Geology and geothermal system near Jackson, Beaverhead County, Montana
by Geoffrey Alan Black

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Sciences
Montana State University

Abstract:
The Jackson, Montana area contains a thick section of Precambrian and a thinner section of Cambrian sedimentary rocks. These are overlain by poorly consolidated Tertiary and Quaternary sediments. The Precambrian sequence consists of clastic rocks of the Missoula Group, McNamara and Garnet Range Formations. These are unconformably overlain by Cambrian sandstones, shales, and dolomites of the Flathead, Wolsey, Meagher, Pilgrim, and Snowy Range Formations.

A regionally extensive thrust fault, the Jackson thrust fault, trends roughly northward through the study area. Movement along this fault placed rocks of the Precambrian McNamara Formation onto the Cambrian sequence. The Jackson thrust fault is interpreted to be an imbricate fault of a large thrust fault that underlies the study area. Post-thrusting extensional motion occurred along this fault as well as on smaller normal faults that parallel the trace of the Jackson thrust fault.

Hydrothermal circulation of the Jardine Hot Spring system takes place within a regime of regional heat flow. Geochemical evidence indicates that the reservoir temperature of the system is probably between IOO0C and 145°C. Circulation depths are estimated' to be between 9,600 and 14,000 feet. Discharge most likely occurs along an inferred normal fault that connects Jardine Hot Spring with a small group of thermal springs northwest of Jackson. Recharge probably takes place along either the trace of the Jackson thrust fault or along an inferred fault on the western margin of the Big Hole Basin. The lack of other thermal springs in the basin supports a local hydrothermal model with both major recharge and discharge occuring in the vicinity of Jardine Hot Spring.

The surface geothermal resource of the study area is capable of supporting an economically viable district heating project for the town of Jackson. The estimated subsurface resource is suitable for a small scale power generation project with a production of at least 400 kilowatts of electricity.

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APPROVAL

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Geoffrey Alan Black

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

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ABSTRACT

The Jackson, Montana area contains a thick section of Precambrian and a thinner section of Cambrian sedimentary rocks. These are overlain by poorly consolidated Tertiary and Quaternary sediments. The Precambrian sequence consists of clastic rocks of the Missoula Group, McNamara and Garnet Range Formations. These are unconformably overlain by Cambrian sandstones, shales, and dolomites of the Flathead, Wolsey, Meagher, Pilgrim, and Snowy Range Formations.

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CHAPTER 1

INTRODUCTION

Purpose of Study

The purpose of this study is to describe the geology near the town of Jackson in southwestern Montana and to establish a circulation model for the hydrothermal system in the area. Conclusions are based on detailed geologic mapping of the structural and lithologic controls on the system and on geochemical surveys of the thermal water, groundwater, and surface water in the area. A second objective is to assess the economic potential of the hydrothermal system by means of a preliminary analysis of geothermal space heating and electrical generation.

The geothermal system at Jackson was selected for this study because of the large discharge of Jardine Hot Spring relative to other thermal springs in Montana, the presence of geologic structures that are consistent with major recharge and discharge channels, and geochemical data that indicate above-boiling subsurface temperatures. In addition, the proximity of the system to the town of Jackson favors the economic utilization of the geothermal resource.

Location

The study area is located near the southeastern margin of the upper Big Hole Basin in southwestern Montana near the town of Jackson (Figure 1). The crescent-shaped valley of the upper Big Hole River is one of the many intermontane basins that characterize southwestern Montana. The basin is approximately 50 miles in length and trends generally north-south. It is bounded on the east by the Pioneer Mountains which are composed primarily of granitic intrusives of Late Cretaceous to Early Tertiary age. The basin's western boundary is the northern Beaverhead Range, composed principally of allochthonous Precambrian sedimentary rocks.

The hot spring itself is within the flood plain of Warm Springs Creek near its confluence with the Big Hole River (Figure 2). The spring discharges from Quaternary alluvium and Tertiary valley fill sediments. In addition, a small group of warm springs, herein termed Lapham Warm Springs, issue two kilometers (1.25 miles) northwest of Jardine Hot Spring and appear to be part of the same hydrothermal system.

Procedure

Field investigations of the study area were conducted during the 1980 and 1981 field seasons. Field work involved geologic mapping of the fifty square miles surrounding Jardine Hot Spring using field reconnaissance and air photographs. Geologic data was plotted on 1:24,000 scale topographic maps.

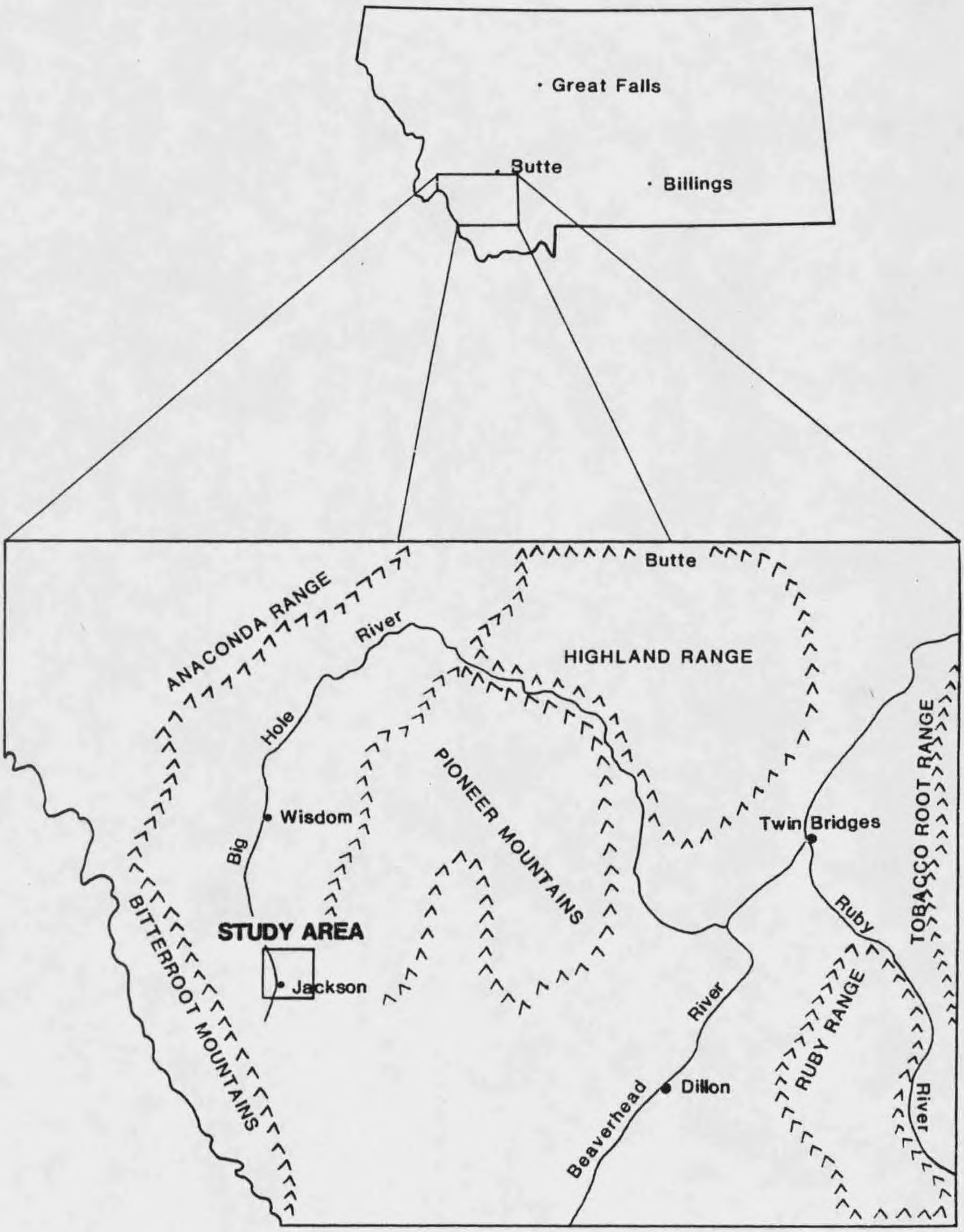

Figure 1. Index map showing location of study area.

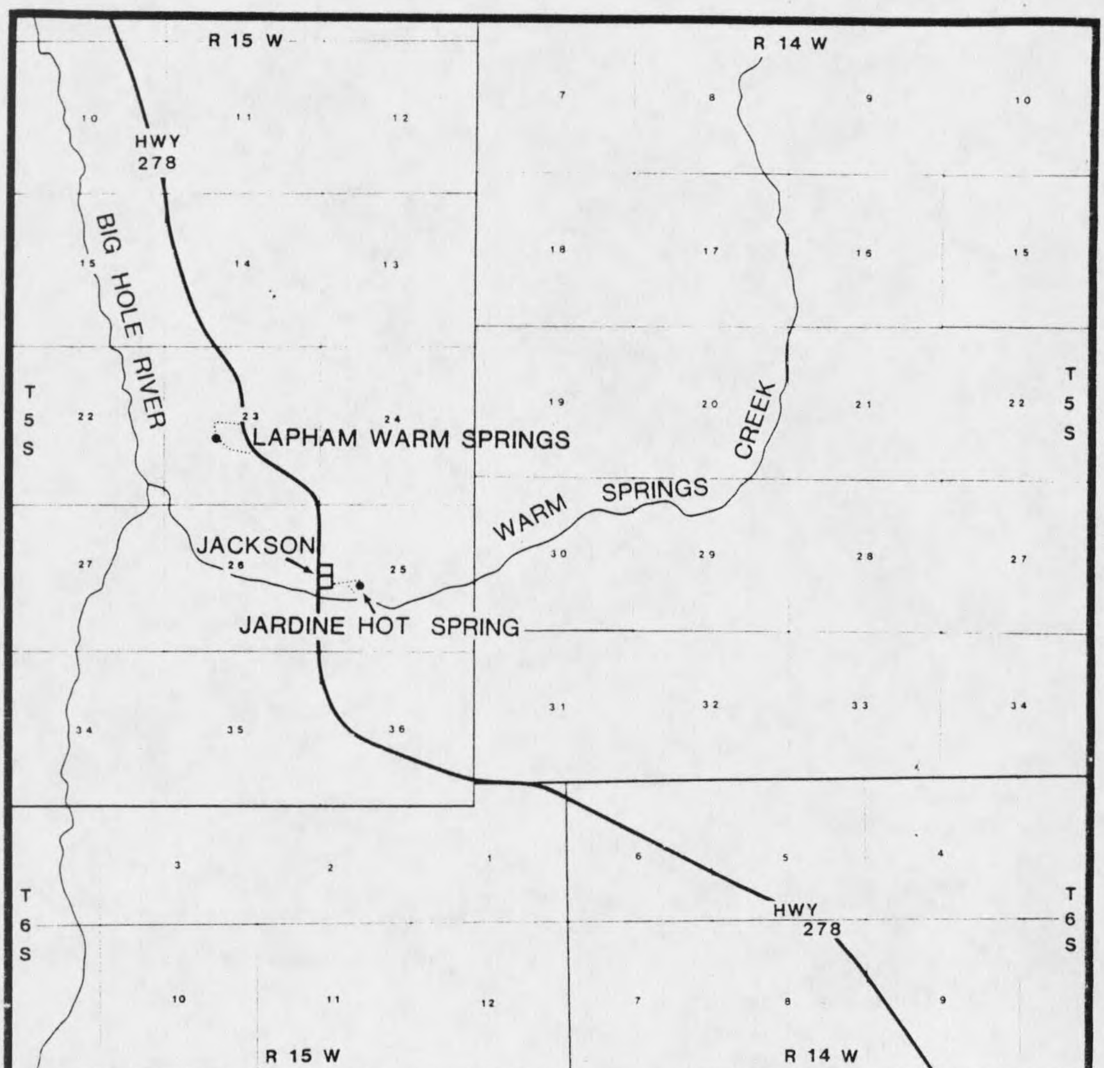

Figure 2. Map of study area.

Thin sections from all major lithologic units in the area were used as an aid to formation description and identification and to estimate intergranular porosity.

Water samples were collected from Jardine Hot Spring and Lapham Warm Springs, as well as from four cold wells, one cold spring, and one stream within the study area. Samples were collected during two sampling periods, in May and November 1981, when groundwater levels were high and low, respectively. Raw, filtered, and filtered-acidized samples were analyzed by the Montana Bureau of Mines and Geology in Butte, Montana.

Previous Investigations

Little work has been done on the geology of the Big Hole Basin. Perry (1934) described the physiography of the basin in a reconnaissance manner and discussed the groundwater supply of the area. Alden (1953) briefly described some of the Tertiary and glacial geology of the valley as part of a regional study. More recently, Hanneman and Nichols (1981) described the late Tertiary sedimentary rocks along the northern and northeastern margins of the basin.

The western Pioneer Mountains have only recently been studied in any detail. Snee and others (1981) compared the age and geochemistry of the plutonic rocks of the west Pioneers to those of the east Pioneers. The area is also included in a current mapping project of the Dillion 1° x 2° sheet by the U. S. Geological Survey.

The Jackson, Montana area is included in a regional Bouger gravity survey of Montana by Bonini and Smith (1974) and an aeromagnetic map of the Pioneer Mountains area by the U. S. Geologic Survey (1979). Shallow resistivity studies of the area immediately surrounding the hot spring at Jackson were conducted by Chadwick, Kaczmarek, and Weinheimer (1977, unpublished data).

Jardine Hot Spring is included in regional hot spring studies by Chadwick and Kaczmarek (1975), Chadwick and Leonard (1979), Nathenson (1981), and Robertson and others (1976).

CHAPTER 2

REGIONAL GEOLOGY

This chapter places the study area within a regional structural and stratigraphic framework. The tectonic history of southwestern Montana is briefly reviewed and the stratigraphy of the Pioneer Mountains region examined. Emphasis is on stratigraphic units of the region that are not found within the study area. A more detailed treatment of both the stratigraphy and structure of the study area is found in Chapter 3.

Regional Tectonic History

The rocks and structural features of the study area record three major stages in the geologic development of southwestern Montana: (1) the origin and evolution of the Cordilleran miogeocline as a major depositional basin in which the Late Precambrian and Paleozoic rocks of the region were deposited, (2) the orogenic disruption of the miogeocline by folding and thrust faulting during the Mesozoic Era and (3) the extension and normal faulting of the region during the Late Tertiary Period through the present.

Late Precambrian and Paleozoic

The earliest record of marine sedimentation in southwestern Montana is contained within the Late Precambrian Belt Supergroup. Sedimentation took place primarily within an east-trending, fault-controlled basin (Winston, 1973; Harrison and others, 1974). These rocks are chiefly shallow water deposits that contain fine-grained clastics and some limestones. In the study area, only the uppermost Belt rocks, the Missoula Group, are exposed.

During the Paleozoic, the study area was positioned along the hinge zone between a major depositional basin, the Cordilleran miogeocline, to the west and a cratonic platform to the east (Figure 3). This relationship persisted throughout the Paleozoic Era into the early Mesozoic Era. Sediments in the miogeocline were deposited as a westward-thickening wedge along a passive, Atlantic-type continental margin (Stewart and Poole, 1974; Dickinson, 1977). Within the study area, sedimentation patterns were determined by the relative emergence and subsidence of the stable shelf and miogeocline. Middle and Upper Cambrian sandstones and dolomites are the only Paleozoic rocks exposed in the study area.

Mesozoic

During the Mesozoic, the Cordilleran miogeocline became disrupted. The thick wedge of Phanerozoic miogeoclinal sedimentary rocks was affected by generally east-directed folding and thrust faulting. Major movement on thrust faults occurred during the Sevier orogeny as early as Late Jurassic time and culminated in the Cretaceous Period (Armstrong, 1968) with local thrusting likely continuing into the Early Eocene Series.

The general features of the thrust belt in southwestern Montana have been outlined by Ruppel and others (1981) and by Ruppel (1982),

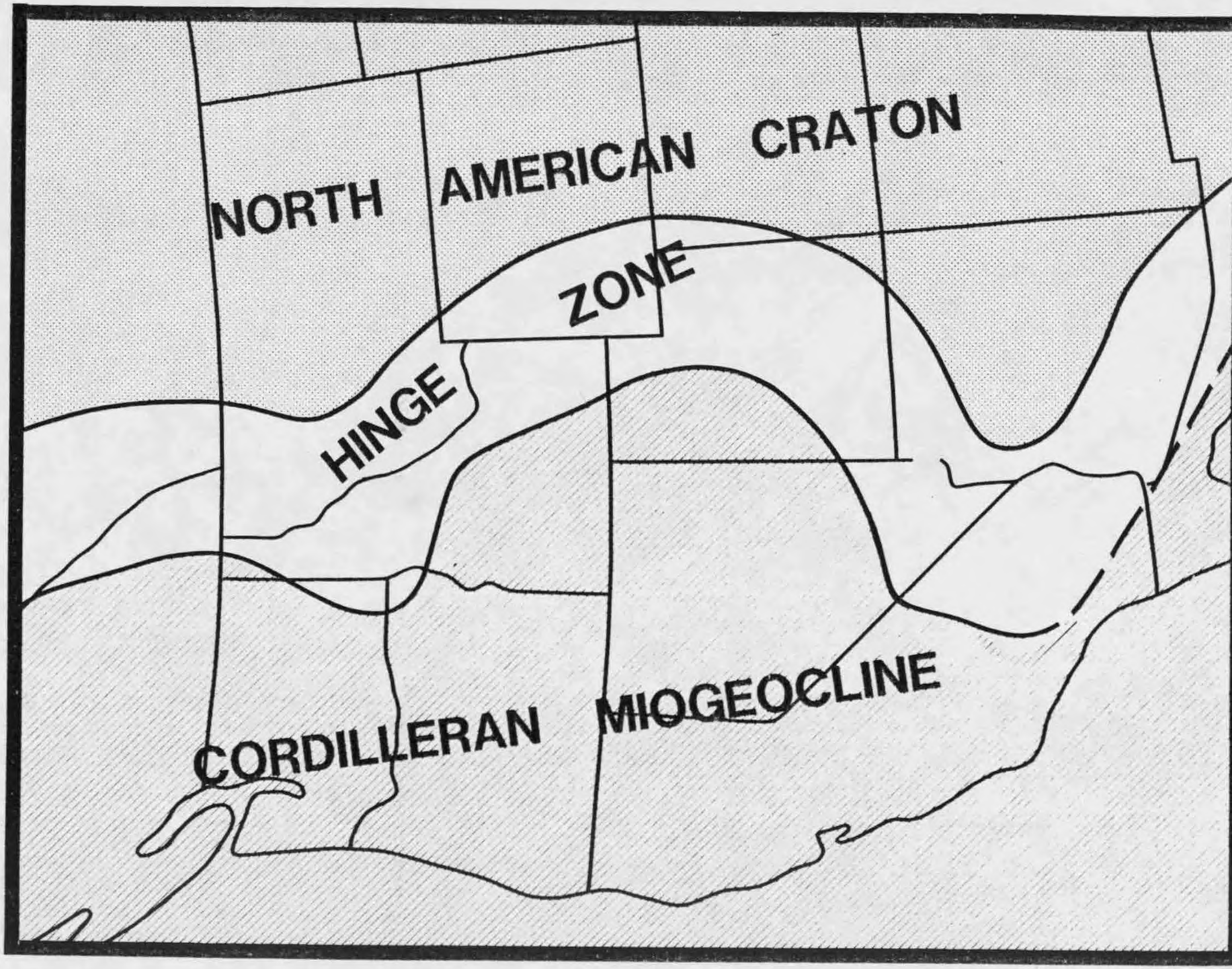

Figure 3. Map showing relative positions of North American craton, Cordilleran miogeocline, and hinge zone.

who describe the region as consisting of two major overlapping thrust plates, the Medicine Lodge plate and the Grasshopper plate. The study area lies just east of the leading edge of the Medicine Lodge plate and is within the Grasshopper plate. These plates are bounded on the east by the "frontal fold and thrust zone" which forms the leading edge of the thrust belt in southwestern Montana (Figure 4). Both the Medicine Lodge and Grasshopper plates are underlain by a basal decollement zone along which east-directed movement took place (Ruppel and others, 1981).

The Medicine Lodge plate, whose leading edge crosses the northern Beaverhead Mountains on the west side of the Big Hole Basin, is described by Ruppel and others (1981) as the dominant thrust plate in southwestern Montana and east-central Idaho. The trace of the Medicine Lodge thrust fault has been traced northward from the north flank of the Snake River Plain for more than 200 km to the west side of the Big Hole Basin, and may continue northwestward to the eastern edge of the Idaho batholith (Ruppel and others, 1981). Ruppel (1978, p. 21) states:

> As a result of movement along the Medicine Lodge fault, Precambrian and Paleozoic sedimentary rocks deposited in the Cordilleran miogeocline in western Idaho have been telescoped and transported as much as 160 km eastward, to rest on rocks of similar age deposited in a marine embayment or seaway in southwestern Montana.

