D.R. MABEY AND K.E. BUDDING, HIGH-TEMPERATURE GEOTHERMAL RESOURCES OF UTAH

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HIGH-TEMPERATURE GEOTHERMAL RESOURCES OF UTAH

By Don R. Mabey and Karin E. Budding

UTAH GEOLOGICAL AND MINERAL SURVEY

a division of

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UTAH GEOLOGICAL AND MINERAL SURVEY

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HIGH-TEMPERATURE GEOTHERMAL RESOURCES OF UTAH

By Don R. Mabey[1] and Karin E. Budding[2]

ABSTRACT

Numerous thermal springs and shallow wells in western Utah are near-surface evidence of geothermal systems containing heat energy that can be produced and put to beneficial use. The seven known geothermal systems in Utah that contain water or steam at temperatures greater than 90°C are in an area of southwestern Utah here defined as the Sevier thermal area. Hot water and steam produced from two of these systems, Roosevelt and Cove Fort, are being used to generate electricity. Two of the systems, Joseph and Monroe-Red Hill, apparently contain only moderate-temperature water that may be useful for space heating and industrial processes but not useful in the foreseeable future for the generation of electricity. The other three systems of Fumarole Butte, Thermo, and Newcastle require additional exploration before their potential can be evaluated. The geological and geophysical data in Pavant Valley indicate a promising area for the occurrence of a high-temperature geothermal resource where no hot water has been discovered. An evaluation of exploration data obtained for the known geothermal systems provides an indication of what geological, geopohysical and geochemical techniques are most useful in Utah for exploring known geothermal systems and in the search for undiscovered systems.

INTRODUCTION

Hot springs, fairly common in western and northwesterrn Utah, are surface evidence of the geothermal energy stored within the earth. Igneous rocks, which formed when molten rock cooled, are another. The term geothermal is used to describe this naturally occurring heat within the earth. Geothermal systems, as used in this report, are local concentrations of heat in the upper part of the earth's crust and the geologic elements that produce the concentration, while the term geothermal resource refers to geothermal energy that might be extracted and used.

In the 1960s, as the United States approached the limit of its ability to produce oil and gas, interest in geothermal heat as a source of energy increased and both industry and government began to examine geothermal resources in Utah and other western states. Two publications by the U.S. Geological Survey, a Water-supply Paper by Stearns and others (1937) and a Professional Paper by Waring (1965), included data on over 50 Utah hot springs. Several general studies since 1965 have been concerned with geothermal resources in all or major parts of Utah. These include "Geothermal Power Potential in Utah" (Heylmun, 1966), "Major Thermal Springs of Utah" (Mundorff, 1970), "Geothermal Energy and Water Resources in Utah" (Batty and others, 1975), "Thermal Waters of Utah" (Goode, 1978), "Geothermal Resources of Utah" (Murphy, 1980), and "Reconnaissance of the Hydrothermal Resources of Utah" (Rush, 1983). Estimates of the energy resource in Utah's geothermal systems are also included as part of National assessments of geothermal resources in "Assessment of Geothermal Resources of the United States - 1975" (White and Williams, 1975), "Assessment of Geothermal Resources of the United States - 1978" edited by Muffler (see Muffler and Guffanti, 1979) and "Assessment of Low-tem-perature Geothermal Resources of the United States - 1982" (Reed, 1983).

In addition to these publications concerned with regional aspects of the geothermal resource, numerous studies have concentrated on individual geothermal systems in Utah. Most of these studies were supported by either the Department of Energy (DOE) and its predecessors, the Energy Resources Development Agency and the Atomic Energy Commission, the U.S. Geological Survey (USGS), or the National Science Foundation. The Earth Science Laboratory at the University of Utah Research Institute (ESL/UURI) and the University of Utah, primarily with support from DOE, have made extensive investigations of some of the geothermal systems. The USGS has studied most of the geothermal systems and the Utah Geological and Mineral Survey (UGMS) has investigated several.

This study is concerned primarily with those geothermal systems that appear capable of producing steam or water at temperatures of 90°C or higher. We classify these as "high-temperature" systems. The USGS in the "Assessment of Geothermal Resources of the United States - 1978" (Muffler and Guffanti, 1979) classified hydrothermal systems as (1) high-temperature—greater than 150°C, (2) intermediate-temperature—between 90° - 150°C, and (3) low-temperature—below 90°C. Thus we include in our high-temperature classification both the high- and intermediate-temperature systems of Muffler and Guffanti. Reed (1983) included systems with temperatures up to 100°C in the low-temperature resource. Temperatures reported in this report are given in degrees centigrade (°C) and may be converted to degrees Fahrenheit (°F) with the equation °F=1.8(°C)+32°.

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Although high-temperature hydrothermal resources have several possible applications, the primary use to date has been the generation of electricity. The only commercial development of geothermal energy in the United States has been through the production of steam or hot water. Considerable research has been directed toward developing techniques to extract energy from hot dry rock and from molten rock. The possible distribution of these geothermal resources in Utah will be discussed briefly.

The most recent National assessment of high-temperature geothermal resources completed by the U.S. Geological Survey (Brook and others, 1979) concluded that the geothermal systems in Utah with reservoir temperatures greater than 150°C (Roosevelt and Cove Fort) contained about six percent of the energy in geothermal systems in this temperature range in the United States outside of National Parks (table 1). In energy content these two Utah systems were ranked seventh and eighteenth among the 52 such systems in the United States. The Roosevelt and Cove Fort systems are now being used for the commercial generation of electricity. In addition, five relatively small geothermal systems in Utah with reservoir temperatures between 90°C to 150°C were identified. Since this assessment was completed considerable research and exploration relating to geothermal resources in Utah has been done but no additional systems with reservoir temperatures 90°C or greater have been reported. In the study reported here the Utah Geological and Mineral Survey, with a grant from the Department of Energy, has examined data relating to the high-temperature (greater than 90°C) systems to: (1) update the assessment of the resource in the known systems and update the total known high-temperature resource of Utah, (2) study the known high-temperature systems to identify characteristics that are useful in determining factors controlling the development of the systems and in designing exploration strategies, and (3) identify areas that appear favorable for the occurrence of undiscovered geothermal systems.

Previous major studies of regional geothermal resources have emphasized hydrology. Mundorff (1970) and Goode (1978) are compilations of information on thermal waters with some information on the local geology. Rush (1983) investigated several thermal areas in considerable detail using hydrologic, geologic, geochemical, and geophysical data; however, he did not concentrate on the regional relationships. The ESL/UURI group has presented case studies of three thermal areas, and several studies have reported on one or more data sets relating to individual areas. This report differs from earlier work in that it emphasizes regional relationships. In discussing individual thermal areas we do not repeat all of the information previously reported but synthesize the earlier work as it relates to understanding the resource and to developing strategies for exploration of similar resources.

GEOTHERMAL SYSTEMS

Vast amounts of thermal energy are stored in the earth; however, normally this energy can be put to large-scale beneficial use only when it is concentrated by natural processes near the surface. Several geologic processes produce such concentrations. When molten rock rises toward the surface of the earth it transfers a great amount of heat from deep hot zones within the earth. If this molten rock is erupted on the surface of the earth, most of this heat is lost into the atmosphere as the rock cools and solidifies. If a large mass of this molten rock

TABLE 1. KNOWN HIGH-TEMPERATURE GEOTHERMAL SYSTEMS IN UTAH

Name of System	Estimated Reservoir Temperature (°C)	Estimated Reservoir Volume (km^3)	Estimated Reservoir Energy (10^{18}J)
Roosevelt			
Roosevelt Hot Springs (McKeans)	265	47	32
Cove Fort			
Cove Fort-Sulphurdale	167	39	16
Thermo			
Thermo Hot Springs	142	8.3	2.8
Newcastle			
Newcastle area	130	6.1	1.9
Joseph			
Joseph Hot Springs	107	3.3	0.8
Monroe-Red Hill			
Monroe-Red Hill Hot Springs	101	4.7	1.1
Fumarole Butte			
Abraham (Baker, Crater) Hot Springs	97	6.1	1.4

Data and names in parentheses are from Brook and others (1979). Energy is in 10^{18}J (joules) which is approximately equal to 10^{15} British thermal units (BTU) and one quad (a quadrillion BTU).

remains below the surface of the earth, it cools slowly by conductive and convective transfer of heat to the enclosing rock. The rate at which cooling occurs is dependent upon the volume and temperature of molten rock, thermal conductivity, and water circulation in the enclosing rock. Large intrusive masses of molten rock may persist as important thermal anomalies for several million years; such thermal anomalies are not common in Utah but they may be involved in some geothermal systems. A more common method of mass transfer of heat within the shallow crust of the earth is by circulation of ground water. Downward circulating water is heated and then rises transferring heat toward the surface.

