson Valley across the Madison Valley (where it is termed the North Meadow Creek fault). Continuing southeastward, the zone of weakness is buried beneath Eocene volcanic rocks, but reappears east of the Yellowstone Valley as a zone of Eocene volcanic and intrusive centers forming the Western Absaroka Belt (Chadwick, 1970) and the parallel Gardiner fault. Total length from the Jefferson Valley to the southern Absaroka Mountains of Wyoming is about 350 km.

Results of numerous studies suggest that during Cenozoic time southwestern Montana was a region of extensive block faulting that developed graben or half-graben intermontane basins generally elongated north-northwest to north-northeast (see, for example, Pardee, 1950; Kuenzi and Fields, 1971; Scholten and Ramspott, 1968). The province probably reflects extension of basin-and-range type structural activity northward from the Basin and Range province proper (Reynolds and Kleinkopf, 1977).

Kuenzi and Fields (1971) believe that basins began to be delineated in Eocene time and that sedimentation began in late Eocene to early Oligocene time. Oligocene sedimentation was extensive and was accompanied by ash-fall deposits indicating one or more episodes of silicic volcanism. Rhyolitic volcanism within Montana may have furnished much of the ash to the basins. Rhyolitic flows and intrusive bodies in the northern Boulder batholith region are dated as 36 to 37 m.y. (Chadwick, 1978).

Extensive erosion of basin sediments in Oligocene to middle Miocene time produced widespread unconformities (Kuenzi and Fields, 1971). Subsidence and renewed sedimentation during late Miocene to Pliocene time were followed by extensive block faulting and local uplift during late Pliocene-early Pleistocene time. Movement continued on some faults during late Pleistocene and Holocene time as evidenced by displacement of glacial deposits and alluvial fans (Pardee, 1950; Witkind, 1975; Weinheimer, 1979).

REGIONAL STRUCTURAL CONTROLS OF HOT SPRINGS

Most of the hot springs in the region are (a) in deeply fractured crystalline rock terrain, or (b) marginal to or within Cenozoic block-fault valleys underlain and bounded by predominantly crystalline rock. At some of the hot springs both conditions are present. Several hot springs also lie along regional faults or shear zones. Locations of the hot springs in relation to lithology and structure are shown in figure 1, and the geologic settings are summarized in table 1.

Eleven hot springs issue directly from deeply fractured crystalline rock masses--mainly the Boulder, Idaho, or Pioneer batholiths or related intrusive or metamorphic rocks. Broadwater and Lolo Hot Springs are located along the southern part of the Montana lineament system, which forms the north boundary of the Idaho and Boulder batholiths. Alhambra, Norris, Sleeping Child, and other hot springs are situated along minor faults within the crystalline rock. At Elkhorn, Gallogly, and Medicine fractures rather than faults

Table 1. Geologic setting of hot springs

Spring	Location	Water temperature (°C)	Lithology	Structure
Alhambra	NE¼ sec. 16, T. 8 N., R. 3 W.	59	Quartz monzonite and aplite of Boulder batholith.	N. 15° E. fault.
Boulder	SW¼ sec. 10, T. 5 N., R. 4 W.	76	Quartz monzonite of Boulder batholith; basin fill to north and east.	Northwest trending range-front fault may extend northward to area; intersected by N. 50° E. fault and N. 45° W. vein system.
Bozeman	SE¼ sec. 14, T. 2 S., R. 4 E.	55	Basin fill overlying Precambrian metamorphic rocks.	N. 85° W. fault 1 km to southwest; northeast anticlinal axis in Precambrian rocks.
Broadwater	NE¼ sec. 28, T. 10 N., R. 4 W.	66	Quartz monozonite satellitic to Boulder batholith.	Within 5 km of corner of Helena valley where two block faults intersect; strong northeast jointing in quartz monzonite.
Chico	SW¼ sec. 1, T. 6 S., R. 8 E.	46	Madison Limestone adjacent to dacite on east; basin fill to west.	N. 50° E. range-front fault; intersected by minor northwest faults; E-W Mill Creek fault system extends into area from east.
Elkhorn	NE¼ sec. 29, T. 4 S., R. 12 W.	48	Pioneer batholith, near south margin.	N-S to N. 15° W. joints.
Ennis	SE¼ sec. 28, T. 5 S., R. 1 W.	83	Basin fill overlying Precambrian metamorphic rocks.	Possible range-front fault 2 km west; subparallel fault inferred at spring.
Gallogly	NW¼ sec. 15, T. 1 S., R. 19 W.	38	Granitic to gneissic border zone of Idaho batholith.	Uncertain.
Gregson	NW¼ sec. 2, T. 3 N., R. 10 W.	70	Basin fill overlying batholith and latitic volcanic rocks.	Inferred N-S fault extending from Warm Springs area.