The study area lies within the Grasshopper plate of Ruppel and others (1981), who describe the plate as being structurally beneath the Medicine Lodge plate and consisting of an entirely allochthonous series of Missoula Group and Lower Paleozoic sedimentary rocks. This agrees

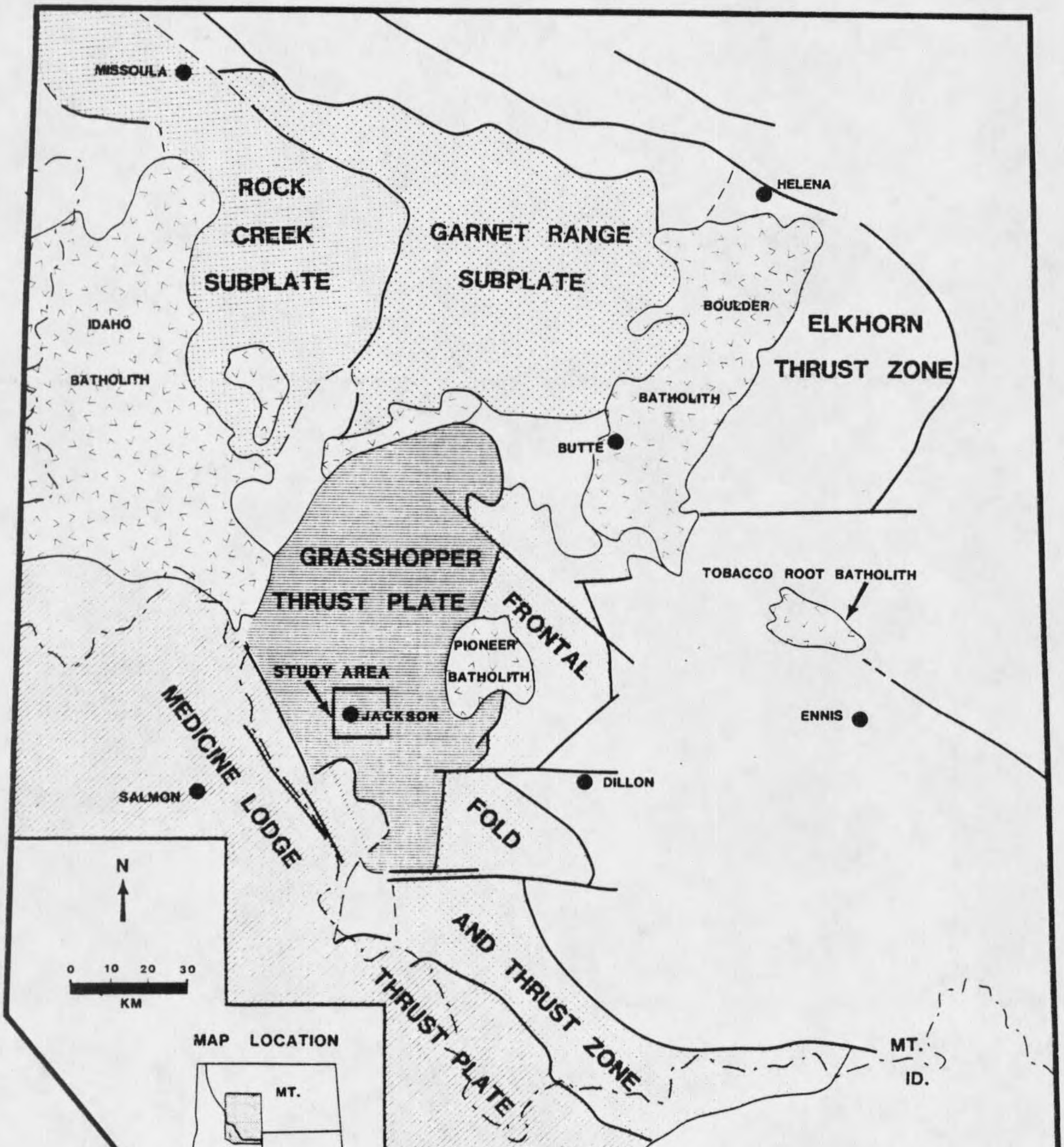

Figure 4. Map of major southwest Montana thrust plates. (Modified from Ruppel and others, 1981)

with the present study since thrust faulted and brecciated Missoula Group rocks overlie Cambrian rocks within the study area and the presence of a regional decollement beneath the study area is suggested (see Chapter 3).

Pioneer Batholith. The complex Pioneer Batholith was intruded into the Grasshopper plate following regional thrusting. This batholith contains several plutons which exhibit age-related compositional changes that range from granodiorite plutons (67-72 m.y.b.p.) to granite plutons (60-62 m.y.b.p.) (Berger and others, 1981). Snee and others (1981) note that radiometric dates from granodiorite plutons in the west Pioneer Mounains are younger than those in the eastern part of the range. They speculate that the younger ages of the western plutons may reflect prolonged cooling related to tectonic loading from pre-emplacement thrusting in the western part of the Pioneers. Alternatively, they note that the age differences of the eastern and western Pioneer Mountains could result from post-emplacement downfaulting of the western Pioneers along north and northeast-trending high angle faults that exposed deeper portions of the batholith to the east. A previously unmapped high-angle fault consistent with the trends described by Snee and others (1981) was mapped in the study area. This fault, referred to as the Warm Springs Creek fault in this study, probably connects with the high angle faults mapped further to the east (Figure 5). This fault is described in more detail in Chapter 3.

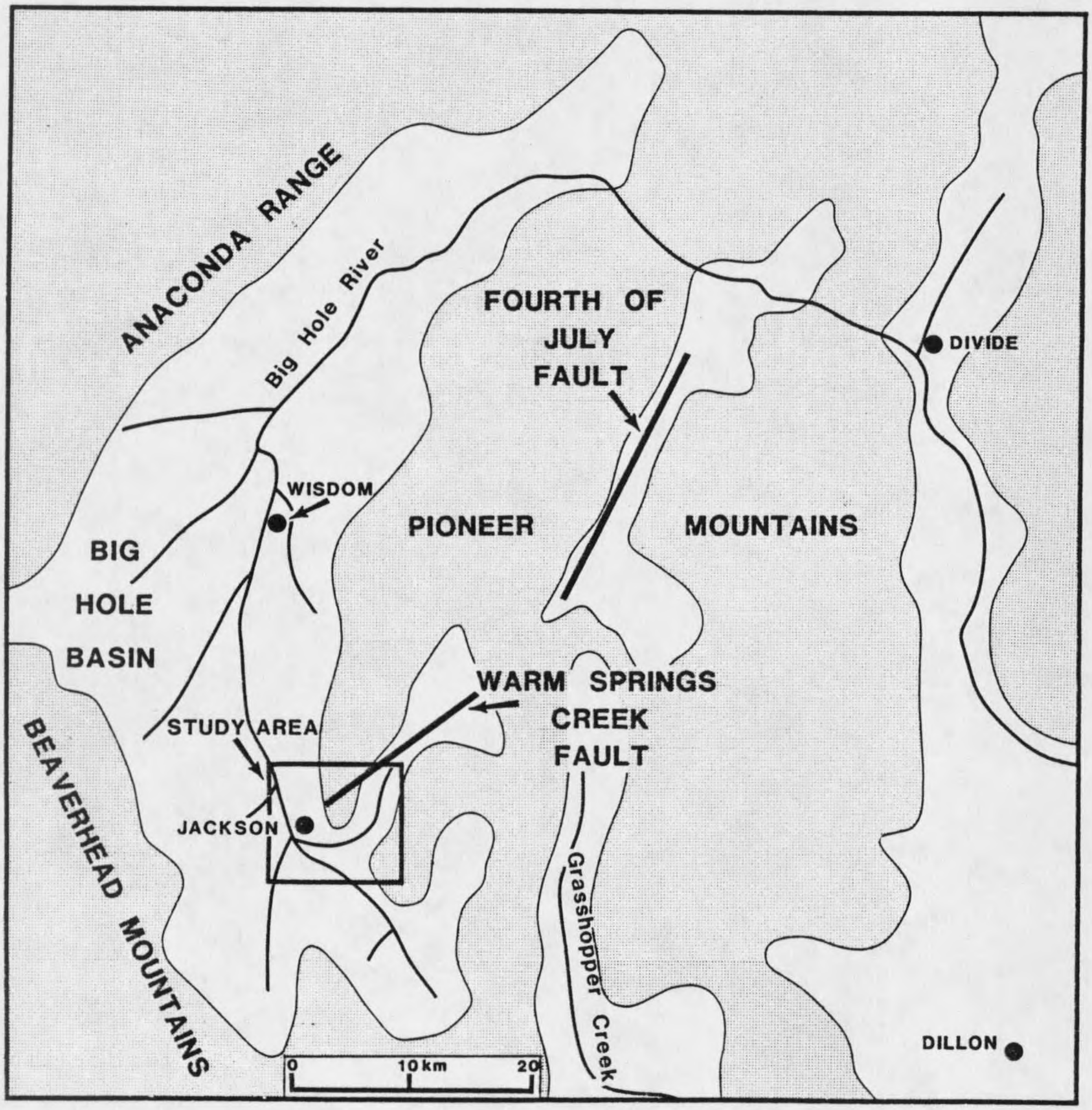

Figure 5. Map of Pioneer Mountains region showing location of high angle faults.

Cenozoic

The most recent tectonic episode affecting the study area involves Late Tertiary through Recent extensional faulting in southwestern Montana. Many of the normal faults within the thrust belt may actually flatten with depth and sole into older thrust ramps (Royse and others, 1975; Allmendinger and others, 1983). Many of the intermontane basins of western Montana were probably formed by downfaulting along these listric normal faults (Bally and others, 1966). Hamblin (1965) has shown that such faults often produce rotational motion in the hanging wall, causing strata to dip toward the fault plane. West dipping listric normal faulting probably occurred along the eastern margin of the Big Hole Basin as evidenced by the eastward-tilted and beveled Tertiary beds approximately twenty miles north of the study area near Wisdom, Montana. Eastward tilted Tertiary beds may also be observed south of the study area.

Regional Stratigraphy

Few studies that deal with the stratigraphy of the region surrounding the study area have been performed. What follows has been adopted primarily from studies of mining districts in the Pioneer Mountains area (cited in the text) as well as more regional studies by Hanson (1952), Moritz (1951), Sloss and Moritz (1951), and Peterson (1981). The units found within the study area, Precambrian Missoula Group and Cambrian sedimentary rocks, are treated only briefly here, as they are examined in detail in Chapter 3.

Precambrian Belt

The Precambrian Y Belt Supergroup crops out over much of western Montana and may attain a thickness of over 40,000 feet. Belt strata consist chiefly of fine-grained clastics that have undergone low-grade metamorphism to quartzites, siltites, and argillites. Ripple marks, dessication cracks, and raindrop imprints that are found throughout the sequence indicate shallow water deposition throughout much of Belt time.

Belt rocks in the vicinity of the study area appear to represent only the Upper Belt Missoula Group, which has been little described in southwestern Montana. This paucity of comparative data, in conjuction with rapid changes in thickness and lithology, make correlations of Missoula Group units in the area uncertain.

Paleozoic

The Paleozoic sequence within the study area contains only the Cambrian Flathead through Snowy Range Formations (see Figure 6). Most formations that constitute the cratonic sequence, however, are present near the study area in the Pioneer Mountains region (e.g. Karlstrom, 1948; Lowell, 1965; Perry, 1934).

Lower Paleozoic sedimentation is represented in the region by the basal transgressive Flathead Sandstone and Wolsey Shale. In areas of Belt deposition, the Flathead Sandstone unconformably overlies Precambrian Missoula Group rocks, as it does in the study area. In areas that lack Belt rocks, such as the Archean terrain east of Dillon, the Flathead is unconformable with Archean basement rocks. Following deposition of the Wolsey Shale, clastic sedimentation was followed by widespread carbonate deposition. Such Middle and Upper Cambrian carbonates are represented in the region by the Meagher and Pilgrim Formation. The more clastic Snowy Range Formation, and especially its conglomeratic facies, may represent a reworking of the previously deposited clastic and carbonate sediments during a rapid advance of the Late Cambrian sea (Lochman-Balk, 1971).

Shallow subtidal and intertidal dolomitized limestones were widely deposited in the region during Ordovician through Middle Silurian time, but were largely eroded during Late Silurian-Early Devonian time (Ross, 1977; Poole and others, 1977). Where it is present in the southern and eastern Pioneer Mountains, the Devonian Jefferson Formation unconformably overlies either the Cambrian Snowy Range or Pilgrim Formations and represents a widespread Devonian transgression. The Late Devonian Antler orogeny apparently did not affect depositional patterns in the vicinity of the study area. Myers (1952) for instance describes a conformable sequence of Upper Devonian Three Forks Formation with Lower Mississippian Lodgepole and Mission Canyon Formations limestones in the Argenta area.

The closest outcrop of the Mississippian Madison Group to the study area is in the Polaris area (Davis, 1980) and is similar to the Madison Group at the Hecla (Karlstrom, 1948), Argenta (Myers, 1952), and Bannack (Lowell, 1965) mining districts. Increasing amounts of evaporites near the top of the Madison Group in the region record a gradual regression of the Madison sea.

The initial phase of Permo-Pennsylvanian sedimentation is represented in the region by the Amsden Formation which unconformably overlies the upper Madison Group (Karlstrom, 1948; Lowell, 1965; Myers, 1952). The widespread Quadrant Sandstone conformably overlies the Amsden Formation and is similar to clean quartz sandstones associated with shelf and platform areas in Montana and Wyoming. Although Shennon (1931) does not recognize the Permian Phosphoria Formation in the Argenta area, Myers (1952) identifies a dark, cherty unit that conformably overlies the Quadrant at Argenta as the Phosphoria. Lowell (1965) also describes a section of Phosphoria, along with the Park City and Shedhorn Formations, in the Bannack area. Karlstrom (1948) notes the presence of the Phosphoria at both Melrose and Hecla.

Mesozoic

Mesozoic sedimentary rocks in the Pioneer Mountains region reflect tectonism associated with the development of the thrust belt. Triassic rocks represent the final deposits of the Cordilleran miogeocline. In the Pioneer Mountains region, the Triassic Dinwoody Formation was deposited with apparent conformity on the Phosphoria Formation and is overlain by the redbeds of the Woodside Formation. Myers (1952) notes that the Woodside, Dinwoody, and Upper Phosphoria Formations in the Argenta area have been truncated by post-Triassic erosion. Such erosion is consistent with depositional patterns elsewhere in the state. Apparently, the regressive sequence in the Woodside Formation was followed by a long period of erosion possibly as a result of orogenic uplift in the westernmost portions of the Cordilleran miogeocline.

Unconformably overlying the Woodside in the Pioneer Mountains region is a thick sequence dominated by clastic facies with subordinate limestone. At Argenta, the base of these beds, which consists of fine-grained clastics and some volcanic material, has been tentatively correlated with the Morrison Formation (Myers, 1952). Lowell (1965) also indicates that a narrow bed of the Morrison Formation may be present at Bannack. Elsewhere in the Pioneer region, however, the basal conglomerate of the Kootenai Formation overlies the erosional surface developed on the Woodside Formation. Near the top of the Kootenai, the characteristic "gastropod limestone" is present, as it is throughout southwestern Montana.

The final Mesozoic deposits in the Pioneer Mountains region are represented by the predominately nonmarine, volcanic-rich, clastic deposits of the Colorado Group. Over much of the region, the Colorado Group represents the youngest sedimentary rocks preserved (Karlstrom, 1948). Following Colorado Group deposition, this part of southwestern Montana became increasingly disrupted as tectonic activity progressed eastward.

Late Mesozoic-Cenozoic

Near the southern end of the Pioneer Mountains, post-Colorado Group deposition is represented by the thick, synorogenic Beaverhead Formation. This formation consists of approximately 10,000-15,000 feet of synorogenic fluvial sandstone and boulder conglomerate that was shed from the tectonic belt as alluvial fan deposits (Ryder, 1968; Ryder and Scholten, 1973). As with most synorogenic deposits, the Beaverhead Formation exhibits an inverted stratigraphy. Lithologic units progressively higher within the formation correspond to progressively older stratigraphic units in the source areas of the formation. The Beaverhead Formation lies on folded and faulted Mesozoic and upper Paleozoic rocks in the southern Pioneer Mountains area and is itself involved in folding and thrusting locally (Lowell, 1965). Deposition may have begun as early as Lower Cretaceous and continued locally into Paleocene or Eocene time (Ryder and Scholten, 1973).

Most Tertiary deposits in southwestern Montana are fluvial and lacustrine sediments assigned to the Bozeman Group of Robinson (1963). The age and names of units within the Bozeman Group differ slightly from basin to basin in southwestern Montana, but are generally assigned to two unconformable, lithologically distinct sequences, the Renova and Sixmile Creek Formations (Kuenzi and Fields, 1971; Robinson, 1967). The Renova Formation, dominated by fine-grained clastics largely derived from volcanic ash, was deposited between approximately Eocene and Early Miocene time (Kuenzi and Fields, 1971; Thompson and others, 1981). The coarse-grained Sixmile Creek Formation, that unconformably overlies the Renova Formation, was deposited between approximately Middle Miocene and Early Pliocene time.

Quaternary

Post-Tertiary rocks in the Pioneer Mountains region consist of unconsolidated sediments deposited during periods of glacial and fluvial erosion of the highlands. Although glacial deposits are mostly restricted to the higher elevations in the Pioneer Mountains, morainal deposits are present in the valley of Warm Springs Creek approximately eight miles northeast of the study area (L. Snee, 1981, personal communication). Glaciers extended into the western Big Hole Valley from the Beaverhead Mountains, as evidenced by well developed moraines along the basin margin.

CHAPTER 3

LOCAL GEOLOGY

The geology of the study area is examined in detail in this chapter. Because a knowledge of local stratigraphic units is a prerequisite to the understanding of the structure of any area, the stratigraphy of the Jackson area is presented first. A section on the local structure follows in which the types of faulting found within the study area are discussed.

Stratigraphy

The stratigraphic sequence within the study area consists of Precambrian and Cambrian sedimentary rocks which are unconformably overlain by poorly consolidated Tertiary and Quaternary sediments. The Precambrian rocks consist of Missoula Group clastic rocks which form two distinct packages in the study area (Figure 6). The stratigraphically lower unit is found in the hanging wall of a southwest-dipping thrust fault, herein termed the Jackson thrust fault, that trends northwest through the study area (Plate 1). This lower unit is assigned to the McNamara Formation of the Missoula Group. The stratigraphically higher unit, assigned to the Garnet Range Formation, is found east of the Jackson thrust fault and is unconformably overlain by Cambrian rocks.

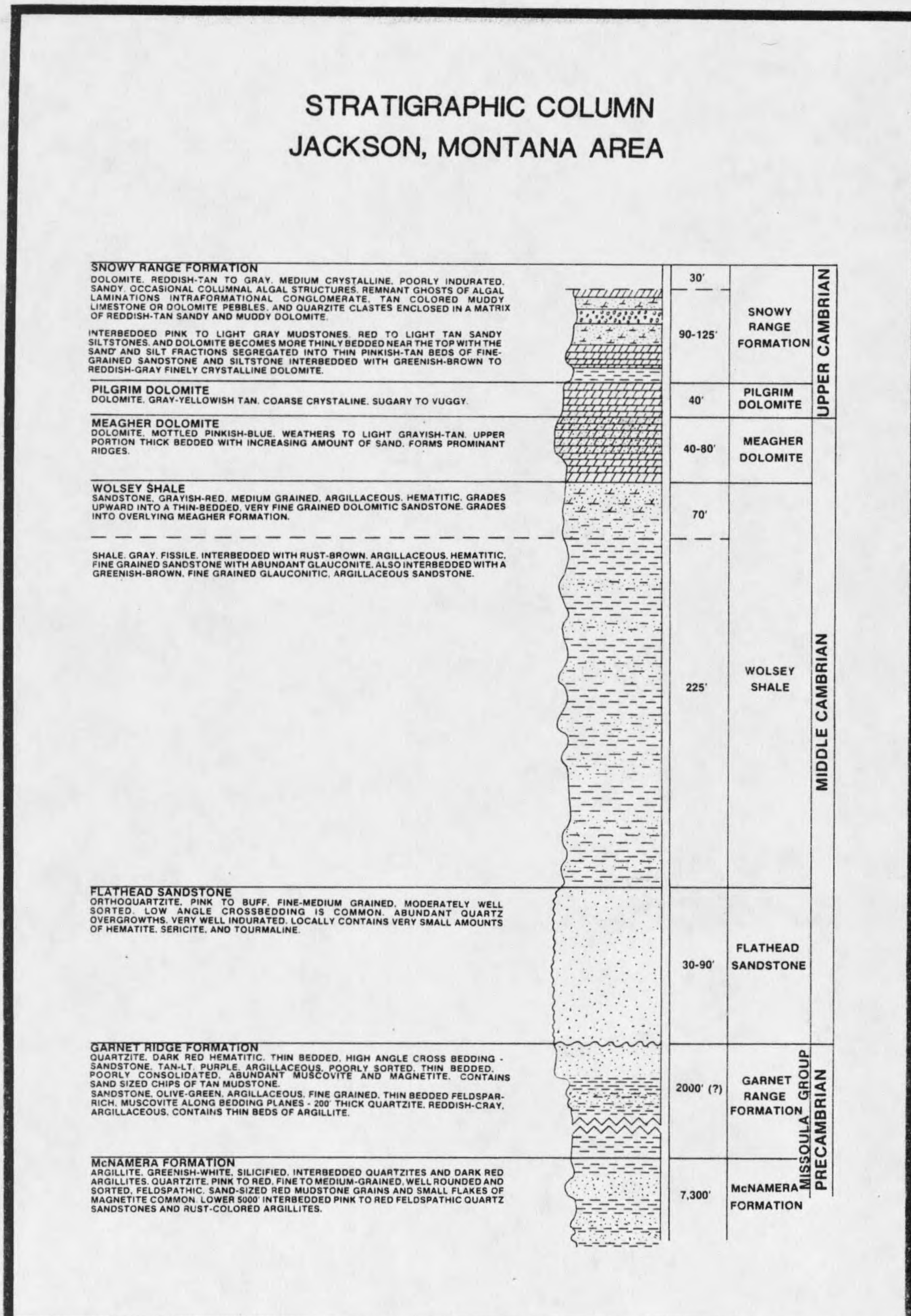

Figure 6. Local stratigraphic column.