Hydrothermal Convection Systems

The geothermal resources of greatest interest in Utah are in hydrothermal convection systems. Energy can be extracted from hot water or steam produced from these systems. Hydrothermal convection systems are often identified because water from the system is discharging at the surface in a hot spring, or hot water is encountered in drilling a well. Thermal water in springs and shallow wells is usually part of a hydrothermal convection system that may contain larger volumes of water at depth, with a higher temperature than the near-surface water. Hot water discharged from springs can be used in applications where small volumes of relatively low-temperature water are required such as for small-scale space heating. But in applications that require either large volumes or high-temperature water, for example the generation of electricity, the deeper and hotter parts of the system must be tapped.

The major elements of a hydrothermal convection system are: (1) conductive heat flow supplying heat to the system; (2) cold water descending through a permeable zone, such as fractured rock, to a zone of high temperatures where the water is heated by direct contact with hot rock; (3) less dense hot water ascending toward the surface in a permeable zone; (4) a reservoir or porous rocks where hot water or steam is stored and where reservoir rock is heated; and (5) hot water discharging into shallow aquifers or, in some systems, at the surface in springs. Geothermal systems are continually transporting heat toward the surface, but the primary resource is the heat that is stored in the water and rock of the reservoir. A large-volume reservoir is an essential element if a geothermal system is to support a sustained high level of production. In some small systems the reservoir may be very limited, perhaps only to a fault zone. Normally the resource is developed by drilling wells into the reservoir and extracting the hot water or steam. Some heat can be supplied to the reservoir by convection in the system during the period of production, but the rate of resupply will usually be small and the reservoir will be depleted. Although the reservoir may eventually recover as heat and hot water recharge the reservoir, the time required normally will be too great to be of interest in a resource study. Therefore, the resource of primary interest is the recoverable energy in the reservoir.

Hot Dry Rock and Magma Systems

Thermal energy is also present in rocks that do not contain large amounts of extractable water because of low permeability. Heat has been extracted experimentally from these "hot dry rock" reservoirs by fracturing the rock and circulating water through the fractures. These experiments may lead to the development of a technology for the commercial extraction of heat from these reservoirs. If this occurs, the potential for the development of geothermal energy in Utah, as in many other areas, will be greatly expanded. However, such development does not appear likely in the near future.

Research is also being conducted in an attempt to develop techniques for extracting heat from molten or partially molten rock. No near-surface masses of molten rock are known to exist in Utah, but such masses may exist at depths greater than several kilometers. No attempt was made in this study to appraise the energy in either hot dry rock or magma reservoirs in Utah.

DISTRIBUTION OF GEOTHERMAL SYSTEMS

Most hot springs and all known major geothermal systems in Utah occur in the western part of the state and this study is focused on southwestern Utah where high-temperature systems are known to occur (figure 1). However, additional high-temperature systems may exist in other parts of Utah and exploration in these areas should be encouraged. Four geothermal systems with calculated reservoir temperatures from 99°C to 149°C occur in southern Idaho within 20 km of the Utah border. Water from one of these, the Raft River system north of the Raft River Mountains, has been used to generate electricity in a pilot plant built by the Department of Energy. The other systems are in or near the northern Cache Valley in Idaho and similar ones may exist in northern Utah.

A petroleum exploration well west of the Promontory Mountains is reported to have measured a bottom-hole temperature of 214°C at 3,660 m below the surface (Murphy, 1980). Further east, a series of warm springs occur along the west side of the Wasatch Range and within the valleys; most are associated with spurs of the Wasatch Range. In the area of the spurs the Wasatch fault dips more steeply than elsewhere and this anomalous geometry appears to favor the development of geothermal systems in which deep circulating ground water is heated to the temperatures measured in the springs. Although no high-temperature systems in northwest Utah have been reported, the possibility that high-temperature systems exist should not be discounted. Eastern and southeastern Utah appear less favorable for the occurrence of high-temperature geothermal systems.

THE SEVIER THERMAL AREA

The seven known high-temperature geothermal systems in Utah occur in or along the margins of intermontane valleys in the eastern part of the Basin and Range geologic province. Six are within 70 km of the eastern province boundary. The Sevier thermal area (figure 1) has been defined for study in this investigation. Several regional geological and geophysical features may relate to the concentration of geothermal systems in the Sevier thermal area. The Intermountain seismic belt extends across the eastern and southern part of the area. Six of the geothermal systems are in the zone where the seismic belt changes from the southerly trend that characterizes the belt in northern Utah to a more westerly trend in the south. Throughout the thermal area north-trending late Cenozoic normal faults reflecting east-west extension are common. A broad oval-shaped feature called the Sevier oval has been defined by T. A. Steven and others (written communication, 1986) on the basis of geology, topography, and geophysical anomalies and is approximately coincident with the northern part of the Sevier thermal area. The oval is a major anomaly in the crust and perhaps the upper mantle. A major zone of igneous rocks of mostly Oligocene age extends across the southern part of the Sevier thermal area. Both intrusive and extrusive rocks are abundant at the surface, and magnetic data suggest that throughout much of the zone intrusive rock is very abundant in the subsurface. In the area of abundant igneous rock, the post-Oligocene deformation has been influenced by the presence of the igneous rocks. Six of the seven geothermal systems are in this igneous zone. Another zone of igneous rocks trending north through the Sevier thermal area has some of the youngest extrusive rocks in Utah, and the north end contains the Fumarole Butte geothermal system in the area of an eruptive center.

The occurrence of geothermal systems in western Utah relates to two geologic conditions that do not exist elsewhere in the state—high regional heat flow and complex structure that allows the development of hydrothermal convection systems. In addition, young igneous systems may be supplying heat to some geothermal systems. An understanding of these three factors is the key to understanding the geothermal systems and assessing the geothermal resources of this region.

REGIONAL GEOLOGIC CONTROLS

Regional Geology

The Sevier thermal area lies partly within the Basin and Range Province and partly in the transition zone. The transition zone lies between the Basin and Range and Colorado Plateau physiographic provinces and involves a mixture of surface and subsurface features of both provinces (figure 1). The western part of the thermal area is dominated by extensive basin areas, part of the Sevier and Escalante Deserts (figure 2). These basins have a cover of Quaternary and Tertiary sediments except in local areas where Cenozoic volcanic rocks are exposed. The ranges adjacent to these basins are composed primarily of sedimentary and metasedimentary rocks of Precambrian through Mesozoic age, Oligocene extrusive rocks, and Oligocene and Miocene intrusive rocks. The rocks exposed in the ranges are presumed to also underlie basin areas. The dominant pre-Tertiary structures in the thermal area are overthrusts and related folds of the Sevier orogenic belt of Cretaceous age. Extensive volcanism in an east-trending belt in the southern half of the area in Oligocene time was followed by general east-west extension in Neogene time with the resulting development of basin and range structures.

The transition zone between the Basin and Range and Colorado Plateau Provinces approximately coincides with the eastern extent of overthrusting associated with the Sevier orogeny. Synorogenic sediments deposited east of the Sevier orogen occur in the transition zone. Salt structures involv-