Table 1. Geologic setting of hot springs--Continued

Spring	Location	Water temperature (°C)	Lithology	Structure
Hunters	SW¼ sec. 9, T. 1 S., R. 12 E.	60	Cretaceous andesitic sandstone and shale.	N. 50° E. faulted anticlinal axis intersected by N. 70° W. calcite vein-fault system.
Jackson	W½ sec. 25, T. 5 S., R. 15 W.	60	Belt quartzite and argillite beneath basin fill.	Possible N-S range-front fault dropping Big Hole Valley on west.
La Duke	SW¼ sec. 32, T. 8 S., R. 8 E.	68	Madison Limestone dragged against Precambrian metamorphic rocks; valley alluvium to west	N. 45° W. reverse fault intersected by N. 15° E. fault.
Lolo	NE¼ sec. 7 T.11 N., R. 23 W.	44	Quartz monzonite near contact with Belt metasedimentary rocks.	N. 50° E. normal faults; northeast Tertiary dikes.
Medicine	SW¼ sec. 12, T. 1 N., R. 20 W.	47	Granitic rock of Idaho batholith.	Uncertain.
New Biltmore	NW¼ sec. 28, T. 4 S., R. 7 W.	54	Folded Paleozoic-Precambrian sequence beneath thin alluvium.	N. 15° E. possible fault along river.
Norris	E½ sec. 14, T. 3 S., R. 1 W.	50	Precambrian metamorphic rocks intruded by Tobacco Root batholith nearby; thin alluvial cover; Tertiary sedimentary rocks cap nearby hills.	Northeast-trending inferred fault intersected by northwest-plunging anticlinal axis in Precambrian rocks.
Pipestone	NW¼ sec. 28 T. 2 N., R. 5 W.	60	Basin fill overlying Elkhorn Mountains Volcanics near Boulder batholith contact.	N. 40° E. trending step faults cross the hot springs area.

Table 1. Geologic setting of hot springs--Continued

Spring	Location	Water temperature (°C)	Lithology	Structure
Potosi	SW¼ sec. 7, T. 3 S., R. 2 W.	51	Quartz monzonite of Tobacco Root batholith.	N. 50° W. Bismark fault system intersected by N. 15-35° E. zone.
Puller	NE¼ sec. 1, T. 8 S., R. 5 W.	44	Basin fill overlying Precambrian metamorphic rocks.	Spring is 2-4 km east of range-front step faults. Other faults lie within valley.
Renova	S½ sec. 32, T. 1 N., R. 4 W.	50	Folded Belt-Paleozoic sedimentary sequence; basin fill to west.	N-S range-front fault intersected by E-W fault; numerous faults to south and east of spring.
Silver Star	SW¼ sec. 1, T. 2 S., R. 6 W.	73	Boulder batholith intruding Lodgepole Limestone; basin fill to east; thin sediment cover.	Probable range-front fault to east; thrust fault to west. Spring is in line with projection of Bismark fault zone.
Sleeping Child	SE¼ sec. 7, T. 4 N., R. 19 W.	50	Biotite schists and gneisses; Tertiary rhyodacite dike.	Minor N-S faults.
Warm	NE¼ sec. 24, T. 5 N., R. 10 W.	78	Thick basin fill.	Gravity data suggest N-S fault dropping valley on west.
White Sulphur	NW¼ sec. 18, T. 9 N., R. 7 E.	46	Fractured Belt argillite and sandstone basin fill to west.	N. 30° W. range-front fault inferred to cross west of springs site.
Wolf Creek	NW¼ sec. 9, T. 10 S., R. 1 E.	68	Glacial outwash overlying rhyolitic welded tuff, sheet gravels, and Precambrian metamorphic rocks.	N-S fault parallel to range-front fault intersected by west-trending fault.

appear to exert major control on hot spring occurrence (Galloway, 1977).

Thirteen of the hot springs are directly related to Cenozoic block faulting. Some of these (Boulder, Chico, Jackson, Pipestone, Renova, Silver Star, White Sulphur) lie along known or inferred major valley-margin faults. Others (Bozeman, Ennis, Gregson, Puller, Warm Springs, Wolf Creek) are situated on or near known or inferred faults within the valley.

La Duke, New Biltmore, and Silver Star occur along known or inferred faults within crystalline-folded sedimentary rock complexes. New Biltmore is located along an inferred fault that may be part of a range-front fault system extending north to Silver Star. Potosi, La Duke, and Silver Star lie along the Bismarck-Gardiner fault zone or its projection. Hunters is located along a major northeast-trending basement shear zone (Garrett, 1972) covered by younger sedimentary rock.

Robertson, Fournier, and Strong (1976) cited various alinements of the hot springs that may represent deep-seated basement control on the springs. No association with known structures was observed along a cited N. 79 E. alinement for Jackson, Elkhorn, New Biltmore, Potosi, Norris, Bozeman, and Hunters Hot Springs. A cited alinement of approximately N. 10° E. includes New Biltmore, Silver Star, Pipestone, Boulder, and Alhambra Hot Springs. These springs appear to be associated with known and inferred faults along the eastern edge of the Boulder batholith and the western edge of the Boulder and Jefferson River valleys. Other cited alinements are more tenuous.

LOCAL STRUCTURAL CONTROLS OF HOT SPRINGS

Local modifications, superimposed on the regional fractured or block-faulted crystalline rock framework, determine the location of most of the springs. For example, fault intersections may provide pipelike conduits to bring thermal waters to the surface. The major local structural controls are (a) intersection of faults, (b) intersection of fault and anticlinal axis, (c) step faults or other intravalley faults, (d) intersection of fault and sedimentary aquifer, and (e) minor faults and fracture zones in crystalline rock.