Missoula Group-McNamara Formation

The McNamara Formation is approximately 7,300 feet thick in the study area. Bedding strike trends generally northwest. The formation dips southwestward between 40° and 60°, with the steepest dips near the thrust fault (Plate 1).

The formation is predominately composed of thin- to medium-bedded, fine- to medium-grained quartzites. Colors commonly range from pink to red on weathered surfaces. In addition, poorly exposed, rust-colored argillites are interbedded with the quartzites. The grains in the quartzites generally exhibit good rounding and sorting and are approximately 10 to 18 percent feldspar. Sand-sized red mudstone grains and small flakes of magnetite are also common.

The lower 5,000 feet of the exposed section is characterized by 200-500 foot thick layers of ridge forming, generally thick-bedded, fine- to medium-grained, red to pink quartzites. These layers are separated by poorly exposed intervals, probably argillite, that average 800 feet thick.

The quartzites are bimodal in character, which may suggest two distinct sources of McNamara Formation sediments. Quartz grains are moderately well to well rounded and sorted, suggesting extensive transport or reworking. The subordinate feldspars, chiefly plagioclase, are much more angular and poorly sorted. Little sedimentary transport or reworking of these grains is suggested because the angularity of the feldspars is not due to overgrowths, nor are the feldspars of obvious diagenetic origin. Since significantly different amounts of transport are evident for the two major grain types in the quartzites, a dual source is likely.

Exposures of the upper part of the McNamara Formation occur about a mile northeast of Jackson within an irrigation ditch. Here, a distinctive greenish-white silicified argillite is interbedded with quartzites and dark argillites. Although present in only a few beds ranging between two and five feet thick, this pale green argillite is diagnostic of the McNamara Formation (W. B. Myers and E. T. Ruppel, personal communication, 1981) and was used as a basis for the designation of this formation. Also present in this part of the section is a thin-bedded (one to four feet) greenish-red quartzite interbedded with rust-colored argillite and dark red feldspathic quartzites containing abundant muscovite along bedding planes and averaging two feet in thickness. The muscovitic bedding planes are probably the result of low-grade metamorphism.

Sedimentary features found within the McNamara Formation indicate that deposition took place in progressively shallower conditions. Generally clean quartzites with planar cross-bedding characterize the lower parts of the exposed section. Mud cracks, asymmetric ripple marks, and raindrop imprints are common in the upper parts of the exposed formation, evidence of deposition in very shallow water.

Missoula Group-Garnet Range Formation

The Garnet Range Formation consists almost entirely of hematitic or feldspathic quartzites that are less indurated than similar units in the McNamara Formation. As a result, Garnet Range units do not form prominent ridges and are comparatively less well exposed. The poor exposures of Garnet Range lithologies are probably also due to a relatively high percentage of clay (9 to 20 percent) in the quartzites. Units with high percentages of clay also have less feldspar (2 to 5 percent) than quartzites in the McNamara Formation and were significantly more friable. Because no complete exposures were found, the total thickness of this section is unclear. However, a 2,000 foot minimum is estimated on the basis of transects performed on exposed sections.

The lowermost sequence of Garnet Range rocks that was observed consists of approximately 500 feet of intercalated gray feldspathic quartzite, a gray argillaceous quartzite with thin beds of argillite, and a reddish-gray argillaceous quartzite containing numerous clay rip-up fragments. Raindrop imprints and mud cracks are very common in this sequence.

Overlying the intercalated quartzites is a generally poorly exposed fine-grained, feldspar-rich, olive green argillaceous sandstone approximately 200 feet thick. Bedding is thin, ranging from 0.3 to 1.3 inches in thickness. This unit contains abundant muscovite concentrated along bedding planes which is probably of metamorphic origin.

A poorly exposed interval approximately 1,000 feet thick overlies the argillaceous sandstone unit. The only exposures are widely spaced and consist of narrow (one to four feet), thin-bedded, poorly consolidated sandstones. Bedding rarely exceeds one inch in thickness. These sandstones are poorly sorted, rich in muscovite and magnetite, and contain sand-sized chips of tan mudstone. The sandstones are light colored, ranging from tan to light purple.

The uppermost Garnet Range unit is approximately 300 feet thick and consists of very compact, thinly-bedded, dark red hematitic quartzite. It is locally massive and exhibits round reduction spots that are grayish-white in color. High angle cross-bedding is common at the top and at the base of the exposures of this unit, as are occasional raindrop imprints. The base of the unit consists of a sharp termination of good exposure where an abrupt transition is made to the underlying, poorly exposed unit.

Cambrian Sequence

The thick section of Missoula Group rocks is overlain by a thin Cambrian sequence of sandstone, shale, and dolomite. The only previously published description of Cambrian rocks in the Jackson area is found in a regional report by Ruppel and others (1981). The present study agrees closely with the lithologic descriptions of Jackson area rocks as found in that report and differs only in detail. The formation designations used in this study follow the nomenclature found in that report.

Flathead Sandstone

The Cambrian Flathead Sandstone overlies the Precambrian Garnet Range Formation with little or no apparent angular unconformity. Coppinger (1974) has also reported that the Flathead Sandstone in Bray's Canyon, about 15 miles southeast of the study area, has essentially the same attitude as the underlying Belt rocks.

Within the study area, exposures of the Flathead Sandstone range in thickness from 30 to 90 feet and consist of fine- to medium-grained, moderately well-sorted, pink to buff orthoquartzite. Low angle cross-bedding is common. The formation typically contains well rounded quartz grains with abundant quartz overgrowths that form the cement. It is very well indurated and locally contains very small amounts of hematite and sericite and a few small grains of tourmaline.

Wolsey Shale

The Wolsey Shale conformably overlies the Flathead Sandstone. Exposures range in thickness from 260 to 395 feet and consist of a basal shale unit and overlying fine-grained sandstones.

The basal unit is approximately 225 feet thick and contains a dark gray fissile shale that is interbedded with a rust brown, argillaceous, hematitic, fine-grained sandstone with abundant glauconite and a greenish-brown, fine-grained glauconitic, argillaceous sandstone. Ruppel and others (1981) describe the greenish sandstone as a separate unit. However, sufficient evidence was not observed in this study to warrant that distinction.

Overlying the basal shale is a 70 foot thick unit consisting of a medium-grained, hematite-rich sandstone with moderately-rounded quartz grains. This sandstone locally contains abundant Scolithes tubes that range from 0.5 to 1.3 inches in length. This sandstone becomes less well-rounded and more argillaceous upward and grades into a thin-bedded, very fine-grained dolomitic sandstone. This upper dolomitic sandstone comprises the majority of the upper part of the

Wolsey unit and is approximately 55 feet thick. Ruppel and others (1981) designate this upper sandstone as a separate unit from the coarser-grained sandstone immediately below it. Both sandstones, however, are lithologically similar. They are treated here as a single unit that becomes more dolomitic and fine-grained upward, grading into the overlying Meagher Dolomite.

Meagher Dolomite

Conformably overlying the Wolsey Shale are exposures of finely crystalline dolomite. Exposures range in thickness from 40 to 80 feet. This dolomite is typically light grayish-tan on weathered surfaces and mottled pinkish-blue on fresh surfaces. The lower part of the formation is generally lighter in color and more coarse grained than the upper portion. The upper portion is more thickly-bedded, contains upwardly increasing amounts of sand, and forms prominant ridges. The upper Meagher also contains grayish, fissile shale and silt lenses that Ruppel and others (1981) suggest may be equivalent to the Park Shale.

Pilgrim Dolomite

The Pilgrim consists of 40 feet of gray to yellowish-tan dolomite. Exposures range from 10 to 40 feet as a result of covering by thrusted Missoula Group rocks. The Pilgrim Dolomite is generally more coarsely crystalline, less well indurated, and cleaner than the underlying Meagher Dolomite. Extensive recrystallization has obscured bedding and is partially responsible for the sugary to vuggy texture of the

the formation. This dolomite tends to form poorly exposed slopes that overlie the ridges formed by the Meagher Dolomite.

Snowy Range Formation

In the study area, the Snowy Range Formation consists of two distinct sequences; a lower sequence of mudstone, siltstone, sandstone, and dolomite and an upper sequence of sandy dolomite and intraformational conglomerate. Ruppel and others (1981) divide the Snowy Range Formation into six units. Observations made during the course of this study, however, did not provide sufficient evidence to divide this formation into more than two lithologically distinct units.

Exposures of the Snowy Range Formation range in thickness from zero to 155 feet in the study area. Although this variation in thickness may be due to regional pre-Devonian erosion, variations are more likely the result of faulting due to the Snowy Range Formation's presence immediately below the Jackson thrust fault where it is often covered by fault-emplaced Missoula Group rocks.

The lower sequence is 90 to 125 feet thick and composed of interbedded pink to light gray mudstones, red to light tan sandy siltstones, and dolomites. These become more thinly-bedded near the top of this sequence with the sand and silt fractions segregated into thin, pinkish-tan beds of fine-grained sandstone and siltstone interbedded with greenish-brown to reddish-gray finely crystalline dolomite.

Above this sequence is a reddish-tan to gray sequence, approximately 30 feet thick, consisting of medium crystalline, poorly indurated, sandy dolomite. This dolomite has occasional columnal algal structures, generally about six inches high. Internal features of these columns have been mostly obscured by dolomitization. Remnant ghosts of algal laminations can occasionally be observed. According to Balster (1971) and McMannis (1965), such columnal algal structures are characteristic of the Snowy Range Formation.

A distingishing feature of the upper sequence is a distinctive intraformational conglomerate composed of tan colored, muddy limestone or dolomite pebbles, generally one-half to one inch in diameter, and smaller quartzite clasts enclosed in a matrix of reddish-tan sandy and muddy dolomite. The upper part of this unit contains increasing amounts of quartz sand. This sand locally forms cross-bedded layers of clean, moderately-rounded, reddish sandstone at the top of the formation.

Structure

Three types of faulting are present in the study area. A regionally extensive thrust fault trends northwest through the area. Roughly coinciding with the trace of this thrust fault is a poorly exposed zone of post-thrusting extensional faulting. Also, a northeastward trending high-angle normal fault is located in the northern part of the study area and apparently joins a major high-angle fault system in the central Pioneer Mountains.

Jackson Thrust Fault

The dominant structural feature in the study area is a northwest-trending thrust fault, herein termed the Jackson thrust fault, that extends completely through the study area (Plate 1). This fault extends both north and south of the study area and is subparallel with the southeastern margin of the Big Hole Basin. In the northern part of the study area, the fault strikes roughly N 50° W and dips 65-75° to the southwest. South of Warm Springs Creek, the fault curves southward and assumes a generally north-south trend.

The Jackson thrust fault is located within the Grasshopper thrust plate of Ruppel and others (1981). The Grasshopper plate is a thrust sheet in the sense used by Boyer and Elliot (1982) in that it is a coherent volume of rock underlain by a basal thrust fault. The basal thrust of the Grasshopper plate is represented south of the Pioneer Mountains by the Kelley thrust of Myers (1952) and north of the batholith by the Johnson thrust of Moore (1956), Fraser and Waldrop (1972), and Calbeck (1975) (Ruppel and others, 1981). The plate is also exposed on the western and southern margins of the Big Hole Basin. Since the Grasshopper plate is present both east and west of the Jackson area, the relationship between the Jackson thrust and the basal thrust fault of the Grasshopper plate must be addressed.

According to Ruppel and others (1981), the Precambrian and Cambrian sedimentary rocks in the Jackson area are correlative with rocks of similar age contained within the Grasshopper plate and are distinct from Precambrian and Cambrian sedimentary rocks contained within the structurally lower "frontal fold and thrust zone." Therefore, the rocks exposed beneath the Jackson thrust fault do not constitute a window through the Grasshopper plate and the basal thrust fault of the plate must completely underlie the study area. Hence, the Jackson thrust is probably an imbricate thrust fault in the hanging wall of the basal thrust of the Grasshopper plate and its position is likely localized by a ramp or step in the underlying basal thrust (e.g. Boyer and Elliott, 1982; Dahlstrom, 1970, 1977; Royse and others, 1975).

Extensional Motion on the Jackson Fault. In much of the study area, hanging wall rocks of the Precambrian McNamara Formation are overlain by a thin mantle of poorly consolidated Tertiary valley fill sediments. These Tertiary sediments can be found in small, isolated patches on Precambrian and Cambrian rocks wherever these rocks are exposed. It appears that the topographic expression of the Precambrian and Cambrian sedimentary rocks in the study area represent, to a large degree, a pre-Tertiary topography that has been covered by Tertiary sediments and subsequently exhumed by erosion.

In the southern part of the study area, the Tertiary mantle is downfaulted against the upper part of the Cambrian section. Although the zone of extensional faulting is generally poorly exposed, it is well delineated in the northeast quarter of Section 9, T. 6 S., R. 14 W. Here, motion occurred along the Jackson thrust fault which is exposed by narrow outcrops of Precambrian McNamara Formation in fault contact with the Cambrian Meagher Dolomite at the base of a brecciated fault scarp which exhibits 140 feet of vertical displacement. The dip of this fault scarp is approximately 65° southwest. The McNamara Formation exposures are isolated and are covered by Tertiary sediments away from the fault scarp.

North of Warm Springs Creek, the fault apparently continues northward where Tertiary rocks are in contact with the Cambrian Meagher Dolomite. In the southeast quarter of Section 18, T. 5 S., R. 14 W., the normal fault is east of the Jackson thrust fault (Plate 1). Here, the Cambrian Flathead is thinned (stratigraphic thickness approximately 20 feet) and brecciated by normal faulting and the Wolsey Shale is abnormally thick. The dip of the fault in this area is approximately 60° to the southwest. The fault continues one mile northwestward where it intersects a high angle, northeast trending fault in the southeast quarter of Section 12, T. 5 S., R. 15 W. The continuation of the northwest trending normal fault northward from its intersection with this high angle fault is uncertain due to poor exposures.

In the southwest quarter of Section 12 and the north half of Section 13, T. 5 S., R. 15 W., another northwest trending normal fault is present one-quarter mile west of the Jackson thrust fault. This normal fault is delineated by a brecciated fault scarp, about nine feet high, that dips approximately 70° southwest. Tertiary rocks are downfaulted against the Precambrian McNamara Formation along this fault. Its continuation to the north and south from this area is uncertain due to covering by Tertiary sediments. Normal faulting, therefore, has occurred on both sides of the Jackson thrust fault in this area. It is uncertain whether extensional motion has taken place along the Jackson thrust fault here as it has in the southern part of the study area.

Northeast-Trending High Angle Fault

A high-angle fault, herein termed the Warm Springs Creek fault, cuts Precambrian Missoula Group and Cambrian rocks in the northern part of the study area (Plate 1). Both Precambrian and Cambrian rocks are extensively brecciated along this fault in the southern part of Section 12, T. 5 S., R. 15 W. Here, the high angle fault cuts the complete Cambrian section in the footwall of the Jackson thrust fault as well as the Precambrian McNamara Formation in the hanging wall of the thrust fault. Exposures of brecciated Cambrian Meagher Dolomite in this area also show an abrupt change in strike toward the high angle fault and appears to be caused by drag along the fault. In addition, the McNamara Formation is offset along the high angle fault as well. Both the drag features and the offset within the McNamara Formation indicate that some component of right lateral motion has occurred along this high angle fault. The scale of these features is not sufficient to depict them on Plate 1. The dip of the fault is near vertical as evidenced by the exposed zones of brecciated rock in this area.

East of Section 12, T. 5 S., R. 15 W., the trace of the fault is uncertain because it apparently trends through the poorly exposed Precambrian Garnet Range Formation. It is probably located just north of Jackson Hill as evidenced by a linear zone of breccia in this area. The high angle fault apparently continues northeastward down a steep gulley into the valley of Warm Springs Creek. It may continue up this valley and join a high angle, northeast trending fault mapped by Snee (personal communication, 1981) who speculates the fault may join the southern portion of the Fourth of July fault mapped by Calbeck (1975) in the central part of the Pioneer Mountains (Chapter 2, Figure 5).

Inferred North Trending Normal Fault

An inferred normal fault is postulated in the study area that is west of, and subparallel to, the Jackson thrust fault. This normal fault is inferred on the basis of subtle surface expression as well as geochemical and hydrologic evidence.

The location of the inferred normal fault is mapped in the southern part of the study area on the basis of Tertiary-Quaternary contacts in Section 11, T. 6 S., R. 15 W., and Section 35, T. 5 S., R. 15 W. (Plate 1). In Section 11, the Tertiary-Quaternary contact is linear and north-trending. This linearity is assumed to be due to normal faulting, although erosion of the Tertiary sediments by a tributary of the Big Hole River cannot be ruled out as a possible cause. In Section 35, the width of the northwest-trending bench of Tertiary sediments narrows abruptly near its terminus along a north-trending linear zone. This zone of narrowing is due north of the linear contact described above and may be the continuation of the inferred normal fault.

North of Warm Springs Creek, the trace of the inferred normal fault is mapped primarily on the basis of a linear zone that extends both northwest and southeast of a warm spring in Section 23, T. 5 S., R. 15 W., along which rapid snow melt occurs. This zone of preferential snow melt extends at least one mile to the southeast toward Jardine Hot Spring. It is hypothesized that the rapid snow melt along this zone is due to the upwelling of thermal water along the inferred normal fault. Geochemical and hydrologic data indicate that the thermal water discharged from Jardine Hot Spring and from the warm springs approximately one and a quarter mile to the northwest is from the same hydrothermal system and that discharge is occurring along a fault zone (Chapter 5). This fault zone is assumed to be the inferred normal fault.

Royse and others (1975) note that a coincidence between the position of young normal faults and older thrust faults exists in the thrust belt and that the position of many normal faults is related to step geometry in an underlying thrust (Figure 7). Examples of this relationship are the Flathead fault in northwest Montana and the Hoback fault near Jackson, Wyoming. More recently, Allmendinger and others (1983) note that high angle normal faults in the Basin and Range Province are truncated or sole into low angle detachments and that at least some of these low angle detachments may be older thrust faults that have undergone Cenozoic reactivation. Within the study area, although the exact trace of the inferred north trending normal fault is speculative, it is evident that normal faulting has occurred both east and west of the Jackson thrust fault fault. Normal faulting in the Jackson area may have taken place along a zone of normal faulting rather than along one normal fault. The zone of normal faulting is approximated by the trace of the inferred normal fault shown on Plate 1.