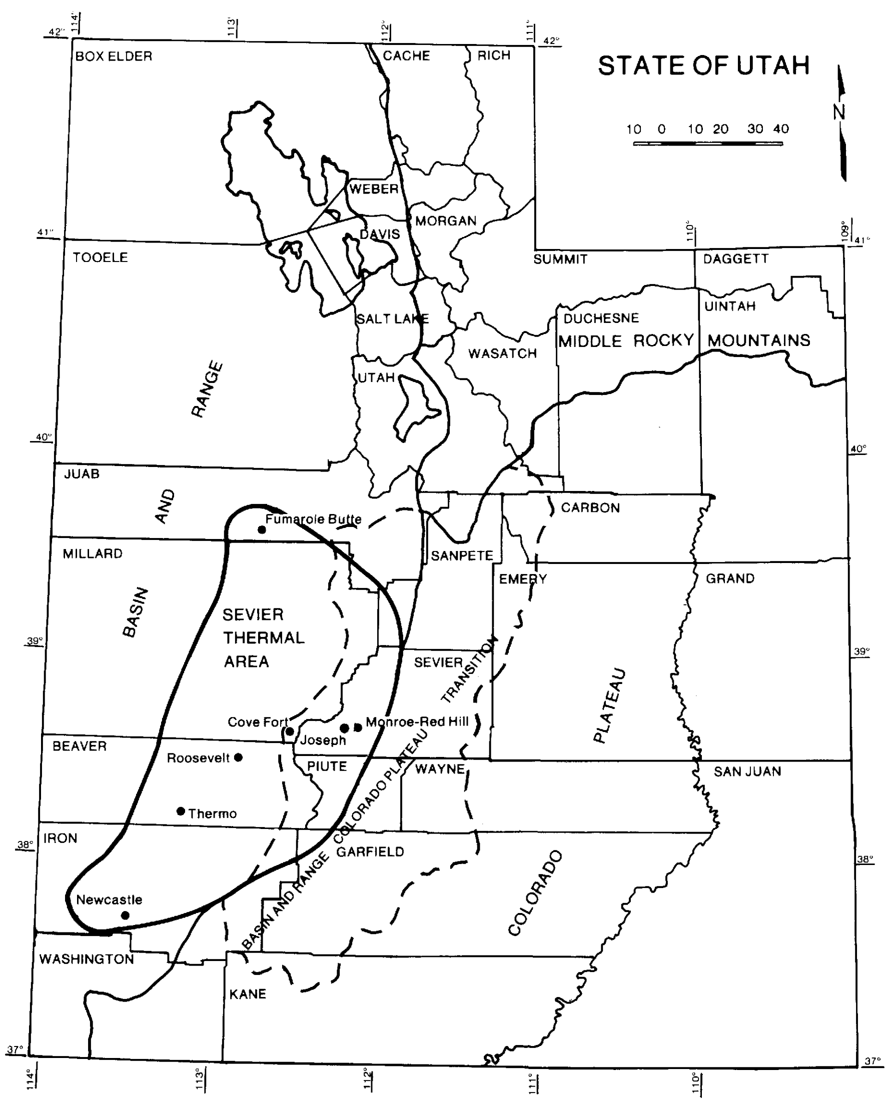


Figure 1. Map of Utah showing the known high-temperature geothermal systems, the Sevier thermal area, and physiographic provinces.

ing Jurassic and younger sediments are also present. Oligocene volcanism extended across the transition zone.

The oldest rocks in the Sevier thermal area are banded amphibolite gneiss, sillimanite schist, and quartzite in isolated outcrops on the west side of the Mineral Mountains. These rocks were formed by regional metamorphism about 1.7 Ga (billion years ago) (Aleinikoff and others, in press; Nielson and others, 1978). The rocks are important reservoir rocks in the Roosevelt geothermal system.

Sedimentation began in latest Precambrian time and continued with minor interruptions until middle Mesozoic time (figure 2). Precambrian and Paleozoic rocks exposed in the region have an aggregate thickness of about 10,000 m. Exposed Precambrian sedimentary rocks about 2,000 m thick are weakly metamorphosed sandy, shaley, and limey strata. The Paleozoic units are about 75 percent limestone and dolomite with lesser amounts of quartzite, sandstone, and shale. All of the exposed Precambrian and Paleozoic sedimentary rocks in and immediately to the west of the Sevier thermal area may be in thrust plates of the Sevier orogenic belt and are perhaps underlain by younger rocks.

Triassic and Jurassic non-marine sedimentary rocks are exposed in thrust windows at Blue Mountain, in the Canyon Mountains, and in the Pavant Range. The aggregate thickness of these rocks is about 1,500 m, about one-third of which is Navajo Sandstone, and they may underlie older rocks in much of the area.

In the Cretaceous the Sevier orogeny produced regional uplift, eastward overthrusting, and folding. Debris shed eastward from the rising area formed Upper Cretaceous conglomerates, sandstones, siltstones, and shales aggregating up to 2,000 m thick in the eastern part of the thermal area. In early Tertiary time, non-marine sediments continued to accumulate in local basins in this area. Locally these units are also about 2,000 m thick. The overthrusting that occurred along several major thrust faults in the study area generally juxtaposed older rocks eastward over younger rocks. The individual thrust plates were coherent over considerable distances. Total lateral displacement of the major thrusts appears to be about 100 km.

During early Oligocene time volcanic activity began in the southern half of the Sevier thermal area and has continued intermittently to the present (figure 3). Andesitic to rhyodacitic stratovolcanos developed before 30 Ma (million years ago) across the thermal area, in the southern Wah Wah Mountains and the southern San Francisco Mountains-Shauntie Hills area to the west, and in the northern Tushar Mountains to the east. By 30-28 Ma, middle Oligocene time, large-scale pyroclastic eruptions occurred to the west of the study area in the Needle Range. Resulting ash flows spread as far east as the Sevier Plateau in the eastern portion of the thermal area (Steven and Morris, 1984).

Volcanism continued in late Oligocene and earliest Miocene time with andesitic to rhyodacitic volcanos forming the thick volcanic pile of the Marysvale volcanic field in the Tushar Mountains. Some of the pyroclastic eruptions between 27 and 23 Ma resulted in source area collapse forming the Three Creeks, Big John, and Monroe Peak calderas (figure 3) (Steven and others, 1984). Lower volume andesitic volcanism continued locally to the west in the Shauntie Hills and southern Wah Wah Mountains.

Paralleling the zone of volcanic activity is a belt of intrusive rocks that become younger to the east (figure 4). The oldest intrusive activity is dated at 29 Ma in the southern San Francisco Mountains-Beaver Lake Mountains, Rocky Range, and Star Range in the west (Lemmon and others, 1973); followed by 25 Ma intrusions in the Mineral Mountains (Aleinikoff and others, in press); intrusive activity occurred between 27 and 22 Ma in the northern and central Tushar Mountains (Steven and Morris, 1983a; Steven and others, 1979); and the youngest intrusives of 22 to 21 Ma are in the northern Sevier Plateau on the eastern side of the thermal area (Steven and others, 1984). The calderas of the Marysvale volcanic field (figure 3) lie within this belt of Tertiary intrusive rocks, attesting to their common source at depth.

In early Miocene, the composition of the volcanic activity changed from a calc-alkalic to a bimodal basalt-rhyolite assemblage roughly coinciding with the onset of basin and range extensional tectonism in the area. Potassium-rich mafic lavas and silicic ash-flow tuff were first erupted about 23 Ma in the western part of the thermal area in the southern Wah Wah Mountains and in the Shauntie Hills. Bimodal activity began about 22 Ma in the Marysvale volcanic field with the eruption of mafic lavas. Eruption of silicic lavas and pyroclastic rocks began about 21 Ma and continued intermittently until about 14 Ma. Ash-flow tuff eruptions resulted in the Mount Belknap and the Red Hill calderas (Steven and others, 1984) in the Marysvale volcanic field about 19 Ma (figure 3). The source areas of the older calc-alkalic volcanics and the younger silicic end member of the bimodal suite roughly overlie each other.

Silicic volcanic activity continued locally in late Miocene time between 10 and 5 Ma producing small rhyolitic flows and domes just west of the Mineral Mountains, at the northern end of Beaver Basin, in the northern Black Mountains, and in the Sevier Plateau (Steven and Morris, 1984).

Bimodal volcanism continued in Pliocene and Pleistocene time producing a north-trending belt from the central Beaver Basin north to Fumarole Butte. Silicic Pliocene volcanism began at Smelter Knolls, a rhyolite flow-dome south of Fumarole Butte, about 3.4 Ma (Turley and others, 1979), and continued with additional rhyolite flows and domes in the Coyote Hills-Twin Peaks area from 2.7 to 2.4 Ma. Basalt was erupted contemporaneously about 2.4 Ma to the south and east of the Coyote Hills area (Crecraft and others, 1981; Nash, 1981).