Intersection of faults

At Renova Hot Springs, Paleozoic and Precambrian Y sedimentary rocks dip toward the valley along the eastern wall of Jefferson Valley, where they have been downthrown along the west side of the Tobacco Root range-front fault (Berg, 1959). The Paleozoic sequence is cut by a number of subsidiary faults, including some which form a radial pattern as shown in figure 2A. The hot springs issue from Cambrian Meagher Limestone about 30 m north of an east-trending fault contact with the underlying Cambrian Wolsey Shale. A shear zone along the mountain front south of the hot springs may be associated with the range-front fault. O'Haire (1977) suggested that the recharge for the springs originates in the hills to the east and flows downward through

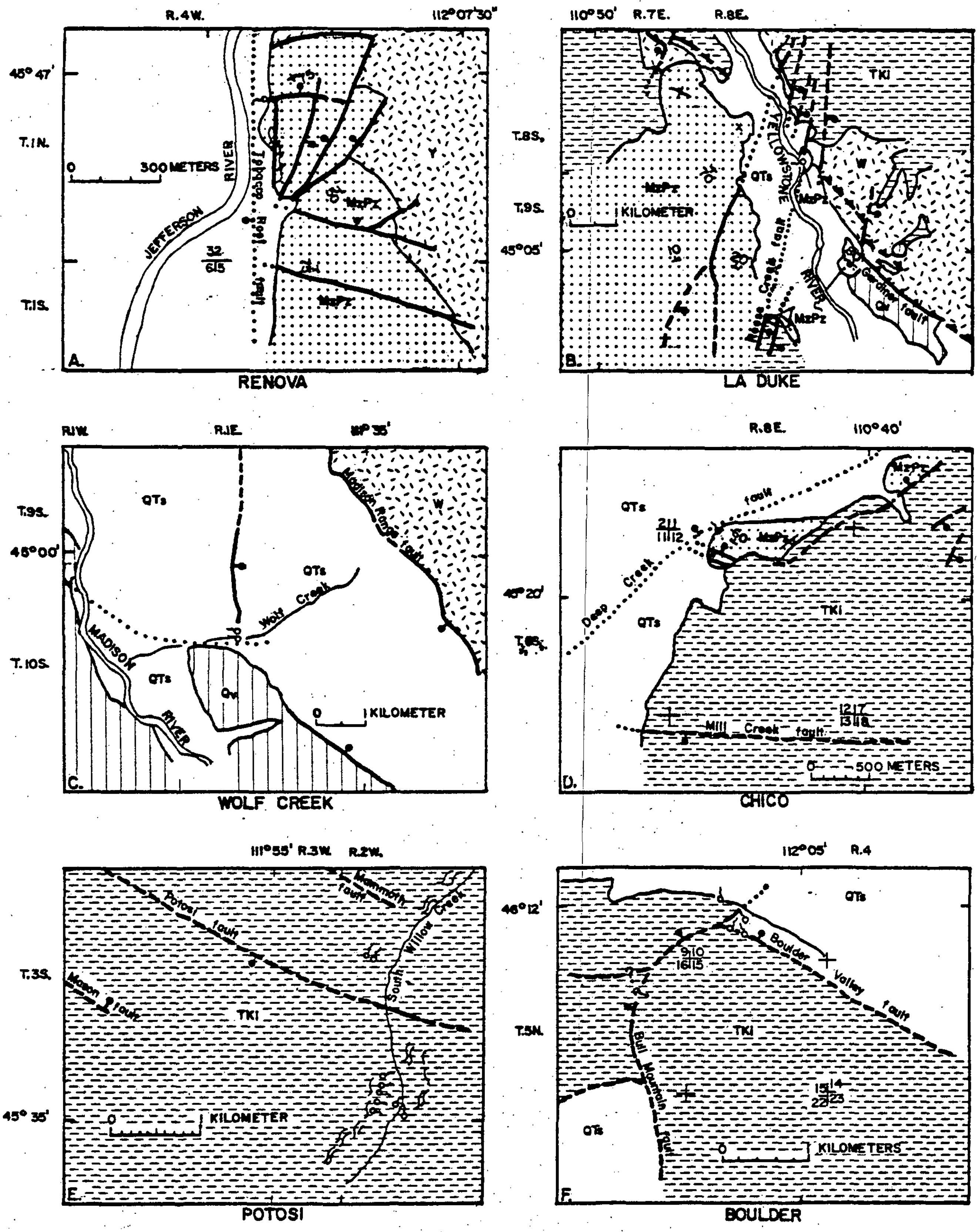

Figure 2.--Hot springs at or near intersections of faults. Maps modified from: A, O'Haire (1977); B, Strusacker (1976); D, Van Voast (1964); E and F, Galloway (1977). Explanation on next page.

EXPLANATION (fig. 2)

Qv	Quaternary volcanic rocks
QTs	Cenozoic sedimentary rocks
TKi	Tertiary and Cretaceous igneous rocks
MzPz	Mesozoic and Paleozoic sedimentary rocks
Y	Precambrian Y rocks
W	Precambrian W rocks
	Contact
?	Normal fault--Dashed where inferred; dotted where concealed; queried where doubtful. Bar and ball on downthrown side
	Reverse fault--sawteeth on upper plate
	Shear zone
20	Strike and dip of beds
80	Strike and dip of overturned beds
	Strike of vertical beds
	Hot spring

the west-dipping Paleozoic beds until it is diverted upward along the intersection of the faults. Some thermal water may reside in the lower part of the valley-fill sequence.

At La Duke Hot Springs, Precambrian metamorphic rocks of the crystalline Beartooth Mountains block on the east are displaced against folded Paleozoic-Mesozoic sedimentary rocks on the west along the Bismark-Gardiner fault zone. The springs are located at the intersection of the N. 45° W. Gardiner reverse fault and the N. 15° E. Reese Creek fault (fig. 2B). Struhsacker (1976) suggests that the thermal water rises along the Gardiner fault where limestone of the Mississippian Madison Group has been dragged to near-vertical orientation. Thermal-water may circulate within fractures and dissolved zones in the limestone to a depth of 3.5 km.