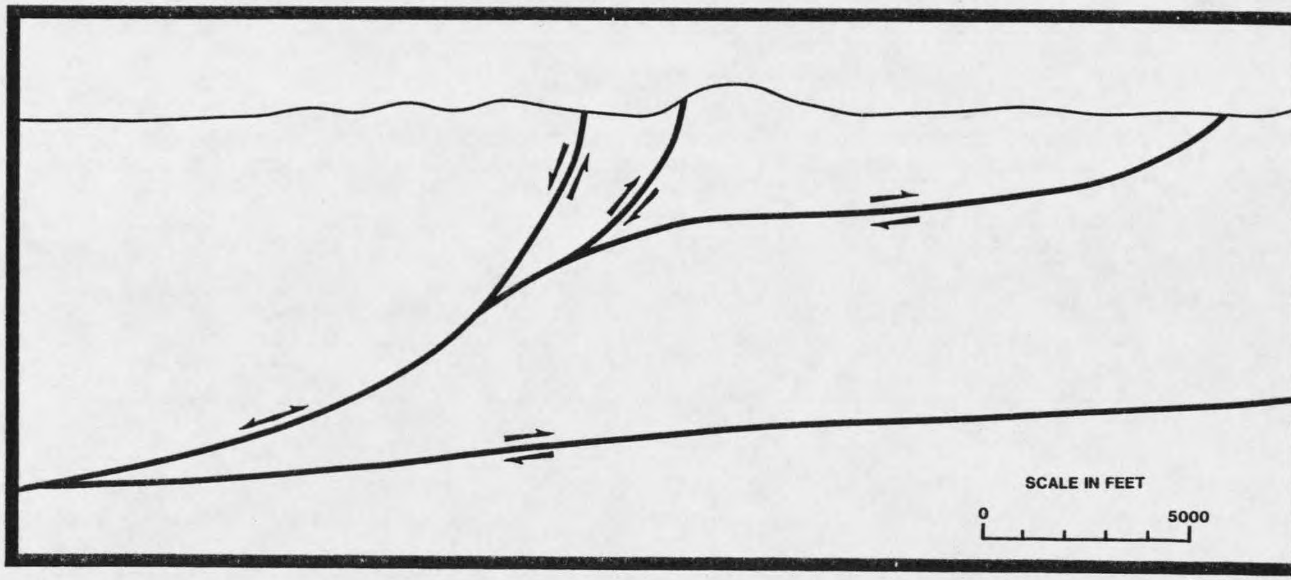

Figure 7. Hypothetical cross section showing the localization of a normal fault by a ramp in a thrust fault. (From Royse and others, 1975)

CHAPTER 4

GENERAL PRINCIPLES OF GEOTHERMAL SYSTEMS

Basic concepts of hydrothermal systems are outlined in this chapter for the general reader unfamiliar with geothermal resources. It is designed as a preface to the succeeding two chapters on the hot spring system of the Jackson, Montana area. Basic heat flow concepts and classifications of geothermal systems are discussed.

Thermal Energy of the Earth's Crust

Much of the Earth's thermal energy is stored in the rocks of the Earth's crust. The movement of this crustal energy toward the Earth's surface is principally the result of heat conduction through crustal rock. Temperature changes between the relatively cool rocks at the Earth's surface and hotter rocks at depth constitute the geothermal gradient, commonly defined as the change in temperature divided by the change in depth from the Earth's surface. Crustal heat flow is a function of both the geothermal gradient and the ability of crustal rocks to conduct heat.

Thermal energy is normally diffuse in crustal rocks. In such a state it is of little economic importance. In some areas, however, geothermal energy is concentrated and transferred directly to the Earth's surface or to relatively shallow depths. Geothermal energy that is concentrated in areas where it can be utilized at costs that are competitive with other forms of energy constitutes the Earth's geothermal resources (Muffler and Guffanti, 1978). Geothermal resources can be divided into two major groups: conductive-dominated and convective-dominated systems (Rybach, 1981)

Conductive Dominated Systems

Heat transfer in conductive systems occurs solely by means of conduction through rock and immobile pore water. Examples of this type of system include geopressured reservoirs and Hot Dry Rock systems. Geopressured reservoirs are zones bounded by impervious shales and which contain water with abnormally high pore pressure. Such zones are effective heat traps that concentrate thermal energy. Hot Dry Rock systems consist of low permeability rocks in areas of high geothermal gradients. Water or some other fluid must be pumped down into the high temperature zones and brought back to the surface after being heated in order to recover the thermal energy from these systems. Due to economic and technical constraints, conductive systems are not likely to be exploited on a significant scale for a number of years (Blair and others, 1982).

Convective Hydrothermal Systems

The vast majority of geothermal resources of commercial interest are of the convective type. In these systems, meteoric water moves downward along permeable aquifers or faults to depths where it is heated by the surrounding rocks. The heated fluid, having undergone a decrease in density due to thermal expansion, rises bouyantly along zones of adequate permeability. The pressure gradient resulting from the density differences between the cold descending water and the hot upwelling fluid is the primary driving force of these systems (Donaldson, 1968). The upward convection of hot fluid effects large amounts of heat transfer to the Earth's surface (Garg and Kassoy, 1981).

Convective hydrothermal systems are divided into two main categories; vapor-dominated and liquid-dominated. Classification depends upon whether steam or liquid water is the dominant phase that controls the pressure of the system in its unexploited state (Brook and others, 1978). The pressure at any given point in the reservoir is approximately equal to that of a standing column of the dominant fluid equal in height to the depth of the reservoir (Donaldson and Grant, 1981).

Vapor-Dominated Systems

These systems are relatively rare. They are thought to develop when deep subsurface water is heated and begins to boil. Boiling occurs at much higher temperatures than at the surface due to pressure effects. If the heat supply is great enough to boil more water than can be replaced by recharge from the margins of the reservoir, a vapor-dominated system will form (Renner and others, 1975; Wright, 1980).

Only three major vapor-dominated systems have been identified in the United States; the Mount Lassen and Geysers systems in northern California and the Mud Volcano system in Yellowstone National Park, Wyoming (Blair and others, 1982). Because vapor must be expelled from these systems, they are likely to have identifiable vent areas (Renner and others, 1975). Hence, it is unlikely that many more large vapor-dominated systems are yet to be discovered in the United States.

Liquid-Dominated Systems

These systems are traditionally divided into two subgroups according to the predominant heat source of the system. One group consists of systems related to shallow intrusions and includes almost all geothermal systems developed for electric power generation (Rybach, 1981). Systems in the other group derive their heat from circulation within a thermal regime of regional heat flow.

Systems which derive their heat from magmatic thermal anomalies are nearly always associated with silicic magma storage chambers at high crustal levels. Viscous silicic magmas tend to lodge in the crust and cool slowly, thereby remaining a heat source for a long duration (Rybach, 1981). Basaltic magmas on the other hand, because of their relatively low viscosity, rise more quickly through the crust and cool quickly, usually at the surface. The intrusion of a silicic magmatic body into high levels of the crust is often accompanied by fracturing and faulting. The fractures commonly allow cold water to descend close to the intrusion, become heated by the intrusion, and ascend to the surface due to the density effects discussed earlier. Smith and Shaw (1975) speculate that a large igneous body on the order of 10 kilometers in thickness may be a significant source of heat for up to 10 million years.

The majority of liquid-dominated systems are not located near silicic intrusions. Instead, most systems derive heat from the geothermal gradient in a region. As near-surface water moves downward along faults, fractures, or permeable stratigraphic beds, it is heated by thermal conduction from the surrounding rock. The temperatures that these systems attain are determined primarily by the depths to which water circulates and by the magnitude of the geothermal gradient. Deep circulation in regions of abnormally high heat flow often results in a high concentration of moderate- to high-temperature hot springs. It is rare, however, for these springs to be at the boiling temperatures which characterize the springs of liquid-dominated systems located near cooling intrusions (Renner and others, 1975).

CHAPTER 5

GEOCHEMISTRY OF THERMAL WATERS AND ITS APPLICATION TO JARDINE HOT SPRING

In this chapter, general principles of the geochemistry of geothermal systems are reviewed and applied to the system at Jackson. Chemical geothermometers and models useful for the recognition of hydrologic mixing are discussed. Using the data presented, geochemical models of the Jackson hydrothermal system and probable reservoir temperatures are proposed at the end of the chapter.

Geothermometry

The discharge water from hydrothermal systems contain chemical concentrations that vary predictably with reservoir temperature. As a result, an analysis of the discharge water can provide an estimate of subsurface temperature. Chemical reactions most commonly used to estimate subsurface temperatures are: (1) the solubility of silica (Fournier and Rowe, 1966; Mahon, 1966), and (2) the cation exchange ratio of sodium, potassium, and calcium (Fournier and Truesdell, 1973).

The solubilities of most common minerals increase with increasing temperature and pressure. In a hydrothermal system, cool shallow groundwater is heated upon descent, dissolving progressively more silicates until a maximum concentration is reached at the hottest part of the system (Fournier and others, 1974). Although water temperature may decrease as the thermal water rises to the surface, only small changes in water chemistry may occur, especially if upward flow is rapid. This is due to the long time required to effect significant chemical changes relative to the flow time of the thermal water. Therefore, the absolute quantity of silicates, chiefly quartz, can be used to estimate the reservoir temperatures of hydrothermal systems.

In addition to chemical concentration, the ratios of certain ions can be used to estimate the subsurface temperatures of geothermal systems. An empirical geothermometer was developed by Fournier and Truesdell (1973) that uses the variation of sodium, potassium, and calcium ions in hydrothermal waters. It was noted that the ratio of these ions change as a function of temperature in response to temperature-dependent exchange reactions between the hot water and the reservoir rock. An empirical equation using the ratio of sodium, calcium, and potassium ions in thermal waters can be used to estimate reservoir temperature.

Assumptions

In using the concentrations of chemicals for geothermometry, certain assumptions must be made about the geothermal system and the chemical reactions that occur within it. It must be assumed that the concentrations of the chemicals used in analysis are controlled by temperature-dependent reactions between the hot water and the reservoir rock. It must also be assumed that the chemical reactions used for geothermometry involve reactants in abundant supply so that reactant availability is not a constraint. In addition, it must be assumed that chemical equilibrium is attained at reservoir temperature with respect to the reactants of interest. Ellis and Mahon (1964, 1967) have demonstrated that these assumptions are met in most geothermal systems.

In addition to the assumptions concerning conditions in the geothermal reservoir, certain assumptions must be made about the thermal water after it leaves the reservoir (Fournier, 1977). It must be assumed that no re-equilibration of reactants occurs as the water flows from the reservoir to the surface. If some re-equilibration does occur, the chemical geothermometers will yield a temperature estimate that is lower than the true temperature of the reservoir. It must be remembered that because of the possiblility of some re-equilibration of the reactants as thermal water flows upward, chemical geothermometry yields minimum temperature estimates.

Anomalously low temperature estimates will also result if the hot water coming from depth is diluted by cold, near-surface water. Chemical nonequilibrium will occur as a result of such mixing. Fortunately, the chemical nonequilibrium resulting from the mixing of two waters can often be used to estimate how much dilution of thermal water is occurring, as well as the temperature of the hot water component prior to dilution. This method is discussed later in this chapter.

Silica Geothermometer

The concentration of dissolved silica in discharge water from high temperature hydrothermal systems can be used to estimate subsurface temperatures in a quantitive manner (Mahon, 1966; Fournier and Rowe, 1966). The method has been discussed in detail by numerous authors (e.g. White, 1970; Ellis and Mahon, 1977; Fournier, 1981) and will only be summarized here. The method is straight-forward provided the assumptions discussed earlier are considered. Additional considerations are the temperature range for the equations used in analysis, the effects of pH on silica concentration, and the specific minerals that control silica solubility.

Temperature effects do not appear to be a problem at the Jackson geothermal system. The equations used to estimate reservoir temperatures are most accurate when applied to thermal waters with temperatures of less than 250°C (Fournier, 1981). The estimated temperatures for the Jardine Hot Spring system are well within this range.

The effects of pH on silica concentration vary with temperature in a nonlinear manner. The maximum effect occurs in waters above 150°C whose pH exceeds eight. Below 150°C, only highly alkaline waters show significant pH effects on silica concentration. Because the pH of Jardine Hot Spring is nearly neutral, pH effects on the silica geothermometer are not considered to be significant.

Although quartz is the most stable polymorphic form of silica within the range of temperatures and pressures in most geothermal reservoirs, other silica phases may exist metastably (Fournier, 1981). Arnorsson (1975) noted that waters above 180°C are in equilibrium with quartz and that waters below 110°C are often in equilibrium with chalcedony. Between these temperatures, silica concentration may range between the equilibrium values for quartz and chalcedony. In this study, equations using the concentrations of both quartz and chalcedony are used for the silica geothermometer.

The equations used for estimating reservoir temperature utilizing quartz and chalcedony solubilities, with silica concentration expressed in parts per million, are (Fournier, 1981):

1. Quartz solubility-

   Temperature (°C) = $[1309/(5.19 - \log SiO_2)] - 273.15$

2. Chalcedony solubility-

   Temperature (°C) = $[1032/(4.69 - \log SiO_2)] - 273.15$

Silica concentrations for Jardine Hot Spring are shown in Table 1 and yield the estimated reservoir temperatures shown in Table 2.

Sodium-Potassium-Calcium Geothermometer

The ratios of certain ions in thermal waters change in response to temperature-dependent ion exchange reactions within geothermal reservoirs. Such reactions occur primarily between the hot water and the reservoir rock. The exchange reactions between sodium and potassium are strongly temperature dependent (Ellis and Mahon, 1977). For this reason, the ratio of sodium and potassium in thermal waters can give reliable estimates of reservoir temperatures, especially above 180°, as shown by White (1965), and Ellis and Mahon (1967). For lower temperature reservoirs, the retrograde solubility of calcium with regard to temperature may affect the exchange of sodium with potassium (Ellis and Mahon, 1977). As a result, Fournier and Truesdell (1973) developed an empirical Na-K-Ca geothermometer that accounts for low

Table 1. Geochemical data for wells, springs and stream in Jackson area

Location and Data Source	Measured Temp.(°C)	SiO_2	Ca	Mg	Na	K	Li	HCO_3	SO_4	Cl	F	pH	Flow Rate
Jardine Hot Spring This Study (11/81)	57.6	49.3	10.3	3.0	226.0	8.5	0.29	615.0	45.5	8.0	1.76	7.36	
Jardine Hot Spring This Study (5/81)	58.6	48.4	10.5	3.3	231.0	10.1	0.31	612.0	47.9	7.6	2.10	7.73	
Jardine Hot Spring Robertson and others, 1976	58.5	49.3	10.5	3.2	428.0	11.0	0.35	632.0	50.0	11.0	1.3	-	60 liters/sec.
Lapham Warm Springs This Study (11/81)	19.4	24.9	10.6	3.5	188.0	8.5	0.23	504.0	40.2	8.0	1.70	7.41	
Lapham Domestic Well This Study (11/81)	15.2	16.5	28.8	3.0	196.0	8.8	0.24	568.0	41.2	6.7	1.73	7.20	
Jackson School Well This Study (11/81)	12.0	16.3	36.3	7.3	20.2	2.8	0.014	176.4	16.5	6.9	0.34	7.55	
Patterson Cold Spring This Study (11/81)	7.0	55.2	26.4	8.0	9.2	4.9	0.007	132.2	8.1	7.0	0.24	7.34	
Fernandez Domestic Well This Study (11/81)	9.5	50.8	27.4	10.0	8.8	4.7	0.006	142.7	5.8	4.0	0.20	7.38	
Work Center Well This Study (11/81)	7.0	34.2	6.2	0.1	47.1	5.4	0.035	119.6	20.6	6.1	0.32	8.12	
Warm Springs Creek This Study (11/81)	5.2	23.4	6.7	2.0	44.0	2.7	0.54	134.2	11.5	2.4	0.49	7.60	

Table 2. Calculated Reservoir Temperatures for Jardine Hot Springs

Data Source	Quartz (No Steam Loss)	Quartz (Max. Steam Loss)	Chalcedony	Na-K-Ca	Na-K-Ca (Mg. Corr.)
This Study (5/81)	101°C	102°C	71°C	142°C	66°C
This Study (11/81)	100°C	101°C	70°C	149°C	65°C
Robertson et al., 1976	101°C	102°C	70°C	145°C	69°C*
Mariner et al., 1976	104°C	–	73°C	148°C	–
Chadwick and Kaczmarek, 1975	99°C	–	–	–	–

*Mg. Corrected estimate from Nathenson, 1981 using chemical data from Robertson et al., 1976 temperature, calcium-rich waters. The equation used (Fournier, 1981) is:

$$\text{Temperature } (^\circ\text{C}) = \frac{1647}{\log(\text{Na/K}) + \beta[\log(\text{Ca/Na}) + 2.06] + 2.47} - 273.15$$

where

β= 0.33 when T°C > 100°C (using β= 1.33) or if [log(Ca/Na) + 2.06] is negative

β= 1.33 when T°C < 100°C (using β= 1.33) and [log(Ca/Na) + 2.06] is positive

and where Na, Ca, and K concentrations are expressed in parts per million. Calculated reservoir temperatures for Jardine Hot Spring using the Na-K-Ca geothermometer are shown in Table 2.

The Na-K-Ca geothermometer can give erroneous results when used on waters with low pH, on waters that issue after subsurface boiling or mixing has occurred, or on waters high in magnesium. As with the silica geothermometer, pH effects are not a problem at Jackson because the water is nearly neutral. Subsurface boiling causes a loss of CO_2, the precipitation of $CaCO_3$, and anomalously low concentrations of Ca^{++}, which will cause the calculated temperature to be too high (Fournier, 1981). Dilution effects are generally negligible if the hot water component is much more saline than the diluting water and if the hot water fraction is at least 20-30 percent of the mixed water (Fournier, 1981). Dilution is covered in greater detail in the following section of this chapter.

Fournier and Potter (1979) demonstrated that water rich in Mg^{++} will yield anomalously high temperature estimates using the Na-K-Ca geothermometer. The following equation is used to correct the Na-K-Ca geothermometer for high Mg^{++} concentration (Fournier and Potter, 1979):

$$\Delta t_{mg} = 10.66 - 4.7415\, R + 325.87\, (\log R)^2 - 1.032 \times 10^5\, (\log R)^2/T - 1.968 \times 10^7\, (\log R)^2/T^2 + 1.605 \times 10^7\, (\log R)^3/T^2$$

where

$R = [Mg/(Mg + Ca + K)] \times 100$, with concentrations expressed in equilivents

Δt_{mg} = the temperature correction in °C that should be subtracted from the Na-K-Ca calculated temperature

T = the Na-K-Ca calculated temperature in °K

R is between 5 and 50.

Magnesium-corrected Na-K-Ca temperature estimates are shown in Table 2.

Subsurface Mixing of Hot and Cold Waters

Many springs, especially warm springs with high flow rates, originate from the mixing of high-temperature water with cold groundwater. This mixing occurs either at the margins of the deep parts of the hydrothermal system or near the surface (Fournier, 1981). Most methods used to recognize mixing are based on comparisons of different springs or wells of a hydrothermal system. Where only one spring or well exists, recognition can be difficult. In single springs, mixing may be indicated by marked chemical non-equilibrium at the spring's discharge temperature, or by systematic variations in the composition or temperature of the discharge water (Fournier and Truesdell, 1974; Fournier, 1981).

If subsurface mixing of hot and cold waters is occurring, the pre-mixing temperature of the hot water component and its volume relative to the cold water fraction can be estimated. The silica-enthalpy mixing model (Fournier and Truesdell, 1974; Truesdell and Fournier, 1977) estimates the enthalpy of the hot water component, the original concentration of dissolved silica, and the fraction of hot water in the warm spring. If it is assumed that no steam escapes from the system, the model will calculate a higher temperature of the hot water component than if maximum steam loss is assumed. Accurate temperature estimates require that no heat is lost through conduction after mixing has occurred. Conductive heat loss after mixing will cause the estimated temperature of the hot water component to be too high (Fournier, 1981). Such heat loss is unlikely in springs with high flow rates (Fournier and Truesdell, 1973), such as Jardine Hot Spring. Temperature estimates must also assume that the initial silica content of the hot water component is controlled by the solubility of quartz, and that no solution or deposition of silica occurs after mixing. This assumption is reasonable because the rate of quartz precipitation decreases drastically as temperature decreases (Fournier, 1981).

Truesdell (1976) developed a computer program that uses the methods of Fournier and Truesdell (1974) to calculate temperatures from five chemical geothermometers and two mixing models. The geothermometers used are: (1) quartz saturation with adiabatic cooling, (2) quartz saturation with conductive cooling, (3) chalcedony saturation with conductive cooling, (4) Na-K-Ca ratios with $\beta = 1/3$, and (5) Na-K-Ca ratios with $\beta = 4/3$. One mixing model is applicable to warm springs and the other is applicable to boiling springs. The boiling spring model was not used for this study because it does not apply to Jardine

Hot Spring. Table 3 shows the results of the silica-enthalpy model using both the graphical method of Truesdell and Fournier (1977) and the computer program of Truesdell (1976). For this study, the computer program was run on an IBM 3380 computer using the PL-1 language.

Table 3. Mixing Model Results (Warm Spring Model)

Graphical Method (Truesdell and Fournier, 1976)	
Temperature of Hot Water Component:	145°C
Fraction of Hot Water Component:	32%
Computer Method (Truesdell, 1976)	
Temperature of Hot Water Component:	140°C
Fraction of Hot Water Component:	36%

Parameters Used:
Temperature of Warm Spring: 58.6°C
Silica Content of Warm Spring: 48.4 mg/1
Temperature of Cold Water Component: 12.0°C
Silica Content of Cold Water Component: 16.3 mg/1

When using the silica-enthalpy mixing model, the temperature and silica content of the local groundwater must be estimated. In order to make groundwater temperature and silica content estimates for the Jackson area, six locations were sampled in addition to Jardine Hot Spring and the warm spring northwest of Jackson (Figure 8). All six locations have measured water temperatures below 20°C. The four locations east of Jardine Hot Spring are supersaturated with respect to silica (Table 2). Because their high silica concentrations indicate non-equilibrium conditions, data from these locations are not used in the mixing calculations.