The majority of the Pleistocene volcanic activity in the thermal area was mafic in composition and erupted through the area of Pliocene volcanic rocks along this north-trending belt (figure 3). Some basalt also erupted at an isolated spot in the southern end of the study area. Basalt erupted about 1 Ma near Black Rock (Crecraft and others, 1981) and at Fumarole Butte (Galyardt and Rush, 1981). Basaltic activity continued in the Cove Fort and Kanosh areas between about 0.7 to 0.5 Ma (Luedke and Smith, 1978; Steven and Morris, 1983a). Younger volcanic activity to the north near Fillmore began about 0.22 Ma and continued through Lake Bonneville time

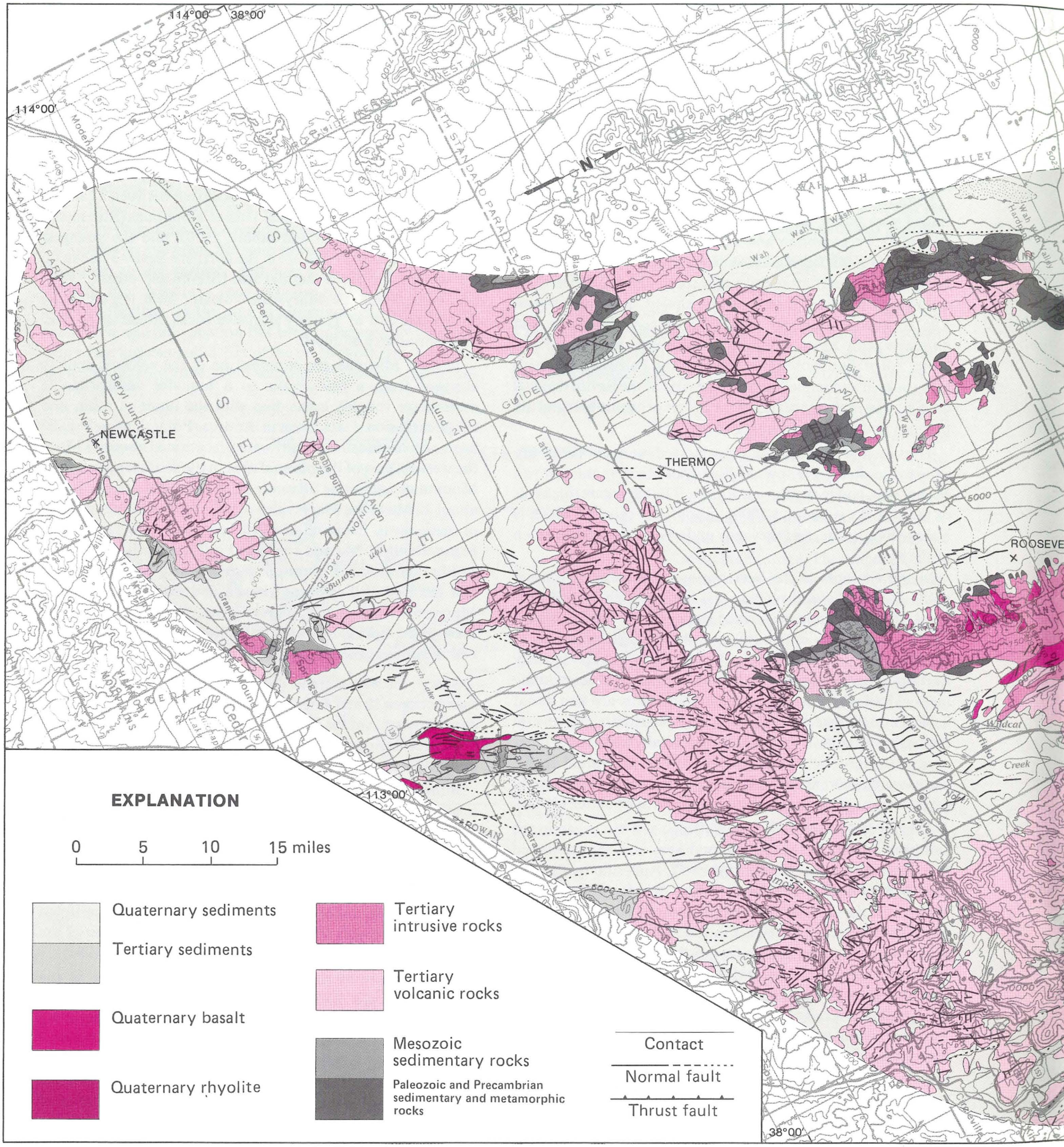


Figure 2. Generalized geologic map of the Sevier thermal area, modified from Hintze (1980).


39°00'
113°00'
ABRAHAM
SEVIER
LAKE
COVE FORT
JOSEPH
MONROE-RED HILL
112°00'
112°00'
39°00'
Delta
Sutherland
Abraham
Lynndyl
Holden
Flowell
Greenwood
Meadow
Hatton
Jacobs Well
Clear Lake
Beaver
River
UNION PACIFIC
Oasis
Hinckley
Deseret
Sugarville
Woodrow
Leamington
Sevier

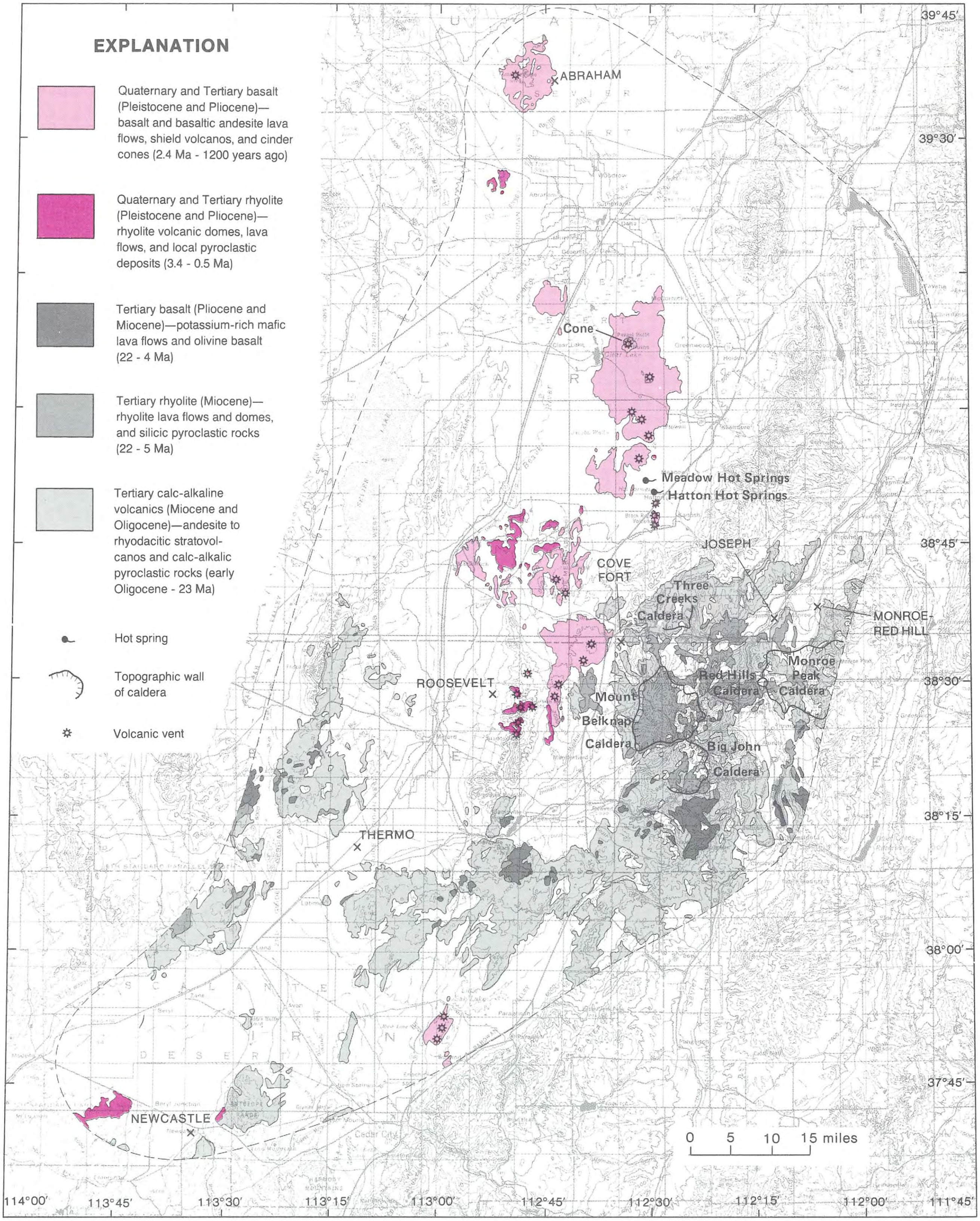


Figure 3. Geologic map of the Sevier thermal area showing the Quaternary and Tertiary volcanic rocks from Galyardt (unpub. map), Hintze (1980), Morris (1978), Schubat and Siders (mapping in progress), Siders (1985a and 1985b), and Steven and Morris (1983b); vents and cinder cones for Quaternary and late Tertiary basalt and rhyolite from Cunningham and others (1983), Hintze (1980), Hoover (1974), Machette and others (1984), and Steven and Morris (1983a).

until the last eruption about 700 years ago (Condie and Barsky, 1972; Luedke and Smith, 1978; Valestro and others, 1972). Pleistocene silicic volcanic rocks in the thermal area consist of a 0.4 Ma rhyolitic flow in Pavant Valley northwest of Meadow and a sequence of 0.5 to 0.8 Ma rhyolite flows, domes, and pyroclastic deposits along the crest of the Mineral Mountains (Lipman and others, 1978).