Wolf Creek Hot Springs (fig. 2C) issue from glacial outwash in the Madison Valley that overlies rhyolitic welded tuff (age 1.9 m.y.), Tertiary gravels, and Precambrian metamorphic basement rocks. The principal vent lies about 75 m north of the postulated intersection of a west-trending fault along Wolf Creek that displaces the tuff, and a north-trending fault that displaces the land surface about 1 m. The latter fault is subparallel to the Madison Range fault, which forms the west boundary of the Precambrian massif of the Madison Range 4 km to the east. Resistivity lows were mapped at the principal vent and at a warm seep along the north-trending fault 150 m north-northeast of the hot spring. Thermal water may ascend from the Precambrian rocks along the conduit formed by the intersection of the two faults and then move laterally and upward through the outwash gravels to the spring orifices (Weinheimer, 1979).

Chico Hot Springs (fig. 2D) lie along the Deep Creek (Emigrant) fault (Horberg, 1940) about 300 m north of its intersection with two minor northwest-striking cross faults in Paleozoic beds (see Van Voast, 1964). The springs lie between two branches of the Mill Creek fault, which cuts across the entire western Beartooth structural block and splays as it approaches the Chico area from the east (Foose and others, 1961). Only the southern branch is shown on figure 2D. The Mill Creek fault system may conduct deep-seated thermal waters from the mountains westward toward the Deep Creek fault.

Potosi Hot Springs (fig. 2E) issue from vents along the wall of South Willow Creek canyon where the creek flows through shattered quartz monzonite of the Tobacco Root batholith. Several northwest-trending faults of the Bismark-Gardiner fault zone transect the batholith in this area (Reid, 1957; Smith, 1970). Near the hot springs, one of these faults, a projection of the Potosi fault, intersects a N. 15°-35° E. zone along which the quartz monzonite is extensively sheared (Galloway, 1977). Crosscutting quartz veins are shattered and slightly offset in a stairstep pattern. The shattered zone, parallel to the canyon, has evidently channeled thermal waters to the surface from depth, where they may have traveled along the plane of the Potosi fault. A prominent west-northwest joint system may permit extensive recharge of water into the adjacent uplands of the area (Galloway, 1977).

Boulder Hot Springs (fig. 2F) issue from a zone of silicified, brecciated, and slickensided quartz monzonite. The zone may represent intersection of the northwest extension of a major Cenozoic block fault along the west side of the Boulder Valley (Hamilton and Myers, 1974) and a second, northeast-trending fault zone. A set of calcite-stilbite-chalcedony veins striking N. 45° W. also crosses the hot springs site (Galloway, 1977). Thermal waters may be channeled along one or more of these structures, rising along the intersections.

Intersection of fault and anticlinal axis

Hunters Hot Springs issue from andesitic sandstone and shale assigned to the Livingston Group of Cretaceous age. The springs (fig. 3A) lie near the intersection of the axis of an anticline trending N. 50° E. and a combined calcite vein and fault zone trending N. 70° W. (Chadwick and others, 1978). South of the map area of figure 3A the anticline has a faulted east flank. The anticline has been interpreted to be the near-surface expression of a major basement shear zone (Garrett, 1972). The calcite vein zone is alined with the McLeod anticline southeast of the area shown.

Results of a shallow resistivity survey delineate a resistivity low under Hunters Hot Springs at a depth of 20 m. With increasing depth, the low becomes elongated along N. 60° W. and N. 50°-70° E. directions, approximately parallel to the major structural features described above (Chadwick and others, 1978). The resistivity low probably defines the configuration of thermal water or thermally altered rock. The waters from Cretaceous or older sandstone aquifers may rise from a depth of about 1.5 to 3.5 km (Chadwick and Kaczmarek, 1975) along the limbs of the anticline, subsequently rising to the surface along fault planes or tensional fractures.

Norris Hot Springs (fig. 3B) lie at the intersection of a N. 30° W. trending anticlinal axis in Precambrian gneisses and a probable fault zone (the Warm Springs Creek fault zone) trending northeastward along Warm Springs Creek (Andretta and Alsup, 1960). An apparent resistivity low at 20-100 m depth may delineate the path of rising thermal water along the Warm Springs Creek fault zone (Chadwick and others, 1978). Small outliers of the Tobacco Root batholith crop out about 1 km south of the springs; the northeast edge of the batholith is about 6 km to the southwest.

At Bozeman Hot Springs, an anticlinal axis in Precambrian gneiss projects N. 45° E. toward the springs (fig. 3C). The axis intersects a fault trending N. 85° W. (Mifflin, 1963) that cuts the Precambrian basement about 1 km southwest of the spring. The springs are located on the downthrown northern block where water-well and gravity data suggest that about 0.2 km of valley sediments overlie the basement. A magnetic low north of the fault (Davis and others, 1965) indicates substantial vertical displacement. Thermal waters are postulated to circulate vertically through faulted bedrock, to move laterally through permeable layers in Cenozoic sedimentary rocks consisting predominantly of clay, and then to rise where fractures extend to the surface. Whether the anticlinal axis or associated fractures in