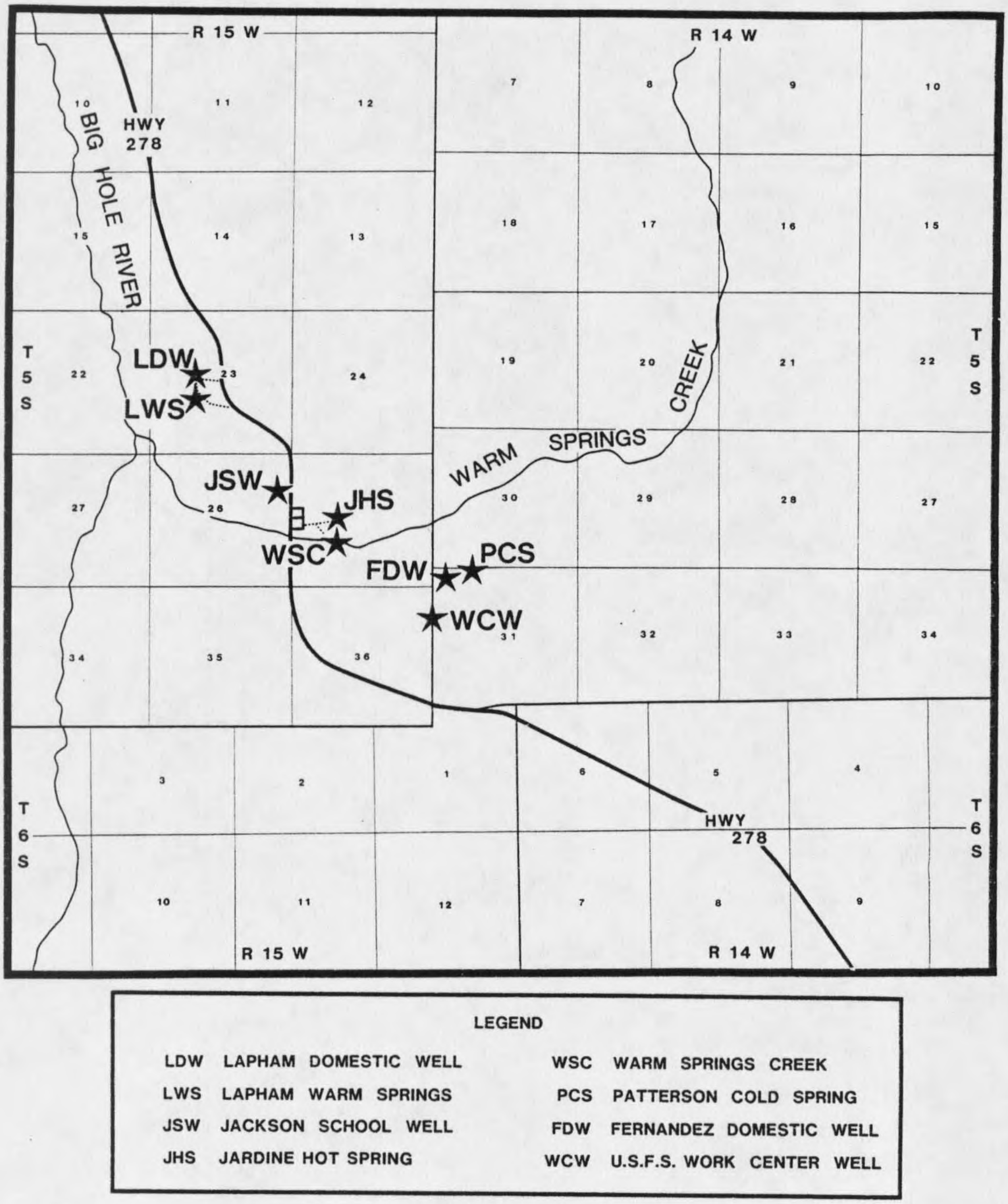

Figure 8. Map of water sample collection stations.

The Lapham domestic well is located near a small group of warm springs, herein termed the Lapham Warm Springs, that discharge into a shallow pond. Measured water temperatures of these springs range from 18°C to 24°C. Large amounts of gas, possibly CO_2, also discharge from these springs. The discharge of the largest spring surges periodically, with a 3-4°C increase in temperature during each surge. These springs have a high sodium content, yielding a Na-K-Ca calculated temperature of 118°C, and a moderate silica content which yields silica geothermometer temperatures of 40°C (chalcedony) and 72°C (quartz). The water from the nearby well is also high in sodium and yields a Na-K-Ca calculated temperature of 94°C. Data from this well was not used because its proximity to the Lapham Warm Springs and its high sodium content indicate that the well water may be influenced in its chemical composition by thermal water.

The sixth location, the Jackson School well, is approximately 1,950 feet northwest of Jardine Hot Spring. It is reported to be 195 feet deep. The measured water temperature is 12°C. Estimated temperatures using both the Na-K-Ca and chalcedony geothermometers are within 22°C, indicating it is near chemical equilibrium. Because of its apparent equilibrium and lack of evidence indicating contamination by warm water, the water from this well is considered to be representative of the cold groundwater. The temperature and silica content of this well were used in mixing calculations.

Geochemical Models

Three geochemical models of the hydrothermal circulation system at Jackson are consistent with the results of the geochemical study. The models are termed the 'warm', 'mixing', and 'precipitating' models for discussion purposes. The 'warm' model assumes that only warm waters exist at depth, with temperatures near 70°C, and that reservoir water is in equilibrium with chalcedony. The 'mixing' and 'precipitating' models assume that the reservoir water is in equilibrium with quartz and that the chalcedony and Mg-corrected Na-K-Ca geothermometers are too low. The 'mixing' model assumes the mixing of separate water types, with a hot water component of approximately 145°C, that comprises about 32 percent of the total mixed water volume. The 'precipitating' model assumes that hot geothermal water, in equilibrium with quartz, precipitates silica as it rises through ashy layers in Tertiary sediments until equilibrium is approached with respect to chalcedony. The relative merits of each model are discussed below in light of the results from the geochemical study.

The 'warm' model is supported by the fact that a magnesium correction to the Na-K-Ca geothermometer will bring it into close agreement with the silica geothermometer. Nathenson (1981) noted that by selective use of the magnesium correction, the cation and silica geothermometers for many of the hot springs in southwestern Montana can be brought into fairly close agreement. According to Nathenson (1981), this agreement implies that the reservoir waters are in equilibrium at the temperatures indicated by the chemical geothermometers, that the silica and Mg-corrected cation geothermometers represent reservoir temperatures, and that subsurface mixing of hot and cold waters is unlikely.

At Jackson, an agreement between the chalcedony and the Mg-corrected Na-K-Ca geothermometers may indicate that the reservoir temperature is between 65°C and 70°C, the silica concentration is controlled by chalcedony solubility, and that no mixing is occurring. The water may be cooled to the discharge temperature of 59°C by heat loss to the surrounding wall rocks during ascent. This model is further supported by a reported sulfate isotope-calculated temperature estimate of 74°C for Jardine Hot Spring (Mariner and others, 1976). This temperature estimate suggests either that the chalcedony and Mg-corrected cation geothermometers are approximately correct, or that sulfide in the thermal water is being oxidized in a shallow aquifer where the quartz-saturated water equilibrates with chalcedony (Mariner and others, 1976). The latter condition would indicate that temperatures higher than 70°C exist at depth.

Temperatures higher than the 65°C to 70°C range suggested by the chalcedony and Mg-corrected cation geothermometers are indicated by the 'mixing' model. Close agreement exists between the uncorrected Na-K-Ca geothermometer temperature estimate of 149°C and the silica-enthalpy estimated temperature of 145°C estimated for the hot water component of a mixed water. If mixing is indeed occurring, use of the magnesium correction to the Na-K-Ca geothermometer may be inappropriate. This is because cold groundwaters are considerably higher in dissolved magnesium than are hot geothermal waters (Ellis and Mahon, 1977).

Magnesium enrichment of the thermal water will occur just prior to discharge and will not have affected cation exchange in the geothermal water and hence would not affect the cation ratios of the hot water component. In addition, mixing with fresh water would cause the silica geothermometers to yield anomalously low results but would have less of an effect on the cation geothermometer. Thus, the apparent agreement between the Mg-corrected cation geothermometer and the chalcedony geothermometer may be coincidental.

The 'mixing' model is further supported by the fact that Jardine Hot Spring has other characteristics that are typical of mixed waters. These characteristics include Jardine Hot Spring's relatively high discharge rate and sub-boiling discharge temperature and a cation-estimated temperature that is much higher than either the discharge temperature or the quartz-estimated temperature (Mariner and others, 1976). In addition, Mariner and others (1976) calculated solution-mineral equilibria for calcite, aragonite, chalcedony, alpha-cristobalite, and fluorite for twenty-one thermal springs in western Montana and found the water at Jardine Hot Spring to be undersaturated with respect to calcite and fluorite. Possible explanations for the anomalously low concentrations of these minerals at Jackson include the absence of calcite in the thermal reservoir, the presence of a high temperature system, or the dilution of the thermal water with cold groundwater. The cold water dilution explanation is favored by Mariner and others (1976).

The possiblity of mixing is also supported by the location of Jardine Hot Spring. A portion of the upward flow component of the cold groundwater flow system may pass through the zone of discharge of the hydrothermal flow system because the hot spring is on the floodplain of a relatively major stream near the base of a large mountain mass. The piezometric head on the cold groundwater could, therefore, force cold water through the permeable valley fill into the hot water discharge route. The resulting mixture would then be discharged at the surface.

The 'precipitating' model, like the 'mixing' model, assumes that the reservoir temperature of the Jackson hydrothermal system is higher than the temperatures indicated by the chalcedony and Mg-corrected cation geothermometers. The model also assumes that quartz solubility controls silica content. The 'precipitating' model predicts that hot water ascending through Tertiary and Quaternary valley fill is influenced by one or more beds containing siliceous ash. In addition to the Tertiary ash beds described in Chapter 2, relatively shallow, ash-rich beds may be indicated by anomalously high silica contents in cold springs and wells east of Jardine Hot Spring (Table 1). Cold groundwaters will often dissolve silica from such ash or glass-rich beds. Conversely, when thermal water that is in equilibrium with quartz encounters volcanic ash or glass, precipitation of silica often occurs until equilibrium with chalcedony is reached. If the ashy layer is near the surface and the upwelling thermal water has been cooled by heat loss to the surrounding rock, the estimated base temperature using the chalcedony geothermometer will approximate the discharge temperature of the hot spring, as it does at Jackson. The silica loss due to precipitation will yield anomalously low results using the quartz geothermometer and true reservoir temperature may be more accurately estimated by the Na-K-Ca geothermometer.

Discussion

Of the three models discussed, the 'warm' model is least favored because it relies principally on a close agreement between the chalcedony and Mg-corrected Na-K-Ca geothermometers. Use of the magnesium correction may not be valid, as discussed earlier. In addition, other evidence indicate that reservoir water in the Jackson hydrothermal system may be in equilibrium with quartz rather than chalcedony. Silica-rich ash beds are likely present in the Tertiary fill in the Jackson area and the geochemical data from sample locations east of Jardine Hot Spring indicate that near-surface water may be encountering such beds. The presence of siliceous beds and thermal water in equilibrium with quartz could account for the geochemical conditions present at Jackson. In short, problems exist with both the chalcedony geothermometer and the use of the magnesium correction to the cation geothermometer. The close agreement between these two geothermometers may merely be coincidence, and the 'warm' model, based on this close agreement, may be suspect.

It is unclear whether the 'precipitating' or 'mixing' model most accurately represents the Jackson hydrothermal system, and further research is required before either model can be demonstrated. The 'precipitating' model is attractive because its chemical relationships are the least complicated and it is consistent with the presence of ash-rich layers in the Tertiary or Quarternary sediment fill. The 'mixing' model is supported by a variety of chemical evidence as well as by the location of Jardine Hot Spring. The supporting chemical data includes the agreement between the uncorrected cation geothermometer temperature estimate and the estimated temperature of the hot water component using the silica enthalpy technique.

Regardless of which model is correct, temperatures higher than those indicated by the chaldedony or Mg-corrected cation geothermometers are implied. Although the 'precipitating' model yields a minimum estimated temperature of 100°C, the implied loss of silica due to precipitation during the ascent of hot geothermal water implies higher reservoir temperatures. The 'mixing' model yields an estimated minimum reservoir temperature of 145°C. Therefore, if one of these models approximates the geochemical conditions of the Jackson hydrothermal system, the minimum reservoir temperature of the system is likely to be between 100°C and 145°C.

CHAPTER 6

HEAT SOURCE AND HYDROTHERMAL CIRCULATION AT JARDINE HOT SPRING

The hydrothermal system at Jackson is a liquid-dominated convective system. Two critical parameters necessary for the understanding of this type of system are the heat source and hydrothermal circulation path of the system. In this chapter, the heat source for the Jackson system as well as possible pathways for recharge and discharge are presented and a hydrothermal circulation model, based on the geochemical and geological data from this study, is proposed.

Heat Source

Convective hydrothermal systems derive their heat from one of two main sources; from deep circulation along faults or permeable aquifers in a regime of high regional heat flow, or from circulation of water near a cooling igneous body. With the possible exception of springs adjacent to Yellowstone National Park (Struhsacker, 1976), hot springs in southwestern Montana, including Jardine Hot Spring, probably derive heat from deep circulation (Chadwick and Leonard, 1979). Three lines of evidence support a deep circulation heat source. One is the spatial coincidence of hot spring concentration and complex structural fabric in southwestern Montana. Another is that the ages of the igneous rocks in the region are too great to provide magmatic heat sources for hydrothermal systems. A third line of evidence is the relative uniformity of estimated reservoir temperatures of the hydrothermal systems in the region, which argues against magmatic heat sources of differing ages and magnitudes.

The vast majority of hot springs in Montana occur in the southwestern part of the state, which is also the most structurally complex region in the state. This spatial overlapping of hot spring concentration and structural complexity suggests that regional structural features may influence hot spring distribution (Chadwick and Leonard, 1979). Southwestern Montana is characterized by large scale folds and thrusts, large granitic batholiths, numerous intermontane basins, and regional structural lineaments. Of the twenty-seven hot springs in Montana that discharge water hotter than 38°C--the temperature used to separate 'hot' from 'warm' springs(Meinzer, 1923; Mariner and others, 1976)--all but two occur in the southwestern part of the state (Chadwick and Leonard, 1979). Because this concentration of hot springs coincides spatially with the complex overlapping of major structural elements, the distribution and hydrothermal circulation of hot spring systems appears to be structurally controlled. Such structural control on a regional level lends credence to the concept of deep circulation along regional fractures as the primary heat source for hot springs in southwestern Montana.

The age of the igneous rocks in southwestern Montana also lends support to the deep circulation concept. Igneous rocks in Montana are chiefly Cretaceous to Oligocene in age. The youngest igneous rocks in the Pioneer Batholith, for example, are of Eocene age (Snee and others, 1981). Igneous rocks younger than Miocene are limited to extrusives near Yellowstone National Park. As discussed in Chapter 4, the largest igneous bodies are capable of supporting thermal anomalies for only 10 million years or less (Smith and Shaw, 1975). Therefore, the igneous rocks of southwestern Montana are too old to provide magmatic heat sources for hydrothermal systems.

The concentration of estimated reservoir temperatures (using the silica geothermometer) of Montana thermal springs into two main groups (Chadwick and Kaczmarek, 1975) is additional evidence against an igneous heat source for hydrothermal systems. In one group, estimated base temperatures average 20°C. The second group averages between 100°C and 120°C. Kaczmarek (1974) and O'Haire (1977) conclude that the high temperature group probably derives heat from deep circulation. If springs in this group were associated with cooling igneous bodies of different ages, sizes and depths, widely varying temperatures would be expected.

From the foregoing evidence, deep circulation in a thermal regime of regional heat flow is the likely primary source of heat for Jardine Hot Springs. Consequently, the regional heat flow in southwestern Montana and its magnitude (geothermal gradient) in the Jackson area are important to the understanding of the hydrothermal system at Jackson. A discussion of these two topics is presented below.

Regional Heat Flow and Geothermal Gradients

Eaton and others (1976) places southwestern Montana within a region of abnormally high heat flow known as the Cordilleran Thermotectonic Anomaly (CTTA). Geothermal gradients in this region are high, averaging 30°C/km (Blackwell, 1969; Blackwell and Robertson, 1973). While part of the heat flow in the CTTA, especially in the southern portion, may be due to crustal thinning resulting from extensional tectonics, the concentration of major granitic batholiths in southwestern Montana and east-central Idaho also contributes to the high heat flow. Although the thermal effects associated with the intrusion of these batholiths have long since dissipated, heat generated by radiogenic decay is continuing. Radioactive elements such as thorium and uranium have a strong tendency to form oxides and silicates in granitic bodies (Press and Siever, 1974). The heat produced by the decay of these elements is a major component of the heat flow in granitic bodies (Birch and others, 1968). Although such a heat source alone is not capable of supporting a hydrothermal system (Blackwell, 1969), it may add significantly to the regional heat flow in southwestern Montana.

Blackwell and Chapman (1977) found that the geothermal gradient near granitic crystalline rocks in western Montana averages 30°C/km. They also suggest that in some of the deep basins in western Montana the geothermal gradients increase to 57°C/km (Blackwell and Chapman, 1977). These basins are generaly marginal to granitic masses and are filled with thick deposits of Tertiary and Quaternary sediments and rocks that are often poorly consolidated. Poorly consolidated sediments have low thermal conductivity relative to crystalline rock, acting as an effective insulator, and may increase the geothermal gradient in some of these basins. Whether an increase to the 57°C/km range suggested by Blackwell and Chapman (1977) actually occurs, however, is uncertain. A well in the west-central Deer Lodge Valley, the Montana Power No. 1-1-22 State in Section 22, T. 7 N., R. 10 W., drilled through 2,536 feet of Tertiary rocks and encountered a geothermal gradient of 30°C/km. A well in the east-central Beaverhead Valley, the North American Resources No. 3-8 State in Section 8, T. 4 S., R. 5 W., drilled through more than 3,000 feet of Tertiary rocks and encountered a geothermal gradient of 35°C/km. It is possible that increasing induration of Tertiary sediments with depth results in an increase in their thermal conductivity and a consequent decrease in the expected geothermal gradient. Projecting the thermal conductivity properties of poorly consolidated surficial Tertiary rocks to significant depths may yield estimated geothermal gradients in Tertiary basins that are too high.

Another example of this type of basin is in the Big Hole Basin. It is adjacent to the Pioneer Batholith and is filled with nearly 16,000 feet of Tertiary rocks on its western side. The geothermal gradient of this basin calculated from the 16,028 foot depth of the Amoco No. 1 Hershey well on the western side of the basin is 33°C/km. As with the two wells mentioned above, bottom hole temperature measurements are not equilibrium temperatures, having been influenced by the introduction of relatively cool drilling mud during drilling, and geothermal gradients calculated from non-equilibrium bottom hole temperatures should be considered minimum values. The close agreement between these three Tertiary basin wells, however, indicates that the actual geothermal gradients are probably close to those estimated by bottom hole temperature measurements. The calculated geothermal gradient of

33°C/km for the Big Hole Basin is used as a minimum value for this study.

Hydrothermal Circulation Model

For a model of the geothermal system at Jackson to be complete it must include, in addition to a proposed heat source and estimated reservoir temperature, the estimated depth of circulation and probable discharge and recharge routes. The geochemical evidence discussed in Chapter 5 suggests three minimum reservoir temperature estimates ranging between 70°C and 145°C, with minimum temperatures between 100°C and 145°C deemed most likely. In order to heat the circulating water to minimum temperatures of 100°C and 145°C in a regime of regional heat flow with a minimum geothermal gradient of 33°C/km, the depths of circulation must be approximately 2.9 km (9,628 feet) and 4.3 km (14,104 feet) respectively, assuming a mean surface temperature of 4.5°C. Using the less likely minimum reservoir temperature estimate of 70°C would require a circulation depth of approximately 2.0 km (6,560 feet). Proposed circulation routes for the Jackson hydrothermal system must be consistent with these estimated depths of circulation.

A circulation model for the Jackson system has been indirectly proposed by Robertson and others (1976). It is their contention that alignments of hot springs in western Montana coincide with the traces of long, linear, deep-seated shear zones and that hydrothermal circulation occurs within the fault breccia of these shear zones. One of these hot spring alignments trends N79°E for 320 km and extends from Big Creek Hot Springs in Idaho, through Jackson to Hunters Hot Spring in Montana (Figure 9). This alignment is not associated with any known structural features. However, an alignment that trends N10°E for 120 km from New Biltmore to Alhambra Hot Springs appears to be associated with known or inferred faults along the eastern edge of the Boulder Batholith (Chadwick and Leonard, 1979). The association of proposed hot spring alignments with deep-seated regional shear zones that control hydrothermal circulation, as postulated by Robertson and others (1976), is tenuous and requires additional evidence.

In the hydrothermal circulation model presented here, proposed routes for recharge and discharge are treated separately. The discussion of the proposed discharge route is concerned mainly with fault control of thermal spring location within the study area. A more general treatment of the hydrothermal circulation is found in following sections.