Extensive basins formed in late Miocene time associated with the development of the Basin and Range Province. Over 1,000 m of sediments accumulated locally in these basins. As normal faulting continued through the late Cenozoic, these basins appear to have evolved into generally smaller, more complex local basins and valleys; however, regional subsidence along the eastern edge of the developing Great Basin produced a major closed drainage basin. About 15,000 years ago Lake Bonneville covered much of northwest Utah including half of the Sevier thermal area. Maximum depth of water in the thermal area was about 200 m. Abundant evidence of older lakes exists but the age, extent, and depths have not been determined. Today the entire thermal area is part of the Sevier Lake drainage basin, a major subbasin of Lake Bonneville basin.

Heat Flow

Average regional heat flow in the Basin and Range Physiographic Province in Utah is high; significantly above the Colorado Plateau, which is near the continental average at 50 mW/m^2 (milliWatts per square meter) (Chapman and others, 1978). Conductive heat flow values have been computed from temperature-gradient measurements in and adjacent to the Sevier thermal area and these data do not define a heat flow anomaly coincident with the area as defined here (figure 5). However, the distribution of heat flow determinations does not establish that an anomaly does not exist. The geothermal map of Utah (Murphy, 1980) shows about 60 heat flow determinations unevenly distributed in and within 25 km of the Sevier thermal area. If values of more than 120 mW/m^2, which likely reflect nonconductive heat transfer, are excluded, the average for 32 determinations is 88 mW/m^2. Undoubtedly some of these 32 values reflect the effects, both positive and negative, of hydrothermal convection systems. The generalized heat flow map of the United States prepared by Sass and others (1976) shows the entire thermal the Sevier thermal anomaly above 1.5 HFU (heat flow units) (63 mW/m^2) with a large area of greater than 2 HFU (84 mW/m^2) in the southeast part of the Sevier thermal area. The greater than 2 HFU anomaly may be more extensive and cover much of the Sevier thermal area. Although existing data are permissive of a heat flow high in the general area of the Sevier thermal area, more measurements will be required to determine if such an anomaly is present.

If regional heat flow over the Sevier thermal area is 80 mW/m^2 and the average thermal conductivity is assumed to range from about 1.25 W/m/K (watts per meter per degree Kelvin) for Cenozoic sediments to about 3 W/m/K for pre-Mesozoic rocks, the normal thermal gradient will be between about 27° and 64°C/km (degrees Celsius per kilometer). Thus for water at an initial temperature of 10°C to be heated to temperatures of 100°C in the normal regional thermal gradient, it would need to circulate to depths of between 1,400 to 3,300 m. Because unconsolidated or poorly consolidated sediments have a lower thermal conductivity than consolidated rocks, thick accumulations of these sediments act as a thermal blanket and consequently raise the temperatures under the Cenozoic basins and refract heat flux into adjoining consolidated rocks. A Cenozoic basin filled with sediment: (1) increases the temperature under the basin relative to comparable depths in adjacent areas, (2) decreases the heat flow in this central part of basin, and (3) increases the heat flow along the margins of the basin.

Thermal gradient measurements have been made in relatively shallow holes in several parts of the area. Although these measurements do not provide an accurate indication of deep heat flow or of temperatures at depths much greater than the depths of the holes, they often indicate the location of geothermal systems. The thermal gradient data can be used to compute near-surface conductive heat flow.

Regional Gravity Data

The regional gravity map (figure 6) was compiled from three primary sources: (1) the gravity map of the Richfield 1 x 2 degree quadrangle (Cook and others, 1981); (2) the gravity map of the Escalante Desert area (Pe and Cook, 1980); and (3) a compilation of part of northwest Utah by M. L. Zoback (written communication, 1983). These three surveys were adjusted to a common datum and recontoured. As this compilation was being prepared, a much more comprehensive effort by the USGS and UGMS was in progress to produce a gravity map and data base for all of Utah by merging the major regional data sets. The complete Bouguer anomaly map of figure 6 is adequate for the purposes of this study but will be superseded by the new Utah State map when it is published.

The Bouguer anomaly values range from -150 mgals in a gravity high along the east edge of the Sevier Desert to -240 mgals in local gravity lows along the east edge of the area. A general inverse correlation between the Bouguer anomaly values and regional surface elevations is apparent and reflects the isostatic effect of an underlying mass deficiency buoying up the higher regions. Superimposed on this broad regional gravity anomaly related to regional topography are local gravity anomalies. Most of the larger local anomalies are produced by the density contrast between Cenozoic sedimentary and volcanic rocks and the generally more dense older rocks. Other generally lower amplitude anomalies reflect mass anomalies within pre-Cenozoic rocks and Cenozoic rocks including major variations in the thickness of overthrust sheets, low-density igneous intrusions, and salt structures.

Regional Aeromagnetic Data

The regional aeromagnetic map (figure 7) is based on numerous surveys flown at several flight-line spacings and flight elevations. The individual surveys for the Richfield 1 x 2 degree quadrangle were adjusted to a common datum and projected to a level 3,680 m above sea level. This compilation was used for the part of the magnetic map in this quadrangle. The remainder of the map is from the Aeromagnetic Map of Utah (Zietz and others, 1976) and sources for this state map.