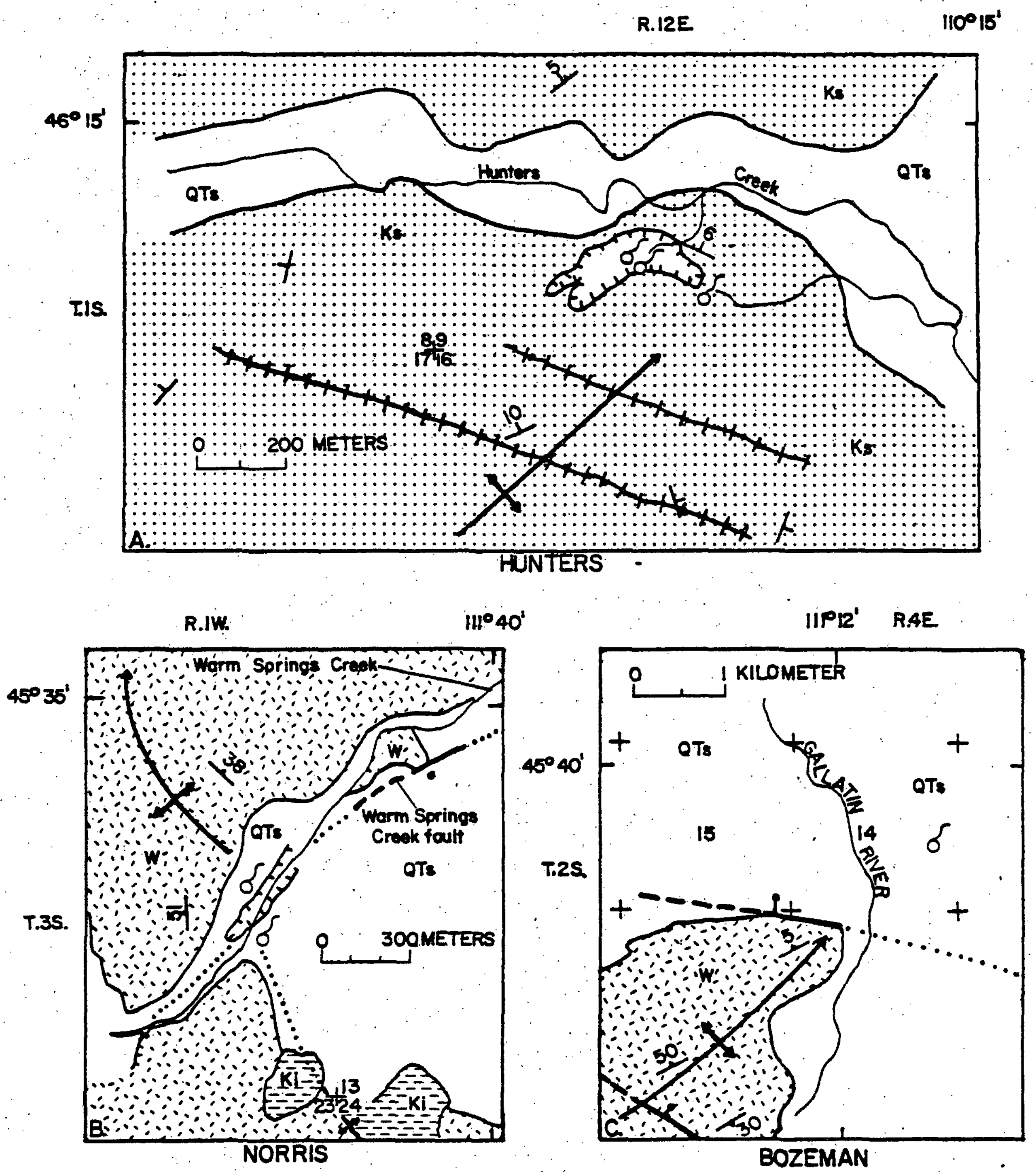

Figure 3.--Hot springs at or near intersections of faults or vein systems and anticlinal axis. Explanation on next page.

EXPLANATION (fig. 3)

QTs	Cenozoic sedimentary rocks
Ki	Cretaceous intrusive rocks
Ks	Cretaceous sedimentary rocks
W	Precambrian W rocks
	Contact
	Normal fault-Dashed where inferred; dotted where concealed; bar and ball on downthrown side
	Anticline with direction of plunge; dotted where concealed
5	Strike and dip of beds
38	Strike and dip of foliation
	Calcite vein
	Resistivity low of less than 50 ohm-meters
	Hot spring

Precambrian gneisses actually control the direction of flow of the thermal waters at Bozeman and Norris is uncertain, but the location of the hot springs along the axes may be more than coincidence.

Step faults or other intravalley faults

Pipestone Hot Springs issue from Cenozoic sedimentary rocks overlying the Cretaceous Elkhorn Mountains Volcanics 2 km east of the eastern contact of the Boulder batholith. The springs lie along a north segment of a postulated range-front fault system along the west side of the Jefferson River valley (fig. 4). This segment consists of two major northeast-trending step faults, each downthrown on the southeast (Prostka, 1966). O'Haire (1977) suggests that the thermal waters rise along the step faults.

Silver Star Hot Springs are located 0.6 km west of the western margin of the Jefferson Valley and about 50 m higher than the valley floor. They issue from a thin sediment layer overlying a contact between quartz monzonite of the Boulder batholith and the Lodgepole Limestone of Mississippian age. A segment of the range-front fault system affecting the Pipestone area is postulated to extend southward past Silver Star to a point about 7 km north of New Biltmore Hot Springs (fig. 4). A sharp gravity gradient at Silver Star (Burfeind, 1967; Abdul-Malik, 1977) and sheared contacts between the Precambrian metamorphic rocks on the west and Cenozoic basin fill sediments on the east (O'Haire, 1977) suggest the presence of a range-front fault east of the hot springs. Silver Star lies near the intersection of the postulated range-front fault and northwestward extension of the Bismark-Gardiner fault zone into the Jefferson River valley (fig. 1).