Fault Control of Discharge

Discharge of thermal water in the Jackson area probably occurs along the inferred normal fault that trends generally northward through the study area (Plate 1). In addition to the evidence for this fault provided in Chapter 3, a study by Nathenson (1981) supports the probability that thermal water is discharged along a fault at Jackson. He uses two end-member models, one involving the ascent of thermal water up a cylindrical conduit and the other involving flow up a planar fault, to explain the relationship between estimated reservoir temperatures, measured flow rates, and discharge temperatures of hot springs in southwestern Montana. As thermal water flows upward along

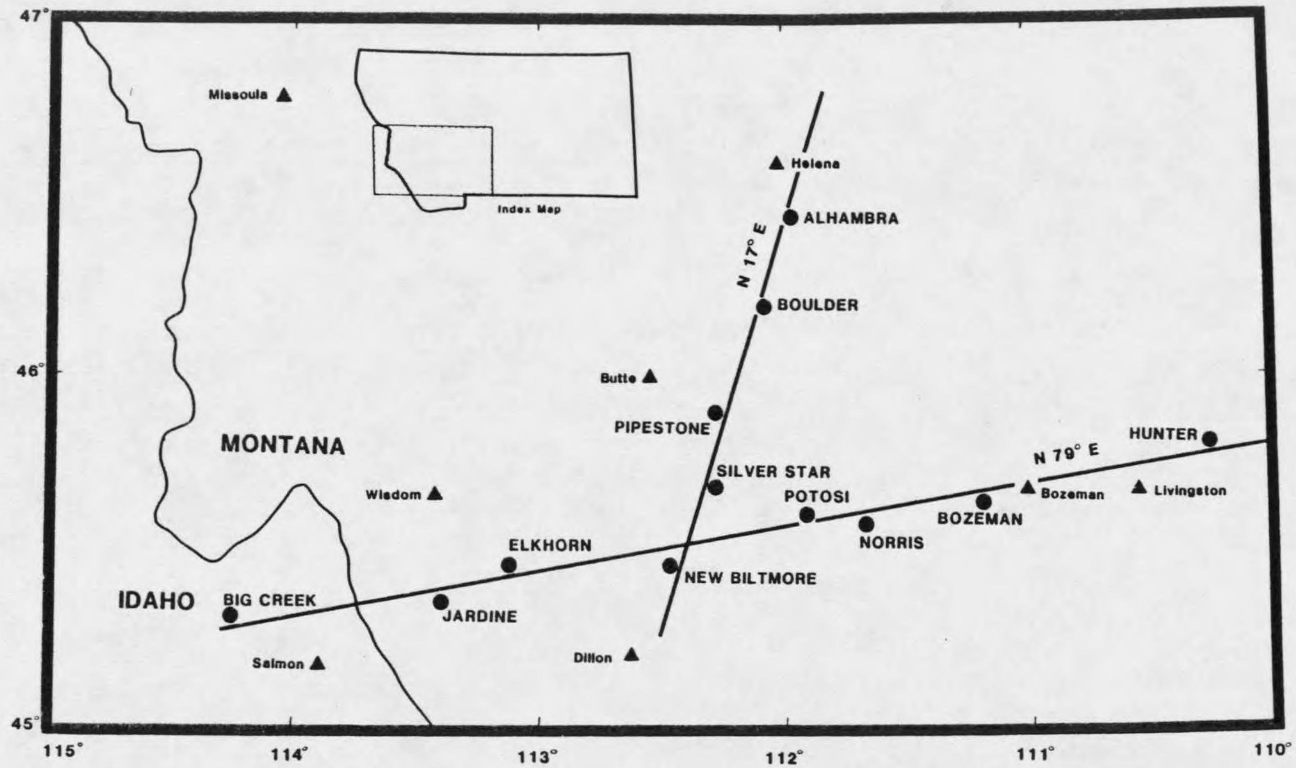

Figure 9. Map of hot spring alignments in southwest Montana.
(From Robertson and others, 1976)

the discharge route, heat loss must occur in order for the water to cool from reservoir temperature to discharge temperature. Nathenson (1981) assumes that heat loss is accomplished primarily through heat transfer to surrounding wall rock during the ascent of thermal water. More heat loss can be accomplished by flow up a planar fault than by flow up a conduit due to the greater surface area of a fault. Using the estimated Mg-corrected Na-K-Ca geothermometer estimated reservoir temperature of 69°C, which is probably too low (as discussed in Chapter 5), a measured discharge temperature of 58.5°C, and an estimated discharge of 60 liters/second, Nathenson (1981) estimates that the heat loss at Jardine Hot Spring is greater than that which could take place through flow up a cylindrical, conduit-like discharge route. He models the heat loss at Jardine Hot Spring as occurring along a fault plane with dimensions of at least 2 km by 2 km. If reservoir temperature is higher than that estimated by Natheson (1981), as is indicated by the geochemical evidence from this study, even greater heat loss would be required. Larger fault dimensions and/or dilution with cold groundwater at shallow depths could provide the necessary additional heat loss.

Additional supporting evidence for hydrothermal discharge along a planar fault rather than along a cylindrical conduit is the presence of the small group of warm springs northwest of Jackson and the linear zone of rapid snow melt that occurs between these warm springs and Jardine Hot Spring. The warm springs are located 2 km from Jardine Hot Spring and the minimum size of a fault connecting them would correspond well to the minimum fault size modeled by Natheson (1981). In addition, the thermal water discharging at Lapham's Warm Springs is chemically very similar to water from Jardine Hot Spring (Figure 10). Samples from these two thermal springs are much higher in sodium, bicarbonate, sulfate, and arsenic than are samples from the other locations used in the geochemical study with the exception of Lapham's Domestic Well which is located less than 300 feet from Lapham's Warm Springs. Like Jardine Hot Spring, the warm springs also discharge large amounts of gas. The chemical similarities of waters from Jardine Hot Spring and from Lapham's Warm Springs indicates that these waters are part of the same hydrothermal system. The presence of a linear zone between these thermal springs along which rapid snow melt occurs indicates that thermal water probably discharges between the thermal springs as well. The discharge of thermal water along this zone may be masked at the surface by the dispersion of thermal water within permeable Tertiary or Quaternary sediments and/or by dilution with cool, shallow groundwater flowing down hydrologic gradient from the highlands east of Jackson toward discharge areas along the Big Hole River.

It appears, therefore, that the discharge of thermal water from the Jackson hydrothermal system occurs along a linear fault zone on the southeast margin of the Big Hole Basin. Ascent of thermal water up the fault occurs along a zone at least 2 km in length and which connects Jardine Hot Spring and Lapham's Warm Spring to the northwest. A major portion of the hydrothermal discharge probably issues from Jardine Hot Spring, with lesser amounts issuing from Lapham's Warm Springs. A portion apparently issues along a linear zone between the thermal

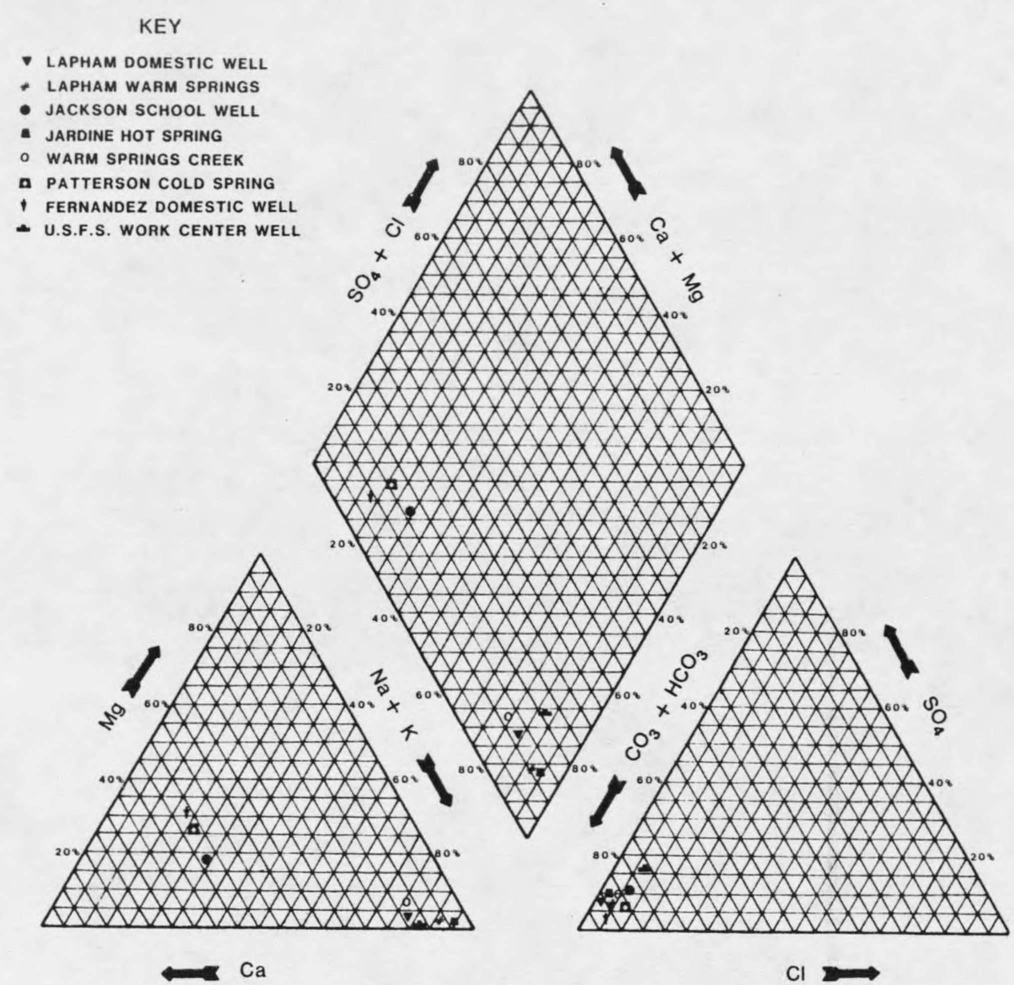

Figure 10. Piper diagram of ion content of waters sampled in the study area. (After Piper, 1944)

springs as well. Cooling from the probable minimum reservoir temperatures of 100°C to 145°C is probably accomplished by heat transfer to surrounding wall rocks during ascent and possibly accompanied by mixing with cold groundwater near the surface.

Recharge Routes

Two basic models of recharge paths for the Jackson hydrothermal system are consistent with the geological and geophysical data. In Model 1 (Figure 11), recharge areas are east of Jardine Hot Spring and descent of recharge water occurs primarily along the plane of the Jackson thrust fault. In Model 2 (Figure 12), recharge areas are on the western margin of the Big Hole Basin, and recharge occurs down an inferred fault zone on the west side of the basin.

Model 1, recharge areas are in the highlands east of Jackson. The Belt metasedimentary rocks on the western flank of the Pioneer Mountains possess abundant fractures, with minor intergranular porosity, that will permit groundwater percolation. The intrusive rocks of the western Pioneer Mountains possess jointing in addition to fracture permeability. Jointing and fracturing can provide important means of recharge, especially where matrix porosity and permeability is low. Groundwater from the Pioneer Mountains may percolate down hydrodynamic gradient primarily by means of joint and fracture permeability toward the Big Hole Basin. Some groundwater moving toward the Big Hole Basin will be discharged from springs in the highlands east of Jackson, but most will likely continue to move down-gradient in the subsurface.

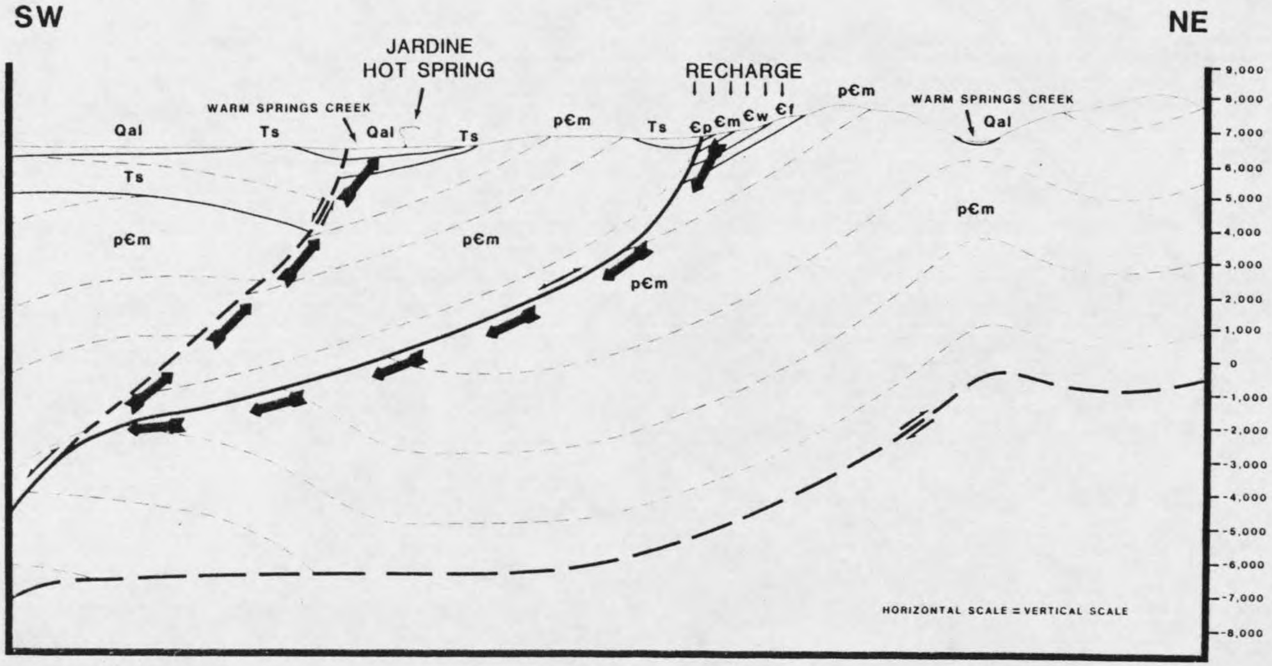

Figure 11. Cross section of Jardine Hot Spring area showing local hydrothermal convection route.

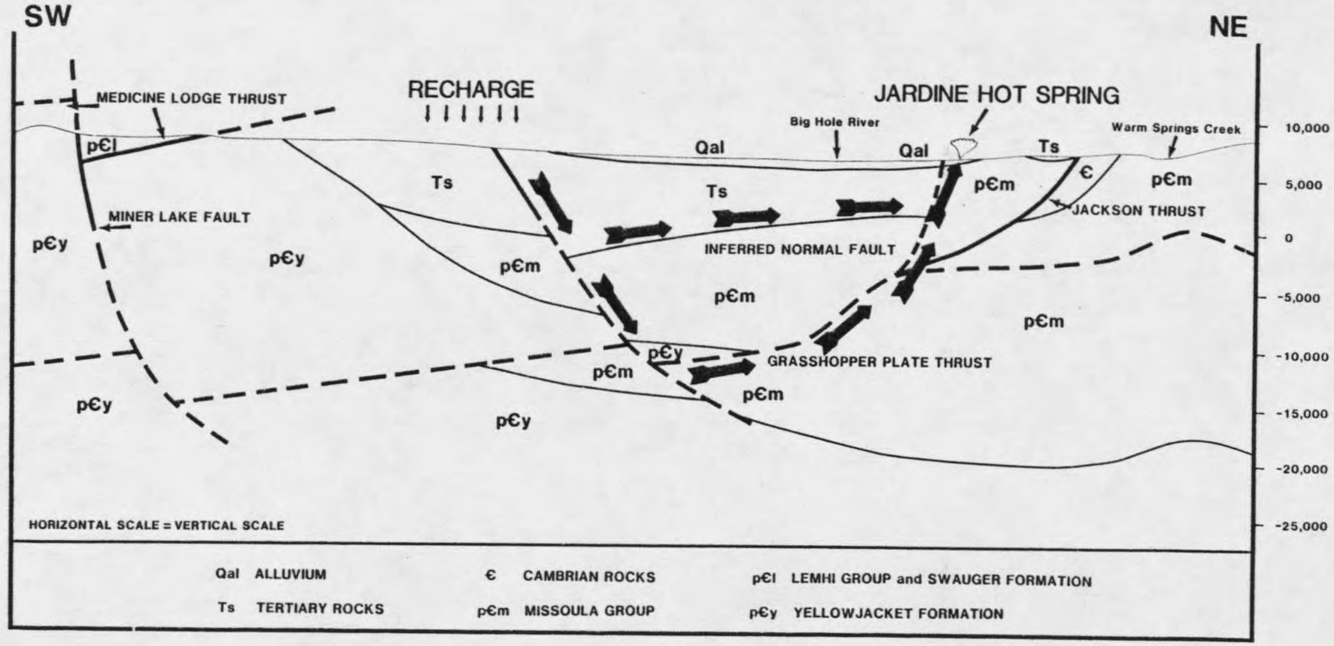

Figure 12. Schematic cross section of southern Big Hole Basin showing regional hydrothermal convection routes.

Groundwater from the highlands of the eastern part of the study area will intersect the plane of the Jackson thrust fault. Post-thrusting extension of the fault probably increased the permeability of the fault, allowing groundwater to descend along the fault plane. Although the surface distance between this fault plane and the hot spring is only two miles, the steep dip of the fault (65° to 75° could allow deep circulation and heating of the water to the temperatures between 100°C and 145°C that are indicated by the geochemical data. The required circulation depth of approximately 10,000 feet needed to attain a reservoir temperature of 100°C is shown in Figure 11. A less rapid flattening of the Jackson thrust fault would increase circulation depth and corresponding reservoir temperature. The precise fault geometry is not known due to a lack of well data and published geophysical data for the area.

In Model 2, the possible recharge route would channel cold water from the highlands in the northern Beaverhead Mountains west of the Big Hole Basin down an inferred fault zone on the western margin of the basin. Groundwater may also percolate through poorly consolidated Quaternary glacial sediments and Tertiary basin fill on the west side of the basin and intersect the fault zone. Water descending the fault zone will become heated during descent. Heated recharge water from the fault zone may follow the basin floor updip, encounter the inferred normal fault on the east side of the basin, and be discharged. Alternatively, the heated recharge water may continue down the fault zone on the western side of the basin until it intersects the basal decollement of the Grasshopper thrust plate (Chapter 3). Thermal water may then ascend the decollement, encounter the inferred normal fault on the east side of the basin, and be discharged along this fault. The amount of Tertiary rocks in the southern part of the basin is thought to be less than that contained in the northwestern part of the basin, as discussed below. Therefore, water flowing along the basin floor at the base of the Tertiary fill, is estimated to reach circulation depths of approximately 10,000 feet. Water descending to the basal fault of the Grasshopper thrust plate may descend to depths up to approximately 16,000 feet.

Although extensive glacial deposits cover the inferred trace of the fault zone on the western margin of the Big Hole Basin, its presence is indicated by well and seismic data in the western part of the basin. Information from the Amoco Productions Company's Jack Hirshey Livestock Corporation No. 1 well in Section 27, T. 3 S., R. 16 W., indicates that the northwest part of the Big Hole Basin contains approximately 16,000 feet of Tertiary basin fill sediments. The highlands just west of the No. 1 Hirshey well lack Tertiary basin fill rocks, indicating that the margin of the basin must dip at least 60° eastward in order to accommodate the thickness of Tertiary rocks encountered in the No. 1 Hirshey well. Southeast of the No. 1 Hirshey well, Amoco drilled a second deep well, the No. 2 Hirshey well, in the central part of the Big Hole Basin. The No. 2 Hirshey well drilled through approximately 9,500 feet of Tertiary rocks before encountering Precambrian rocks. The information from these wells indicate that the floor of the basin probably shallows to the east and may shallow to the south as well. This is supported by proprietary seismic data from the northern part of the basin that indicates the basin is a half-graben that deepens to the west where it is bounded by a steeply dipping fault.

Summary

In summary, the circulation pattern of the Jackson hydrothermal system takes place within a regime of regional heat flow with a geothermal gradient estimated to be a minimum of 33°C/km. Geochemical data indicates that minimum circulation depths are likely to be between approximately 9,600 feet and 14,000 feet. Discharge probably occurs along an inferred normal fault that connects Jardine Hot Spring with a small group of thermal springs 2 km to the northwest. Two recharge routes can be postulated, one on the eastern margin of the Big Hole Basin and the other on the western margin of the basin. A recharge route on the eastern side of the basin would indicate that the Jackson hydrothermal system is a local system, with primary recharge and discharge routes separated by a distance of only two miles at the surface. Recharge on the western side of the basin would require thermal water to circulate beneath the entire width of the Big Hole Basin, indicating the presence of a more regional hydrothermal system. The apparent lack of thermal springs in the basin other than those near Jackson supports the presence of a localized hydrothermal system. Consequently, a recharge route on the east side of the Big Hole Basin is favored by the evidence from this study. The hydrothermal circulation model shown in Figure 11 is consistent with the geological and geochemical data from this study and probably closely approximates the hydrothermal system at Jackson.