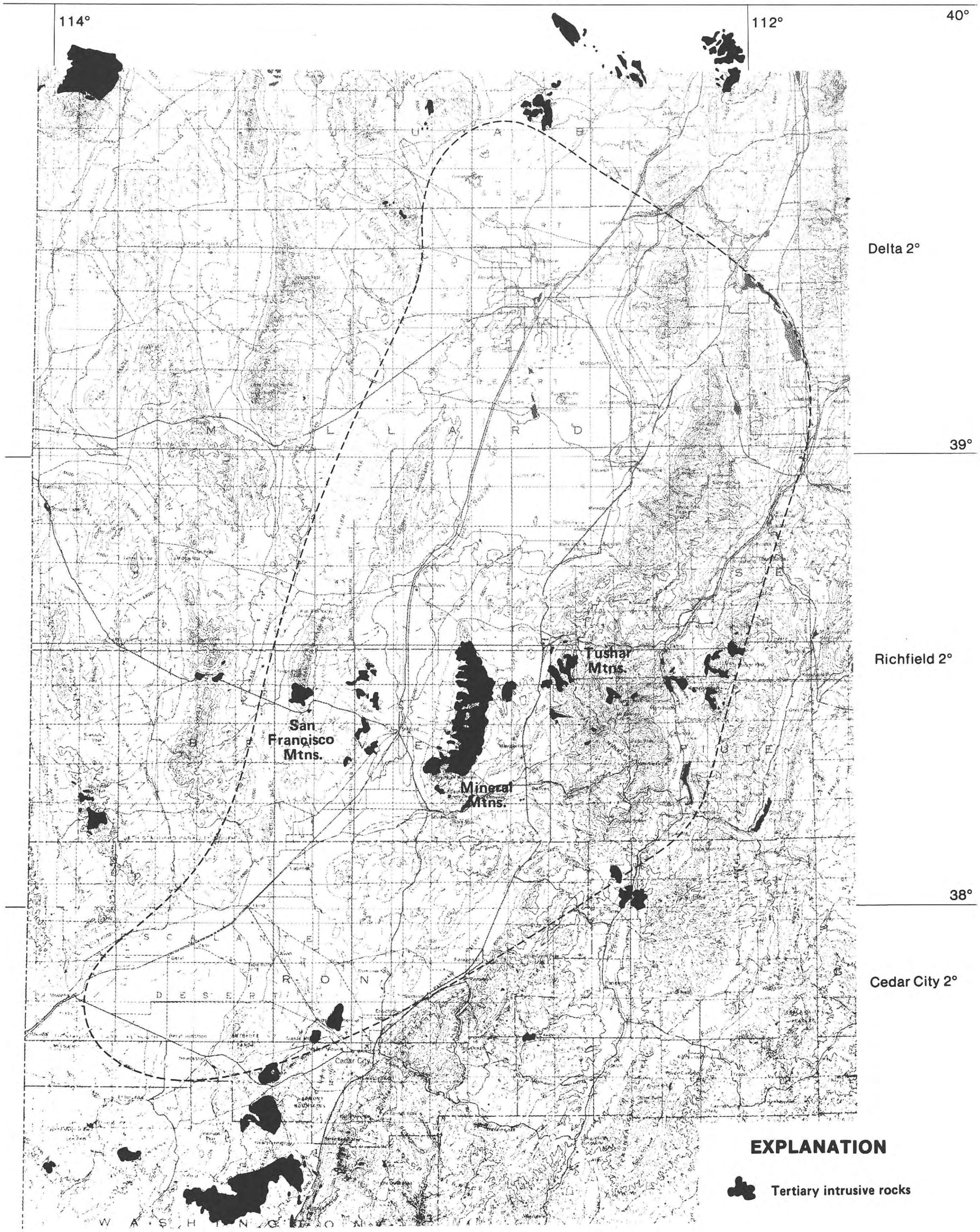


Figure 4. Map of the Sevier thermal area showing the Tertiary intrusive rocks from Hintze (1980) and Steven and Morris (1983b).

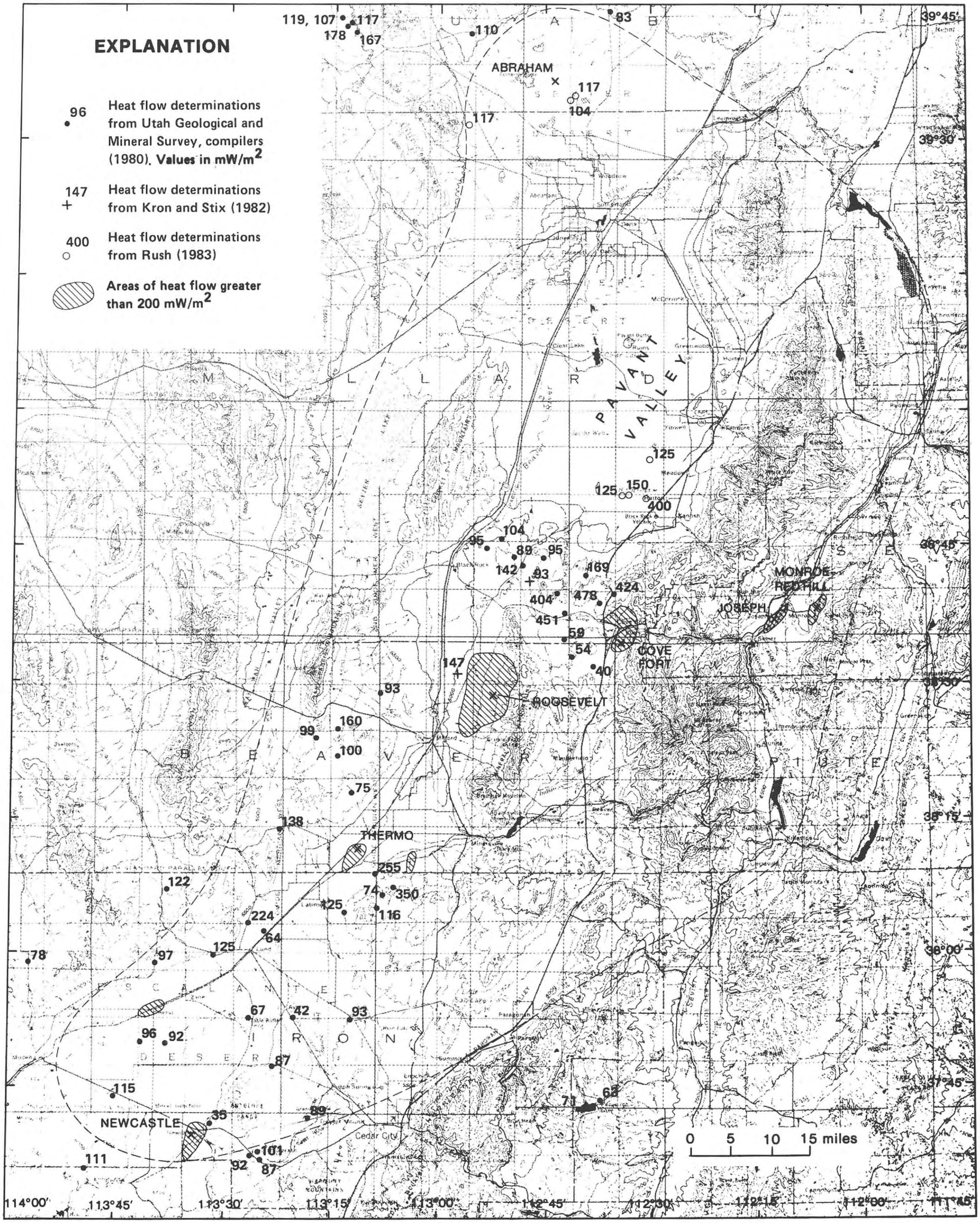


Figure 5. Heat flow map of the Sevier thermal area. Data range in quality from good to marginal with some values based on measurements in shallow drill holes and estimated thermal conductivity. Many values reflect shallow rather than deep conductive heat flow. Outline of areas of greater than mW/m^2 are based on data from Hulen and Sandberg (1981), Kron and Stix (1982), Murphy (1980), Rush (1983), and Ross and Moore (1985).