A resistivity low at an apparent depth of 20 m extends about 200 m east from Silver Star Hot Springs (O'Haire, 1977). At greater depth, the low is located farther to the northeast until, at an apparent depth of 100 m, it is centered 200 m N. 65° E. of the springs and is elongated to the northeast. A deeper survey conducted by Abdul-Malik (1977) also showed a resistivity low having a northeast trend several hundred meters east of the springs. The configuration of the anomaly may describe the upper surface of a zone of thermal water or altered rock east of a range-front fault or a subsidiary fault.

The geologic picture at Silver Star is complicated by the presence of the quartz monzonite-limestone contact beneath the springs and by a major thrust fault in the hills to the west. Whether the thermal water flows into a fault zone from deep valley fill, or from fractures in the crystalline rock of the upthrown block, is unclear.

Near Ennis Hot Springs (fig. 1), the western boundary of the Madison Valley is marked by a shear zone in Precambrian gneiss at the contact with valley sediments. The springs issue from valley fill about 2 km to the east. A resistivity low trending N. 30° E. at the site of the springs may indicate channeling of thermal water up an intravalley step fault and lateral

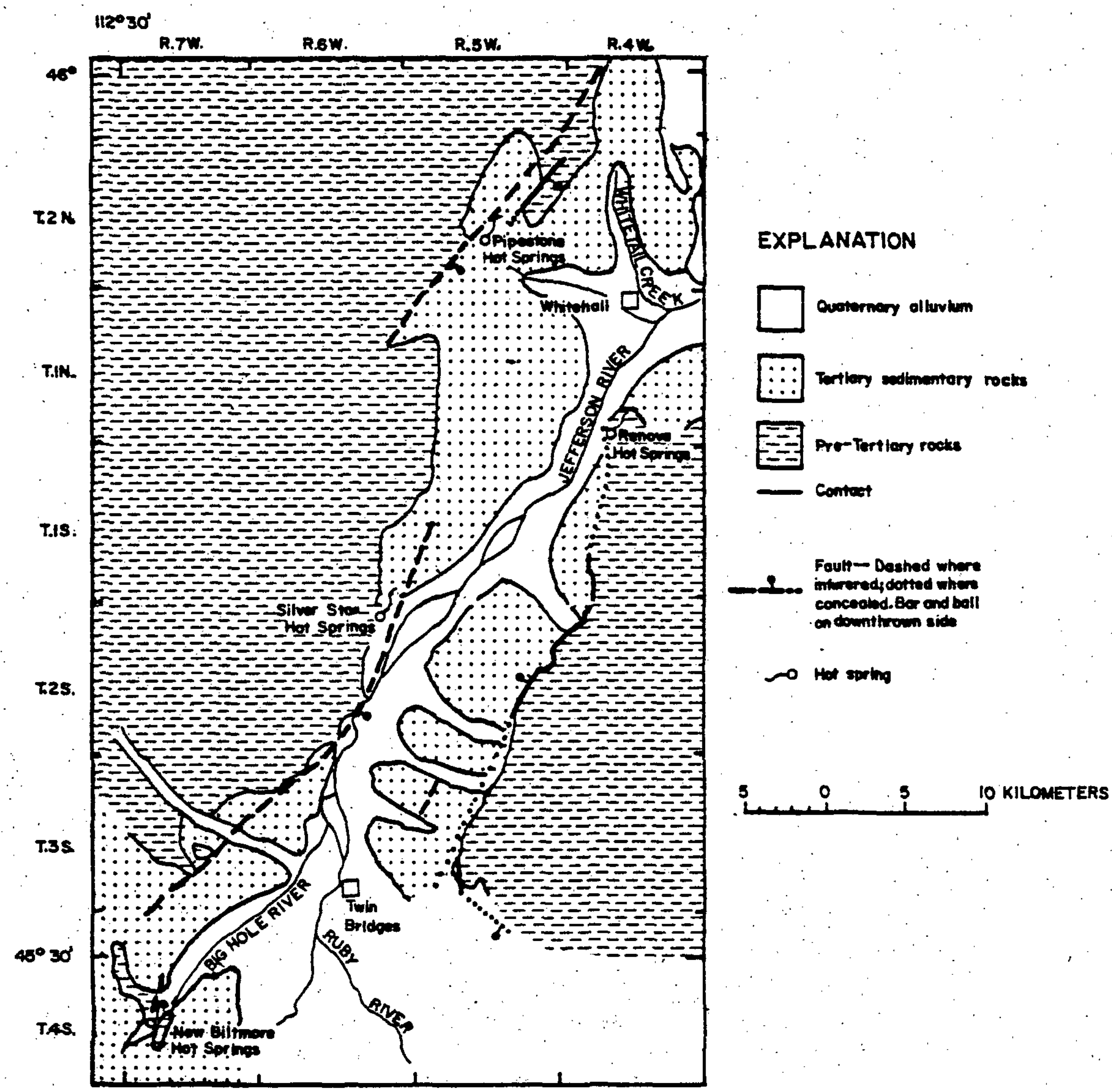

Figure 4.--Range-front faults of the upper Jefferson Valley. Map modified from O'Haire (1977).

spreading through valley sediments along the local ground-water gradient.