CHAPTER 7

THE ECONOMICS OF GEOTHERMAL DEVELOPMENT

This chapter presents an overview of moderate-temperature hydrothermal resource utilization and discusses the suitability of the geothermal resource at Jackson for various types of potential uses. In addition, economic evaluations are provided for two potential uses--district space heating and power generation. These analyses identify some of the internal development costs of each project and estimate annual net revenue in order to provide a preliminary assessment of each project's attractiveness for potential investors. Because factors such as external costs, royalties, taxes, and government incentives are not addressed, the analyses provided are not intended as rigorous feasibility studies.

Moderate-Temperature Resource Utilization

Historically, geothermal resource utilization has been divided into two broad categories; direct use (non-electric) applications and electric power generation, measured in Megawatts thermal (MWt) and Megawatts electric (MWe) respectively. Direct utilization of geothermal energy means that geothermal heat is used directly and is not converted to other forms of energy such as electricity (Anderson and Lund, 1979). Utilization for conventional power generation has been confined primarily to resources with temperatures generally in excess of 150°C. Direct use applications utilize low to moderate temperature resources ranging from about 20°C to 180°C.

On a worldwide basis, the direct utilization of geothermal resources is significantly greater than utilization for power generation. In 1979, for example, direct use of geothermal energy amounted to approximately 7,000 MWt compared to about 1,800 MWe utilized for the production of electricity (Lineau, 1980). Of the 7,000 MWt for direct use, over 12,000 MWt are used for space heating and cooling; approximately 5,500 MWt for agriculture, aquaculture, and animal husbandry; and over 200 MWt for industrial processing (Lineau, 1980).

In the United States, geothermal energy utilization has emphasized high temperature resources suitable for power generation. This emphasis is primarily due to the remoteness of geothermal resources from urban areas in this country compared to countries such as Japan, Hungary, and Iceland where direct utilization is emphasized. Direct use projects must be located near resource locations because significant heat loss increases rapidly with transportation distance. Conversely, electricity produced from geothermal resources can be transported with relative ease from the resource site to urban markets. In recent years, however, interest in developing moderate temperature resources in this country has increased and the number of direct use projects has grown, particularily in the West where geothermal resources are concentrated.

Direct Use Applications

Direct utilization of geothermal resources is economically attractive for a number of reasons. First, direct use applications generally require lower temperature resources than does conventional electrical power production and these low temperature resources are much more abundant than high temperature resources. Second, much of the energy contained in geothermal resources is lost when it is converted to electricity - directly using geothermal heat has a much higher energy conversion efficiency. Third, most direct use applications can be economically operated at a much smaller scale than conventional power generation projects, resulting in smaller initial costs and reduced time for planning and implementation.

Most direct use applications fall into three categories; agricultural-related uses, industrial processing, and space heating and cooling. Temperature requirements are generally lowest for agricultural uses and are highest for industrial processing (Table 4).

Agriculture-related projects account for the majority of direct use applications of geothermal heat. The principal agricultural application is greenhouse heating. Major greenhouse operations exist in the Soviet Union, Hungary, and Japan. Geothermal energy is also used on a large scale as a heat source for raising livestock, primarily poultry and other small animals. In addition, geothermal water is used in aquaculture operations. Hatcheries produce shrimp, eels, catfish, trout, and other fish and require only slightly heated water (Armstead, 1978). In Idaho, approximately 500,000 pounds of catfish are raised

Table 4. Temperatures of geothermal fluids required for various agricultural and industrial processing uses. (After Rinehart, 1980).

Temperature (°C)	Uses
180°	Evaporation of concentrated solutions Refrigeration by ammonia absorption
170°	Heavy water via hydrogen sulfide process Drying of diatomaceous earth
160°	Drying of fish meal Drying of timber
150°	Alumina via Bayer's process
140°	Drying of farm products at high rates Canning of food
130°	Evaporation in sugar refining Extraction of salts by evaporation
120°	Fresh water by distillation Most multiple effect evaporations
110°	Drying and curing of light aggregate cement slabs
100°	Drying of organic materials; seaweeds, grass, vegetables
90°	Drying of fish stock Intense de-icing operations
80°	Space heating; Greenhouses by space heating
70°	Refrigeration (lower temperature limit)
60°	Animal husbandry Greenhousing by space and hotbed heating
50°	Mushroom growing Balneological baths
40°	Soil warming
30°	Swimming pools, biodegredation, fermentations Warm water for year round mining, de-icing
20°	Hatching and raising of fish

annually using 32°C water (Lineau, 1980). Other agricultural applications include using geothermal heat for mushroom raising and for soil warming for certain crops and trees to increase growth rates (Anderson and Lund, 1979).

In addition to agricultural uses, geothermal heat is used in the processing of a variety of industrial and agricultural products. Many processing projects require significantly higher temperatures than agricultural applications, with temperatures ranging to 150°C often needed (Anderson and Lund, 1979). A wide variety of processing procedures currently utilize geothermal heat. Among the largest projects are a timber, pulp and paper drying operation in New Zealand (Lineau, 1980) and a diatomaceous earth processing project in Iceland (Armstead, 1978). In Nevada, the world's first geothermal vegetable dehydration unit uses 120°C water for onion dehydration. Other potential processing uses for geothermal heat include preheating, washing, sterilizing, and refrigeration (Anderson and Lund, 1979).

Geothermal energy may also be used directly for space heating and cooling. Conventional space heating systems such as forced air or circulating water systems are compatable with a geothermal heat source. Cooling applications utilize the heat-absorption cycle of available heat pumps and can use hot water with temperatures between approximately 80°C and 130°C without extensive custom design work (Anderson and Lund, 1979). When using a geothermal heat source for space conditioning, existing systems must be converted (retrofitted) to the new energy source or new systems must be designed and installed.

Forced-air systems are the most readily retrofitted and circulating water systems are generally favored for new systems (Engen, 1978).

Currently, most space conditioning applications of geothermal energy are large-scale district heating projects. The largest space-heating project is the Reykjavik municipal heating system in Iceland. Geothermal water is used in a large, centralized system to heat 16,000 homes (Lineau, 1980). A more de-centralized system is in place in Klamath Falls, Oregon where over 400 wells are used.

The suitability of the Jackson geothermal resource for processing or agricultural uses appears limited, primarily by the estimated size of the resource and its remote location. The surface resource, with a measured temperature close to 60°C and a flow rate of 60 liters per second (950 gpm), is not sufficient to support most processing applications. Expected reservoir temperatures in the range of 100°C to 145°C increase the number of potential processing uses of the resource, but the locations of supplies of raw materials and potential markets for finished processed products would require long transportation distances. For example, only one town with a population over 20,000 is located within a 100 mile radius of Jackson. Although the Jackson resource is suitable for a medium-scale greenhouse or aquaculture operation, such an operation would also have difficulty overcoming the transportation costs of supplies and products.

District space heating for the town of Jackson appears more promising due to the town's proximity to the resource. The potential for utilizing the geothermal resource at Jackson for a small-scale district heating project is evaluated later in this chapter.

Power Generation

In addition to direct use applications, interest in power generation using moderate temperature geothermal resources has also increased, primarily due to technical advances which have greatly increased the thermodynamic efficiency of extracting power from low-enthalpy energy sources. Of particular interest to geothermal development are refinements in the Rankine cycle engine, used by major thermal power-generating plants, which employes a working fluid that absorbs heat, vaporizes, produces work, and condenses in a closed system. In conventional power generation systems, the working fluid is water, principally because of its low cost. Water is not an efficient fluid to use with low to moderate temperature heat sources, however, because steam has a very low power density at low temperatures. The low molecular weight of steam requires enormous volumes of steam at low temperatures to produce power in commercial quantities, necessitating the use of extremely large and expensive expansion engines, pumps, pipes, condensers, and related equipment. When high density fluids such as freon, isobutane, or isopentane are used as the working fluids, the high molecular weight of the vapor in the cycle enables more work to be produced during vaporization with less vapor flow than can be produced by steam at similar temperatures. The development of Rankine cycle engines using high-density working fluids, therefore, enables moderate temperature resources to be used for power generation much more economically than has been possible with conventional techniques.

The geothermal resource at Jackson appears suitable for small-scale power generation, although hotter water temperatures are needed than exist at the surface. A survey of manufacturers of Rankine cycle generators determined that resources with water temperatures less than about 65°C necessitate the use of enlarged heat exchangers and other costly equipment, making these resources uneconomical. The surface temperature and flow rate of 60°C and 60 liters per second (950 gpm) at Jardine Hot Spring are, therefore, insufficient. Geological and geochemical evidence from this study (Chapter 6), however, indicates that subsurface temperatures in the range of 100°C to 145°C can be expected. A decrease in flow rate may be experienced by a geothermal well, especially if some degree of shallow mixing of hot and cold water is occurring near the surface (Chapter 6). A subsurface temperature of 100°C and a flow rate of 32 liters per second (500 gpm), therefore, are consistent with the observations from this study. These parameters, then, are used in a preliminary economic evaluation of the Jackson resource for power generation.

Economic Evaluation of District Heating

The following section assesses the district heating needs of the Jackson community and the ability of the geothermal resource to meet those needs. In addition, a preliminary estimate of a district heating project's initial (front-end) and annual costs is made to determine the feasibility of a geothermal district heating system for Jackson.

Because climatological data, used extensively in this section of the study, is measured in degrees Fahrenheit by the National Oceanic and Atmospheric Administration and because specifications for applicable equipment and materials are generally not available in SI units, units of measurement in this section are given in the English system.

Heating Requirements

District heating requirements for Jackson are estimated using the methods of Engen (1978), who defines an average residential unit as having 1,800 square feet of floor space with average construction and energy efficiency. The averages appear to be a close approximation of size and construction variables in Jackson. Factors that must be considered when estimating the heating demand for an area include; the minimum outside design temperature (t_o), the desired inside temperature (t_i), and the annual Fahrenheit degree days (ADD) for the project area (Engen, 1978). The minimum outside design temperature for the Jackson project is taken to be -25°F. The inside design temperature is assumed to be 70°F. The annual degree days (ADD) for Jackson was determined by averaging data from the National Oceanic and Atmospheric Administration's Jackson weather station, recorded from 1955 through 1972 (Table 5). Parameters calculated using ADD data are the maximum heat load (HL_{max}), the annual heat load (HL_a), the delivered heat (H), and the fluid flow rate required from the geothermal resource (w) (Table 6). The equations used apply to one average-efficiency residential unit of 1,800 square feet.

When calculating the required flow rate per residential unit, it is necessary to estimate the temperature drop across the required heat

Table 5. Annual degree days for Jackson, Montana*

Year	Jan.	Feb.	Mar.	Apr.	May	June	Jul.	Aug.	Sept.	Oct.	Nov.	Dec.	Total
	Month												
1955	1628	1417	1440	1014	689	432	249	192	498	763	1324	1458	11,104
1956	1508	1412	1234	879	550	389	230	319	456	807	1210	1263	10,277
1957	1745	1145	1191	933	548	396	183	236	440	807	1203	1256	10,083
1958	1405	1018	1320	965	436	383	290	167	501	729	1119	1199	9,532
1959	1356	1270	1206	861	786	346	253	306	570	805	1161	1325	10,245
1960	1583	1448	1205	884	721	410	137	353	454	806	1136	1407	10,544
1961	1270	1048	1129	968	612	234	181	169	682	826	1149	1474	9,742
1962	1607	1245	1276	805	666	406	298	358	486	704	1060	1251	10,162
1963	1673	1024	1159	968	596	446	275	239	344	688	1082	1384	9,878
1964	1481	1368	1383	963	691	445	171	312	566	725	1148	1475	10,728
1965	1400	1211	1487	869	769	442	257	301	715	676	1020	1350	10,497
1966	1436	1298	1216	930	601	469	176	269	388	824	1043	1380	10,030
1967	1321	1193	1219	1043	689	425	178	189	372	831	1114	1603	10,177
1968	1493	1216	1062	1014	762	448	223	394	578	866	1211	1569	10,836
1969	1461	1406	1370	801	530	441	243	169	428	1002	1012	1331	10,194
1970	1402	1148	1330	1184	681	307	178	193	625	(976)	1184	1512	10,720
1971	1272	1250	1302	(915)	(627)	450	284	128	622	(909)	(1163)	1677	10,599

* Jackson weather station: Elevation 6477', Latitude 45°22'N, Longitude 113°25'W. Figures in parentheses from the Wisdom weather station: Elevation 6058', Latitude 45°37'N, Longitude 113°27'W.

exchangers. Using the measured resource temperature of 140°F, the temperature drop is 12°F as determined from Figure 13. The required flow rate, therefore, is calculated to be 15.83 gallons per minute of 140°F geothermal water per residential unit. There are approximately 40 units in Jackson, including small businesses. Assuming 100% market penetration, the total system requirement is 633 gpm. Because the flow rate of Jardine Hot Spring is approximately 950 gpm, the resource as it exists at the surface is adequate to supply the needs of the district heating system.

Table 6. Calculated Resource Requirements

Maximum Heat Load

$$
\begin{aligned}
HL_{max} &= 800 \text{ Btu/hr } (t_i - t_o) \\
&= 800\ [70° - (-25°)] \\
&= 76{,}000 \text{ Btu/hr}
\end{aligned}
$$

Annual Heat Load

$$
\begin{aligned}
HL_a &= 19{,}200 \text{ Btu (ADD/year)} \\
&= 19{,}200 \text{ Btu } (10{,}315/\text{year}) \\
&= 1.98 \times 10^8 \text{ Btu/year}
\end{aligned}
$$

Delivered Heat

$$
\begin{aligned}
H &= 1.25\ (HL_{max}) \\
&= 1.25\ (76{,}000 \text{ Btu/hr}) \\
&= 95{,}000 \text{ Btu/hr}
\end{aligned}
$$

Fluid Flow Rate*

$$
\begin{aligned}
w &= \frac{H}{500\ (\Delta t)} \\
&= \frac{95{,}000}{500\ (12)} \\
&= 15.83 \text{ gpm}
\end{aligned}
$$

* Where Δt is the estimated heat exchanger temperature drop for the appropriate resource temperature.

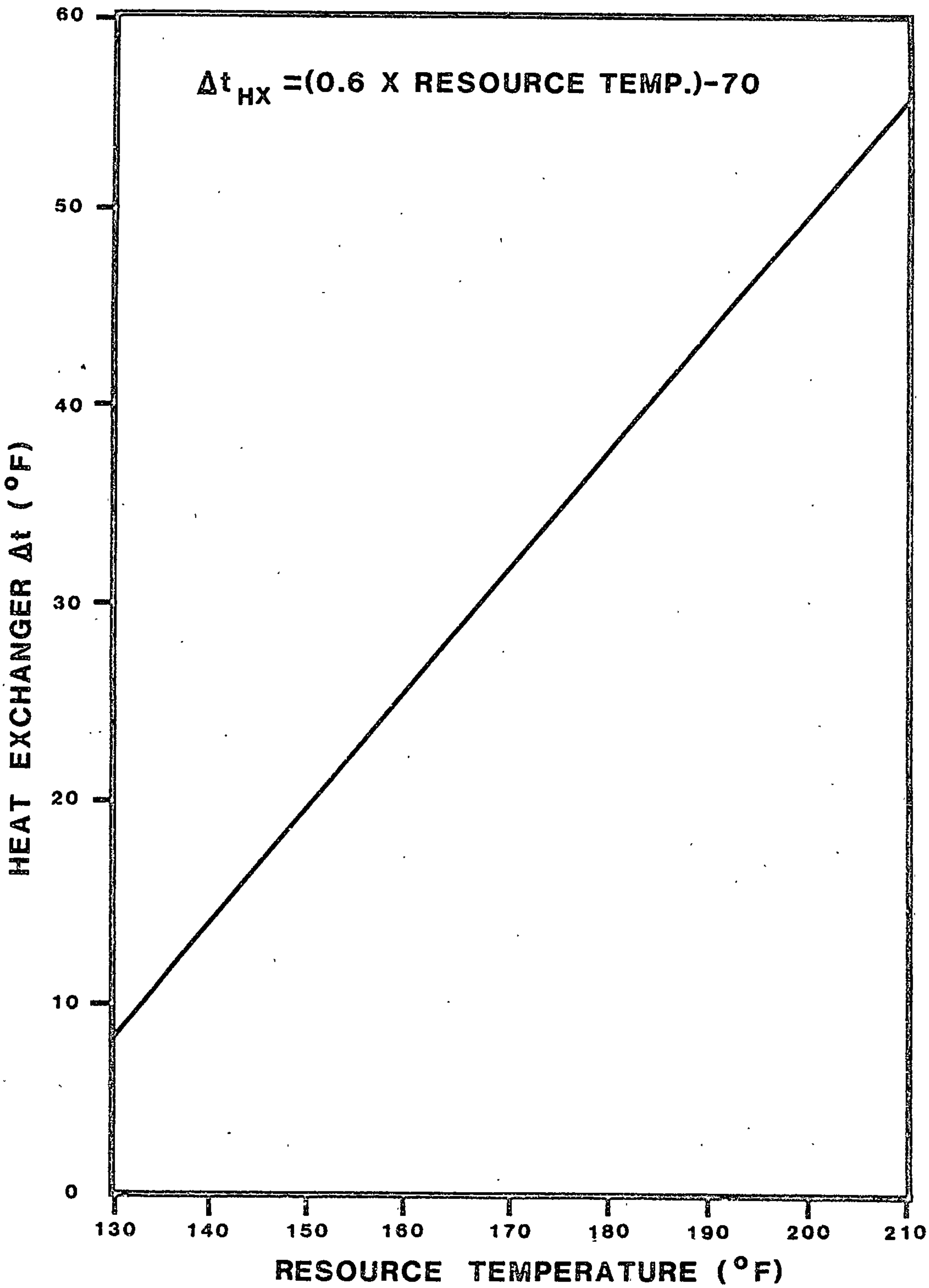

Figure 13. Required resource temperature vs. heat exchanger temperature drop. (After Engen, 1978)

Project Costs

The initial capital expenditures for the project are primarily a function of transmission costs. Additional expenses consist of heat exchanger costs, expenses associated with retrofitting existing residential heating systems, and design costs. The major transmission costs involve the purchase price and installation of the transmission system pipe. These costs, in 1981 dollars, are approximately $34/foot for 8-inch fiberglass-reinforced pipe and $9/foot for installation (Struhsacker, 1981). Approximately 5,300 feet of pipe is required for the project, at a cost of nearly $228,000.

The remaining front-end expenses consist of heat exchanger, circulation pump, retrofit, and design costs. A central heat exchanger and circulation of fresh water, rather than geothermal fluid, through the transmission system is assumed in order to reduce maintenance costs due to scaling and corrosion. The cost of a heat exchanger capable of handling 600 gpm of geothermal water is $22,000, installed, as taken from a report by Engen (1978) and inflated 10% annually to December, 1981 dollars. The cost of a pump needed to move hot water through the transmission system, including installation, is approximately $4,000. Retrofit costs are estimated at $1,000 per residential unit and design costs at 15% of the total system cost.

The annual operating expenses of the system consist primarily of maintenance and power costs. Unless specific maintenance items are known, annual maintenance costs are generally assumed to be a fraction of the initial capital cost (Engen, 1978). Annual maintenance expenses, assumed to be two percent of the total system cost, excluding design, are estimated at $5,880. Similarly, annual power costs are assumed to be 1% of the total system cost, excluding design, and are estimated at $2,940.

Discussion

The attractiveness of a district heating project at Jackson to potential investors depends primarily on the profitability of the venture in light of alternative investment opportunities and on the capital risk involved in the project's development and operation. The profitability, in turn, depends in large measure on the price that potential customers are willing to pay for heat from the project in light of alternative sources such as natural gas, heating oil, or wood.

The profitability of a venture is assessed in terms of the return on capital invested and is commonly measured as the venture's internal rate of return (IRR). The IRR of a project is defined as that rate of discounting future net revenue such that the initial cost of a project is equal to the sum of discounted future net benefits. It is commonly written as:

$$C_o = \sum_{t=1}^{n} \frac{R_t}{(1+i)^t}$$

where i is the internal rate of return, C_o is the initial cost of the project, R_t represents the net revenues from the project during period t, and n is the expected life of the project. The IRR is comparable to the yield to maturity on a bond or the rate of return on a savings account.

In addition to return on investment, investors are sensitive to investment risk which affects investment behavior in such a way that the greater the risk involved in a project, the higher the IRR must be in order to attract investors. On low-risk projects, the IRR need only be slightly greater than the yield available from relatively safe, low-liquidity investments such as U. S. Treasury bonds or bank certificates of deposit. The precise assessment of the risk associated with ventures similar to the district heating project at Jackson is a relatively complex process and beyond the scope of this study. However, some factors influencing the projects's risk can be briefly addressed qualitatively. Because the surface resource at Jardine Hot Spring is sufficient to supply the heating requirements of the project, risk due to resource uncertainty commonly associated with hydrothermal development projects is greatly reduced. In addition, equipment for the transmission system, retrofitting, and heat exchange are all "off the shelf" items, and development risks are consequently diminished. Assuming 100% market penetration, the investment risks are therefore assumed to be moderately low for the project.