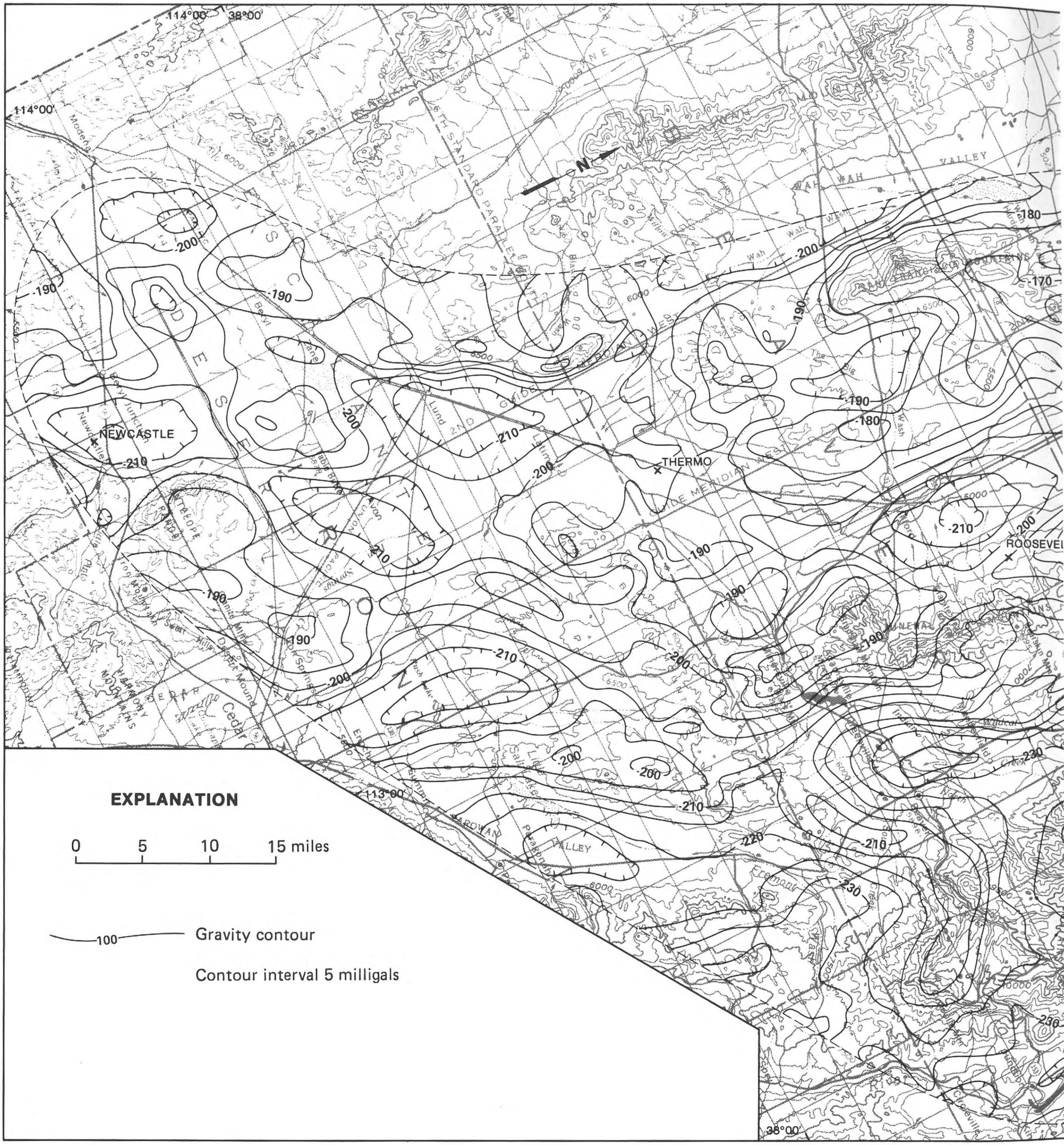
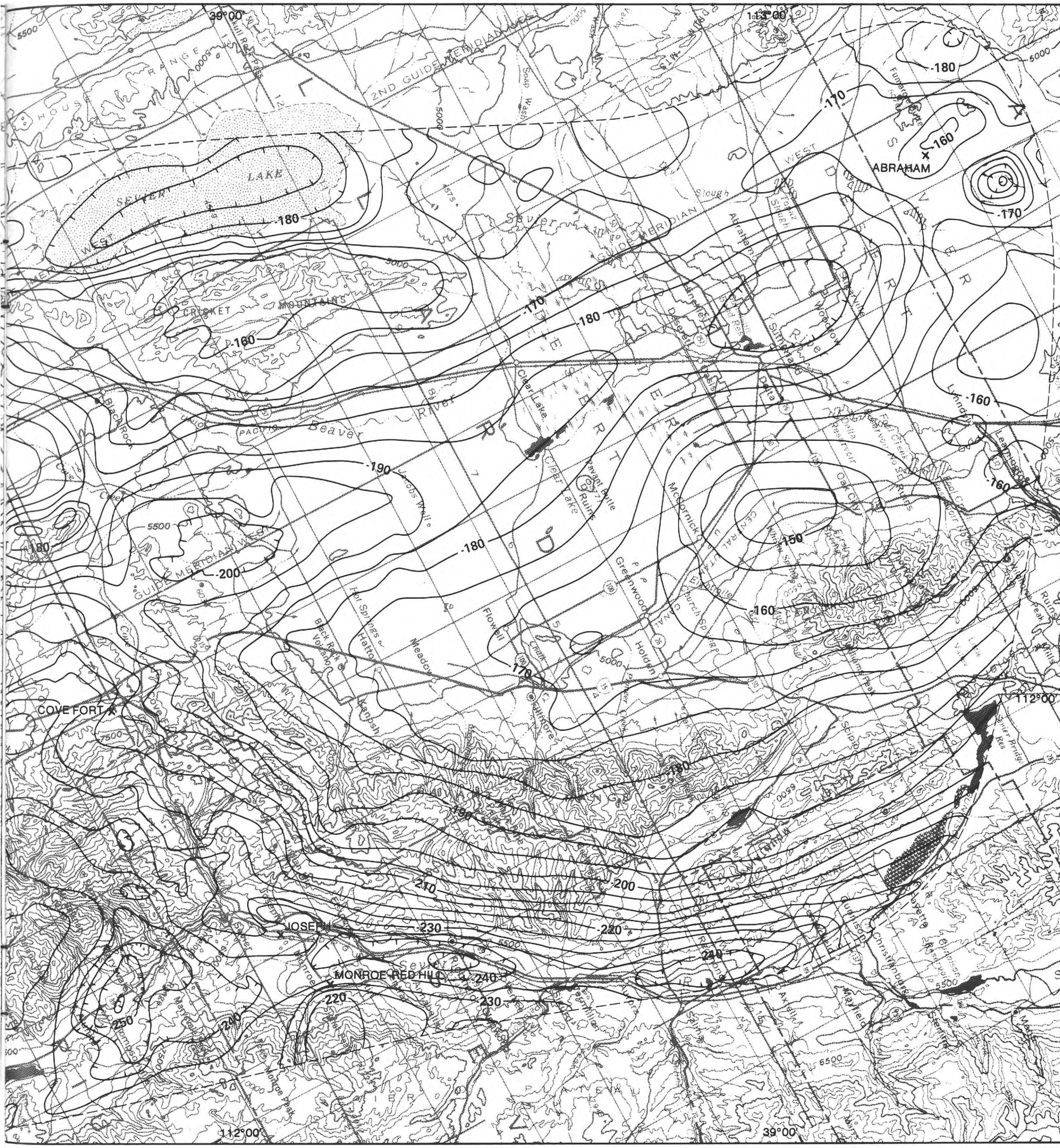


Figure 6. Complete Bouguer gravity anomaly map of the Sevier thermal area. Map is based on data from files of the University of Utah and the U.S. Geological Survey and a few stations established by the authors. Cook and others, (1981) Pe and Cook (1980), and U.S. Geological Survey unpublished



contour maps by M.L. Zoback, T.G. Holdenbrand, and D.R. Mabey are the primary data sources. Gravity datum is the Utah base station network (Cook and others, 1971). Data were reduced using an assumed density of 2.67 g/cc and the International Gravity Formula of 1930.

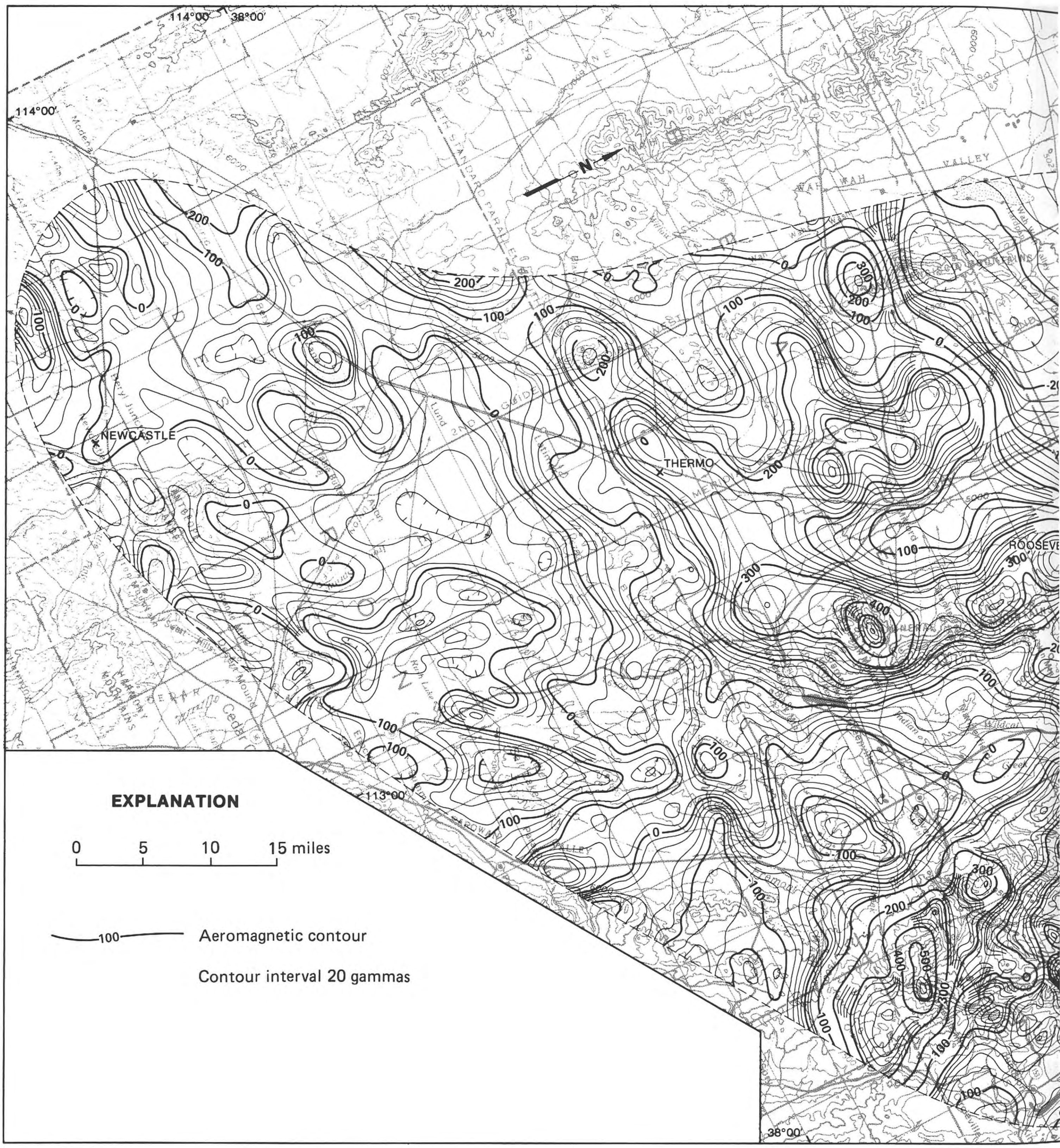


Figure 7. Residual aeromagnetic map of the Sevier thermal area. The data are from published sources and are primarily the same compilation used to produce the Aeromagnetic Map of Utah (Zietz and others, 1976) with modification in the Richfield 1 x 2 degree quadrangle. The data are from surveys