Puller Hot Springs are located 2 to 4 km east of two step faults bounding the Ruby Valley on the west. Travertine deposits and one active warm spring (Trudau Lake) are located along several of the numerous intravalley faults in Tertiary sedimentary rocks mapped in this area by Dorr and Wheeler (1964) and Monroe (1976). Whether thermal waters from Precambrian basement rocks at Puller reach the surface along step faults or along one or more intravalley faults connected to the range-front faults at depth is uncertain. Lateral movement of water in the valley fill is probable.

Intersection of fault and sedimentary aquifer

New Biltmore Hot Springs issue from alluvium in an exceptionally narrow and straight reach of the Big Hole River. The valley near the springs appears to be fault-controlled. Evidence includes (1) its straight nature; (2) the presence of numerous minor faults, shear zones, and drag folds in the Precambrian metamorphic and steeply dipping Paleozoic sedimentary rocks that cross the valley; and (3) the presence of a major shear zone near the springs. This shear zone is subparallel and adjacent to the straight reach of the stream and may represent a sliver of the major range-front fault system that appears to extend southward from Silver Star (fig. 4). The hot springs issue from the inferred fault where it displaces steeply dipping Meagher Limestone. Water may travel through the limestone from its outcrop belt in the highlands to the north and rise to the surface where tapped by the fault.

Near White Sulphur Hot Springs (fig. 1), a post-middle Miocene block fault bounds the Smith River valley on the northeast, trends N. 30° W. slightly west of the hot-spring site, and displaces Tertiary sedimentary rocks against Precambrian Belt rocks (M. W. Reynolds, oral commun., 1978). The fault approximately coincides with a northwest-trending zone of high soil temperature (Chadwick and others, 1977) and with seismic discontinuities in near-surface bedrock. A 270 m drillhole penetrated fractured argillite, siltite, and sandstone carrying thermal waters on the upthrown (east) side of the postulated fault. The thermal water possibly ascends fractures parallel to the adjacent range-front fault.

Jackson Hot Springs issue from alluvium near the mouth of Warm Springs Creek along the eastern edge of the upper Big Hole Valley. The straight mountain front to the north suggests the presence of a northerly trending range-front fault that projects through the springs (fig. 1). A resistivity low at the springs has an eastward elongation parallel to the strike of Precambrian Y (Belt) quartzite and argillite beds exposed to the northeast and presumed to underlie the springs. Thermal water may travel along the bedding planes of the fractured Belt rocks after rising along the postulated north-trending range-front fault.

Warm Springs issue from thick Cenozoic sedimentary rocks of the Deer Lodge Valley. A steep gravity gradient (Konizeski and others, 1968) and a steep magnetic gradient (Johnson and others, 1965) suggest the presence of a north-trending fault, downthrown on the west, along the eastern edge of the Deer Lodge Valley structural block (fig. 1). The high concentration of calcium in the waters of the hot springs (Leonard and others, 1978) and a cone of travertine about 13 m high suggest that thermal waters ascend the fault from limestone, either Paleozoic carbonate or Tertiary lime-rich sedimentary rocks.

Minor faults and fracture zones in crystalline rock

Alhambra Hot Springs lie along a minor fault trending N. 15° E. through quartz monzonite and aplite of the Boulder batholith (Smedes, 1966). Resistivity data suggest that thermal water at shallow depth moves from the east and ponds against silicified, brecciated country rock which occurs on the west side of the fault (Galloway, 1977). Leonard and Janzer (1978) suggest that the hot, radioactive water ascends a fractured zone in the fine-grained crystalline rocks adjacent to the fault.

Broadwater Hot Springs issue from a quartz monzonite satellite of the Boulder batholith that intrudes Precambrian Y (Belt) dolomite, limestone, and argillite (Knopf, 1963). Although the trend of the predominant jointing is N. 50° E., resistivity and hydrologic data suggest that the thermal water ascends a north- or northwest-trending fracture zone (Chadwick and others, 1977). The zone is subparallel to the Bald Butte fault along the boundary of the Helena valley several kilometers to the east of the springs.

Lolo Hot Springs (fig. 1) and nearby Granite Hot Springs (not shown) occur along a swarm of northeast-trending faults and Tertiary granite porphyry dikes that cut the Lolo Hot Springs batholith. The flow of thermal water may be controlled by the faults and the contact of the batholith with Belt metamorphic rocks (Galloway, 1977).

Gregson Hot Springs issue from thin alluvium overlying granitic and volcanic rocks in southern Deer Lodge Valley. The north-trending inferred fault passing through Warm Springs may continue southward through the Gregson site and provide a conduit for deep circulation of thermal waters.

Medicine, Sleeping Child, and Gallogly Hot Springs issue from fractured granitic and gneissic rock of the Idaho batholith and the surrounding metamorphosed aureole of Belt rocks. Elkhorn Hot Springs occur in fractured crystalline rock of the Pioneer batholith. No association with major faulting has been established at any of those sites. The hydrothermal system may be controlled entirely by fractures in the crystalline rock.

THERMAL-WATER CIRCULATION

The regional thermal gradient in crystalline rocks of the Boulder batholith and nearby intruded rocks averages about 30°C/km (degrees Celsius per kilometer) (Blackwell, 1969; Blackwell and Robertson, 1973; Leonard and others, 1978). Higher gradients measured in certain localities have been ascribed to hydrothermal convection (Blackwell and Morgan, 1975; Chadwick and others, 1975).