The annual revenues from the project, and hence the IRR, are limited by the price potential customers are willing to pay for the heat. The average residential unit in Jackson, based on an informal survey, pays between $100 and $125 for heat on a monthly basis, in 1981 dollars. At these rates, with annual operating expenses estimated at $8,820, the net annual revenue from the project's 40 units would be between $39,180 and $51,180, yielding internal rates of return between 10.5% and 14.8%. These rates, coupled with public policy initiatives designed to accelerate geothermal development such as loan guarantees and tax credits, may make the project attractive to potential investors.

Economic Evaluation of Power Generation

In this section, the economics of power generation using the Jackson geothermal resource is addressed. Unlike many direct use applications whose economic value is primarily a function of savings over conventional fuel costs, the generation of electric power yields revenue directly through the sale of electricity. The economic viability of a power generation project of this type can be addressed by determining if the revenue from the project justifies the initial capital investment. To consider a specific application at Jackson, the parameters that must be estimated are the project's costs and net annual revenue. With these, the project's estimated internal rate of return can be determined.

The surface conditions at Jardine Hot Spring cannot support a viable power generation facility, as discussed earlier in this chapter. Resource quality at depth, therefore, is a critical factor affecting the generation capability of the project, which in turn affects initial costs, annual earnings, and return on investment. Although a geophysical survey can be conducted to identify specific target areas for drilling, an exploratory well is the only way to confirm suspected flow and temperature characteristics and estimate the economic life of the resource (Blair and others, 1982). Because of resource uncertainties, the risks associated with project development are estimated to be high.

Project Costs

The major factors affecting the initial project expenses are the generator size and drilling costs (Table 7), both of which are sensitive to resource characteristics at depth. These parameters are estimated for purposes of this analysis. Based on the geochemical and geological data discussed earlier in this study, a minimum resource temperature of 100°C and a flow rate of 32 liters per second (500 gpm) are expected to be encountered by drilling. Because discharge of thermal water is believed to occur along a fault (Chapter 6), drilling should be conducted so that the fault is intersected at a depth below that where significant heat is lost through mixing with shallow groundwater and/or by conduction to surrounding rock. Well depth, therefore, is roughly estimated to be between 1,000 and 2,500 feet.

Table 7. Power Generation Cost Summary

Initial Project Cost	2,500 Foot Well	1,000 Foot Well
Generator (installed)	$500,000	$500,000
Geothermal Well	100,000	40,000
Cold-Water Well	4,000	4,000
Down-Hole Pump	5,000	5,000
Rights Acquisition	4,000	4,000
Geophysical Survey	10,000	10,000
Total	$623,000	$563,000

* Annual pumping and maintenance costs are deducted from annual earnings figures.

Computer projections using the temperature and flow rate estimated above were performed in 1981 by KSPS/ORC, Inc. of Miami, Florida. Those projections indicate that the estimated subsurface resource at Jackson would drive a generation system capable of producing in excess of 400 kilowatts (KW) of electricity at the point of delivery to the transmission line. Total installed cost for this size generator was estimated by KSPS/ORC, Inc. at $1,250 per KW in December, 1981 dollars. Included in this estimate are all pumps and surface piping, all electrical interface equipment necessary to tie in with an existing transmission line near the hot spring, and the necessary concrete foundation. The generator output was determined after deducting expected drains on the generating system due to pumping-related power requirements, therefore pumping expenses other than purchase and installation costs are not included in the cost figures.

Drilling and completion expenses are estimated from a survey of well-drilling firms in the region. A price of $40 per foot is used for a well with ten inch casing. The casing size needed for a flow rate of 500 gallons per minute (gpm) was determined using Figure 14 (Engen, 1978). With well depths of 1,000 and 2,500 feet, costs are estimated at $40,000 and $100,000 respectively. In order to allow more precise definition of the discharge fault's location and geometry, and thereby select optimum drilling sites, the cost of detailed geophysical surveys are included in the project's expense estimate. These surveys, performed by personnel from the Montana Bureau of Mines and Geology (MBMG), include a gravity, shallow resistivity, and shallow seismic surveys. The cost, as determined by the MBMG, is $10,000.

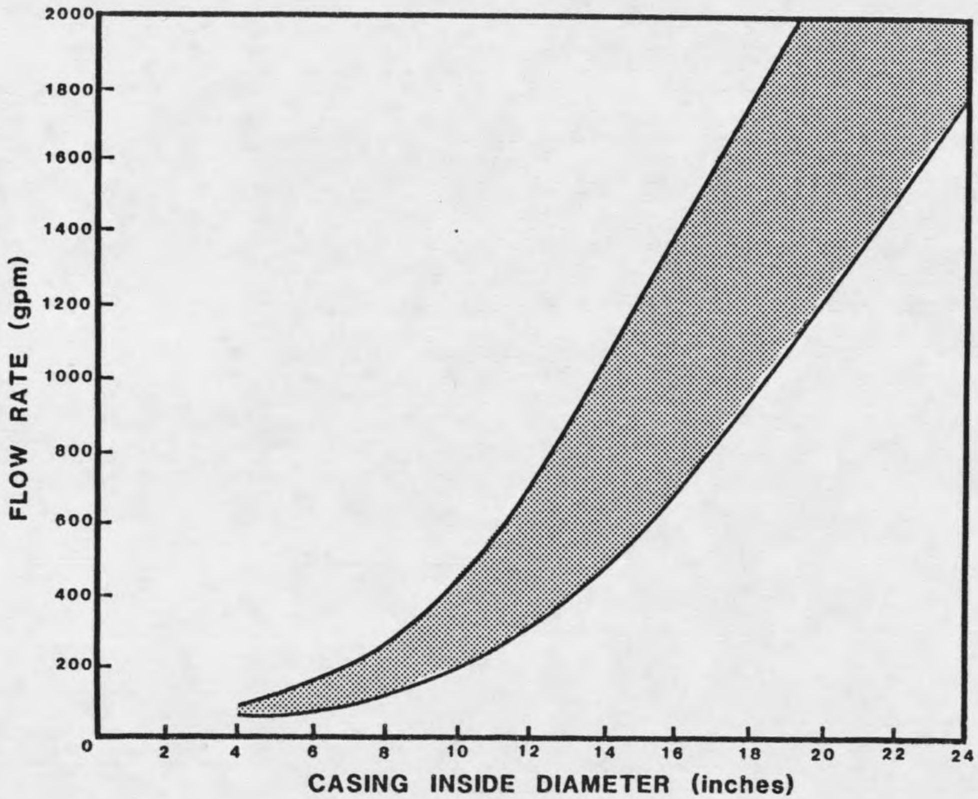

Figure 14. Production rate versus well casing diameter.
(After Engen, 1978)

Additional front-end costs include expenses for the acquisition of water and mineral rights. In Montana, geothermal energy has not yet been determined to be specifically a mineral right or a water right. The Montana Department of Natural Resources and Conservation (1980) recommends that both rights be secured by the developer. In the case of Jardine Hot Spring, these rights in the immediate vicinity of the hot spring are owned by the same individual, which lessens the complexity of rights acquisition. Pending the results of the geophysical survey, the mineral rights of an estimated 3,000 acres surrounding the hot spring should be acquired. Lease payments for mineral rights average one dollar per acre in the area and, with bonuses, are estimated to total $4,000.

Annual Earnings

Revenue from this project depends upon the price at which the power will be sold. A small power generating facility utilizing geothermal energy would qualify for consideration under Sections 201 and 210 of the Federal Public Utility Regulatory Policies Act of 1978, Public Law 95-617. This is an important consideration because projects which qualify under this act are assured the sale of produced power. A utility must purchase power from qualifying projects at a rate not less than the avoided costs the utility would have incurred if the energy was not obtained from the project. In practice, the purchase price would be close to the utility's marginal cost for electricity rather than the average cost which the utility charges its customers.

The power from the project at Jackson would be sold to the Vigilante Electric Cooperative of Dillon, Montana. The Cooperative operates a 7,200 volt transmission line that passes within 300 meters (1,000 feet) of Jardine Hot Spring and is capable of handling the electricity produced from that project. The Cooperative is part of the Bonneville Power Administration (BPA) system. It is not clear at present to what extent BPA is subject to the avoided cost requirements of Public Law 95-617. BPA, therefore, has not specified an avoided cost rate. From a survey of the other utilities in the area, a rate of \$0.50/kilowatt hour was estimated. This figure is used, along with the amount of electricity delivered to the transmission line, to determine estimated annual revenue. Annual expenses for maintenance and pumping costs are deducted from the revenue generated through the sale of power to arrive at the expected net annual earnings, which are estimated to be \$167,000.

Discussion

In geothermal power generation projects, the producer is principally concerned with maximizing the profitability of delivering power from the resource, while the utility is concerned with minimizing the cost of producing reliable baseload power. The link between these two concerns is the price of delivered power, which is established by the least expensive available alternative type of new baseload generation at the point of delivery to a transmission system (Blair and others, 1982). With the price established, net annual revenues are determined based on generating capacity and annual costs as seen above.

Coupled with an estimation of the project's initial costs, the profitablility of the venture can be assessed in terms of the project's internal rate of return. Prior to discussing the venture's IRR, factors influencing risk, which in turn affects investment behavior, will be addressed.

The chief elements of investment risk affecting the project are related to the uncertainties of resource quality and drilling depth. A precise treatment of risk associated with these parameters requires that a probability distribution analysis be performed, which is beyond the scope of this study. This procedure, as applied to geothermal development projects, is discussed in detail by Blair and others (1982). For purposes of this study, resource quality is assumed to approximate the minimum estimates used in this analysis (100°C and 32 liters per minute) and uncertainty with regard to drilling depth is dealt with in terms of increases in initial project cost estimates and their effect on project internal rates of return.

The probability of encountering the temperature and flow rate used in this analysis is assumed to increase with depth. In other words, hotter water is more likely to be found at deeper, rather than at shallower, depths. With a 1,000 foot well assumed, the project's initial costs are estimated to be $563,000 (Table 7) and the IRR is 29.6%. Assuming a 2,500 foot well increases the estimated initial costs to $623,000 and decrease the IRR to 26.7%. If estimated drilling depth is increased to 5,000 feet, the IRR decreases to 22.8%. These rates of return on capital invested in the power generation project are all significantly higher than those estimated for the district heating project and are commensurate with the higher risks involved. The relatively high risk may be partially offset by the increased number of tax advantages and other public sector incentives offered to geothermal projects of this type. Such incentives may make the project very attractive to potential investors.

CHAPTER 8

SUMMARY

Hydrothermal circulation near Jackson, Montana is apparently fault controlled and occurs within a regime of regional, rather than magmatic, heat flow. Recharge probably takes place on the eastern margin of the Big Hole Basin with meteoric waters descending the plane of a northwest trending thrust fault. This fault places Precambrian Missoula Group rocks onto a sequence of Cambrian clastic and dolomitic sedimentary rocks. The permeability of the fault plane was enhanced by post-thrusting extensional motion. Discharge probably occurs along an inferred north-trending normal fault connecting Jardine Hot Spring with a small group of warm springs two kilometers to the northwest.

Chemical geothermometers yield a range of estimated minimum reservoir temperatures. Specific temperature estimates depend upon certain assumptions concerning reservoir chemical conditions as well as heat loss during ascent of thermal water. A best estimate of reservoir temperature is a range between 100°C and 145°C, assuming that silica concentration is controlled by quartz solubility. If heat loss occurs primarily by conduction, a reservoir temperature of at least 100°C is likely. If thermal water mixes with cold water during ascent, reservoir temperatures near 145°C are calculated. Although less likely, reservoir temperatures as low as 70°C are possible assuming chalcedony solubility controls silica concentration. Control by quartz, however, is more consistent with the geologic setting of the thermal system.

The measured discharge temperature, coupled with a high flow rate, makes Jardine Hot Spring an attractive target for development. The proximity of the hot spring to the town of Jackson favors the direct utilization of the resource. The existing surface resource could support the heating requirements of a district space heating project and preliminary analysis indicates that such a venture may be profitable. The estimated subsurface resource may be attractive to potential developers as well. Expected temperatures of approximately 100°C and flow rate of 32 liters per second (500 gpm) could support a 400 kw power generation project, which could yield moderately high returns on investment.

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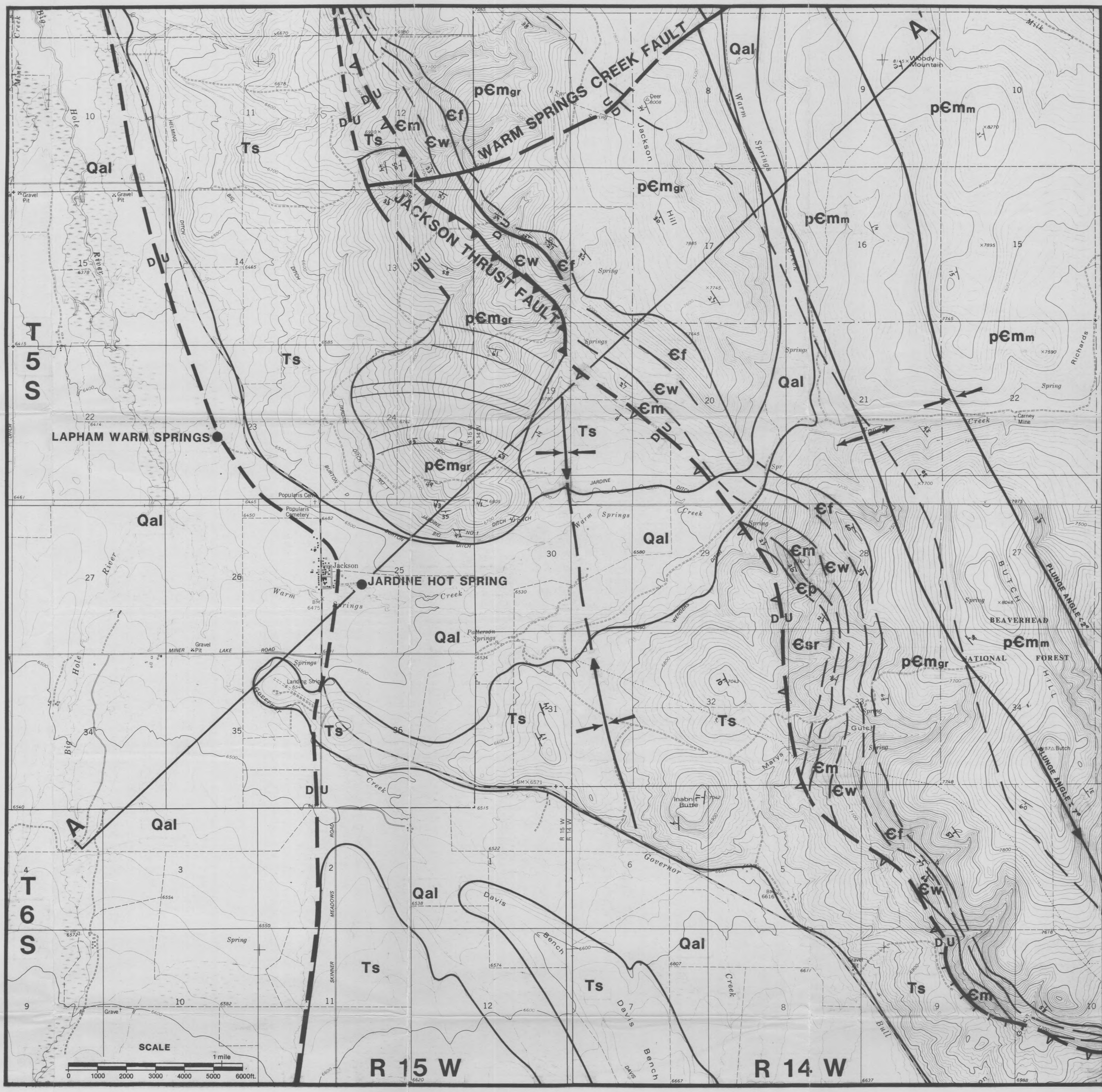


LEGEND

Qal	QUATERNARY ALLUVIUM	Єf	CAMBRIAN FLATHEAD SANDSTONE
Ts	TERTIARY SEDIMENTS	pЄmgr	PRECAMBRIAN MISSOULA GROUP GARNET RANGE FORMATION
Єsr	CAMBRIAN SNOWY RANGE FORMATION	pЄmm	PRECAMBRIAN MISSOULA GROUP McNAMARA FORMATION
Єp	CAMBRIAN PILGRIM DOLOMITE		
Єm	CAMBRIAN MEAGHER DOLOMITE		
Єw	CAMBRIAN WOLSEY SHALE		

- THRUST FAULT
- NORMAL FAULT (reactivated thrust fault)
- NORMAL FAULT
- FAULT SCARP
- FOLD, ANTICLINE
- FOLD, SYNCLINE
- FORMATION CONTACT
- BED LINES
- STRIKE AND DIP OF BED

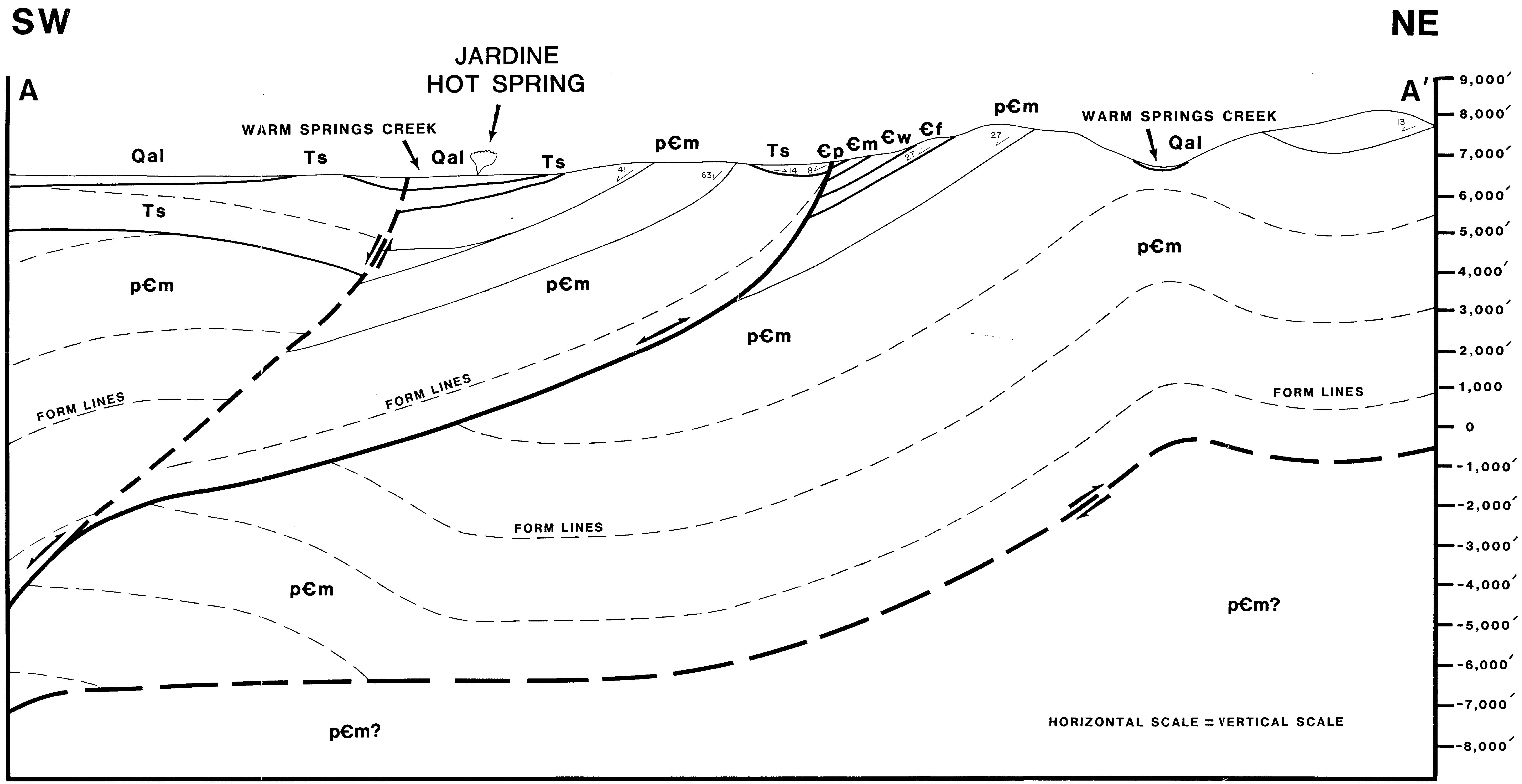
N

GEOLOGIC MAP OF JACKSON, MONTANA AREA

PLATE 1

BASE FROM U.S. GEOLOGICAL SURVEY MAPS OF JACKSON, FOX GULCH, JACKSON HILL, AND BUTCH HILL QUADRANGLES, MONTANA

G.B. '83



GEOLOGIC CROSS SECTION : JACKSON, MONTANA AREA

PLATE 2