Radiometric dating of 13 surface rock samples by the potassium-argon method as part of this study revealed no igneous activity postdating 19 m.y. except near the southern borders of the State. A basalt at Gardiner, on the north edge of Yellowstone National Park, yields an age of 1.2 ± 0.6 m.y. In the upper Madison River valley, to the northwest, two samples of welded tuff were dated as 2.0 and 1.9 m.y. and probably represent extension of the Huckleberry Ridge tuff of Christiansen and Blank (1972) northward from the Island Park-Yellowstone caldera complex.

Smith and Shaw (1975) calculate that residual magmatic heat from all but the largest magma chambers is probably dissipated within 1 or 2 m.y. Thus, the evidence from surface samples suggests that igneous rocks of the study area are too old to be sources of residual magmatic heat for the hot springs. Instead, normal regional heat apparently is sufficient to maintain the hydrothermal systems without enhancement from cooling igneous bodies.

In the absence of young volcanism, deep circulation of thermal water is required to attain the temperatures measured at the springs (table 1) and those estimated on the basis of chemical geothermometers for the reservoirs. (See, for example, Fournier and others, 1974.) Most of the temperatures are estimated to be between 60° and 120°C (Mariner and others, 1976; Leonard and Janzer, 1978). A normal conductive gradient of 30°C/km, starting with a mean annual surface temperature of 7°C, projects to a reservoir temperature of 60°C at a depth of 1.8 km, and to a temperature of 120°C at a depth of 3.8 km in crystalline rock.

Many of the hot springs issue from faults bounding or within Cenozoic intermontane valleys filled with sediments having low thermal conductivity. Assuming a typical gradient in valley fill of 60°C/km (see Blackwell and Chapman, 1977) and uniform heat flow, the estimated depth at which a temperature of 120°C would prevail at valley-fill-related springs could be 1.9 km rather than 3.8 km in crystalline rock. The estimated depth for a temperature of 60°C would be 0.9 km. At some springs where the water circulates in both crystalline rock and valley fill, the apparent depth of circulation would lie between the estimates for crystalline rock or valley fill alone. Several studies have shown that water can circulate to depths of 2-4 km along major structures similar to those described above (see, for example, Blackwell and Morgan, 1975; Evans, 1966).

Two examples of differing reservoir interpretations are given. At Renova Hot Springs (figs. 1 and 2A), the thickness of adjacent valley fill is about 1.4 km estimated by Burfeind (1967) on the basis of gravity data. If the thermal gradient were 60°C/km, the estimated base temperature of 92°C (Na-K-Ca method) could be acquired by circulation near the basal part of the valley fill or into adjacent crystalline rock. Temperatures of less than 92°C estimated from chalcedony or quartz geothermometers would imply shallower circulation.

At Ennis Hot Springs the thermal waters probably circulate well below the estimated thickness (0.3 km) of the valley fill underlying the springs (Burfeind, 1967) to attain a temperature of at least 83°C (measured at the surface). Higher base temperatures estimated from chemical geothermometers imply circulation to depths greater than 3 km--well within the Precambrian crystalline bedrock.

CONCLUSIONS

Thermal waters that issue as hot (more than 38°C) springs in southwestern Montana appear to be structurally controlled. Deep fractures resulted from movement along major lineament zones and from Cenozoic block faulting, which produced generally north-trending graben and half-graben intermontane basins. Waters circulate to depth along these fractures to acquire their heat. Several hot springs occur along the Bismark-Gardiner fault zone, Montana lineament, and eastern margin of the Boulder batholith. Thirteen hot springs are spatially related to known or inferred Cenozoic block faults in the Deer Lodge, Big Hole, Jefferson, Ruby, Madison, Gallatin, Boulder, Yellowstone, and Smith Valleys.

Geological and geophysical data at various hot springs delineate conduits through which thermal waters may rise from depth. Intersections of faults evidently control waters at Renova, La Duke, Wolf Creek, Chico, Potosi, and Boulder Hot Springs. Faults intersect anticlinal axes at Hunters, Norris, and Bozeman Hot Springs. Pipestone, Silver Star, Ennis, and Puller Hot Springs are found along step faults or other complex range-front fault systems or along inferred intravalley faults. At New Biltmore, White Sulphur, Jackson, and Warm Springs, inferred faults intersect sedimentary aquifers, which may be permeable along bedding planes, fractures, or dissolved zones. Hot springs that issue along fractures or faults with no known cross-structures are Alhambra, Broadwater, Lolo, Gregson, Medicine, Sleeping Child, Gallogly, and Elkhorn. Beneath Wolf Creek, Hunters, Norris, Silver Star, Ennis, Jackson, Alhambra, and Broadwater Hot Springs, shallow resistivity lows may delineate the path of thermal water ascending the shallow conduit systems feeding the hot springs.

Normal regional heat apparently is sufficient to maintain the hydrothermal systems without enhancement from cooling igneous bodies. The thermal gradient normally is higher in low thermal conductivity sediments of the block-fault valleys than the 30°C/km average for crystalline rock. To attain the reservoir temperatures indicated by chemical geothermometers (generally 60°-120°C), waters would have to circulate to depths of about 2 to 4 km in crystalline rock and about 1 to 2 km in valley-fill sediments. The Renova Hot Springs hydrothermal system exemplifies circulation near the basal part of the valley fill or into adjacent crystalline rock. At Ennis Hot Springs, in contrast, waters appear to circulate well below the valley fill.

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