flown at several flight elevations mostly from 2700 to 3700 m above sea level. In the Richfield quadrangle the data have been projected to 3700 m above sea level.

The surveys used in compiling the state map were flown at 2,760 and 3,680 m above sea level. As with the gravity data, the USGS and UGMS are currently preparing a data set and map by accurately merging all the major aeromagnetic data sets of Utah.

The larger magnetic anomalies are in two west-trending belts. These are two of three major belts of Tertiary igneous rock that extend across western Utah (Mabey and others, 1978). Individual anomalies within these belts reflect primarily large intrusive bodies but extrusive units also produce substantial anomalies. Between these belts are local magnetic anomalies produced by Cenozoic volcanic rocks, and by deeper sources that may be Mesozoic plutons and/or Precambrian basement.

Seismicity and Quaternary Faults

The Intermountain seismic belt is an area of numerous earthquakes that extends north from southwestern Utah (figure 8). The belt is generally thought to reflect an intraplate boundary that, in Utah is approximately centered along the eastern edge of the Basin and Range Province. The eastern and southern two-thirds of the Sevier thermal area, which includes six of the seven geothermal systems, are within the belt (figure 8). Here the seismic belt arcs westward to become part of a west-trending zone that connects with active seismic areas in Nevada and California. The basin and range structures that characterize most of the Sevier thermal area suggest regional extension in an approximately east-west direction. However, fault-plane solutions for earthquakes occurring in the southeastern part of the area indicate a more complex strain pattern with a major component of horizontal compression indicated by some earthquakes.

Quaternary faults provide an indication of tectonic activity over the last 1.6 million years. The compilation of Quaternary faults in figure 9 is based on Nakata and others (1982). Several major zones of faults are apparent and some of the thermal systems are within these zones. Much of the northern part of the area was inundated by Lake Bonneville and to a lesser extent by earlier and later lakes. Surface evidence of Quaternary movement on some faults was likely destroyed by these inundations. Also shown on the map are areas where gravity data indicate the Cenozoic rocks to be more than 1 km thick. Most thick sequences of Cenozoic rock indicated by the gravity data in the valley areas are predominately Miocene and Pliocene sediments deposited in local basins.

Regional Geochemistry

A literature search to compile existing chemical data on water samples in the Sevier thermal area with temperatures greater than or equal to 20°C was conducted. Data were taken from the following sources: Bliss, 1983; Capuano and Cole, 1982; Cole, 1983; Goode, 1978; Klauk and Gourley, 1983; Lee, 1908; Mariner and others, 1983, McHugh and others, 1980, 1981; McHugh and Miller, 1981; Moore, 1980; Mower and Cordova, 1974; Mundorff, 1970; Sandberg, 1963; and Union Oil Company, 1978d, 1978e, 1979b. The analyses were entered into the elemental analysis program ELE at the ESL/UURI which generates trilinear diagrams and calculates geothermometers (Withrow, 1983). The discussion on the geochemical signature of the thermal water is based on common ion chemistry because trace elements were not available for most samples.

The locations and grouping of water samples are plotted on figure 10 along with directions of ground-water movement (Mower, 1965 and Sandberg, 1966). All ground-water flow in the study area is toward Sevier Lake. Ground-water flow is northward from the Escalante Valley, through Milford Valley, into the southern Pavant Valley through a narrow gap along the Beaver River north of Black Rock (Mower and Cordova, 1974). Ground water flows southward and westward into the Sevier Lake from the Sevier Desert and the Pavant Valley. Surface water in the Sevier River Valley on the east side of the study area is contained in the northward-flowing Sevier River which enters the Sevier Desert near Leamington.

Water samples are grouped geographically on figure 10 and are described by these divisions. Groups have been combined on the trilinear plots when applicable. Table 2 lists samples that form each group, the measured and calculated total dissolved solids (TDS), temperature, pH, and calculated geothermometer temperatures.

The following geothermometers were used to calculate reservoir temperatures for thermal water in the study area: 1) quartz conductive (Fournier, 1981), 2) chalcedony (Fournier, 1981), 3) Na-K-Ca (Fournier and Truesdell, 1974), and 4) Na-K-Ca with Mg correction (Fournier and Potter, 1979). The geothermometers generally calculate the maximum temperature of the thermal fluids along their travel path. The reliability of these geothermometers is dependent on five assumptions: 1) temperature-dependent reactions occur at depth; 2) an adequate amount of components are available for the temperature-dependent reactions; 3) water-rock chemical equilibrations occur at the reservoir temperature; 4) only minor equilibration occurs at lower temperatures as the water flows from the reservoir to the surface; and 5) hot water rising from depth does not mix with cooler, shallow ground water (Fournier and others, 1974).

If a Na-K-Ca geothermometer-calculated temperature is less than 100°C, then the silica content of the thermal water is a function of chalcedony and the chalcedony geothermometer should be used; a Na-K-Ca temperature of greater than 100°C indicates the silica temperature calculated assumes a silica content as a function of quartz solubility (Fournier, 1977). The Na-K-Ca geothermometer calculates reservoir temperatures that are too high for thermal waters with a large Mg content. A Mg-corrected geothermometer should be used when the Na-K-Ca geothermometer calculates a temperature greater than 70°C and R is between 5 and 50, $R = [Mg/(Mg + Ca + K)] \times 100$ using equivalent units of concentration (Fournier and Potter, 1979). In most instances the chalcedony and the Na-K-Ca geothermometers are discussed because they fit the above criteria and most consistently give similar results.

Generally ground water in higher elevations in the study area where surface discharge is low is of Ca HCO_3 to Na SO_4 character, while at lower elevations, in areas of increased discharge, water is Na Cl in character. All thermal water (greater than or equal to 20°C) is neutral to slightly basic, with the exception of the slightly acidic Roosevelt geothermal system water. Water from moderate- to high-temperature thermal areas is Na-Ca Cl in character, and the majority of the water outside of thermal areas is classified as Ca-Na HCO_3-Cl-SO_4, although the southwest portion of the study area has Na-Ca Cl-SO_4-HCO_3 water. The nomenclature used to describe water types is from Back (1961). TDS concentrations are discussed separately following the description of the thermal water in the study area. Some geochemical anomalies are evident.

Water samples from Abraham Hot Springs (Group A1) range from 82° to 84°C with a pH of 6.5 to 7.4. The water samples group closely on the trilinear diagram (figure 11) and are Na-Ca Cl in nature. Average reservoir temperature indicated by the chalcedony and Na-K-Ca Mg corrected geothermometers is 85°C. Water to the south (Group A2) in the vicinity of Delta in the northern part of the Sevier Desert ranges from 18° to 29°C with a slightly basic pH of 7.5 to 8.7. There are many low-temperature (less than 90°C) thermal wells in the Delta area. For most analyses Na concentrations were reported as Na + K so trilinear plots and the Na-K-Ca geothermometer calculations were not possible. Two samples that could be plotted fall in the Ca-Na HCO_3-Cl-SO_4 category. Chalcedony geothermometer temperatures average 51°C.

Group B1 in southern Pavant Valley is comprised of Meadow and Hatton Hot Springs which group closely as Na-Ca Cl water type (figure 12). Temperatures range from 29° to 41°C in the hot springs with pH between 6.5 and 7.5 (table 2). In October 1985, a temperature of 63°C was measured at Hatton Hot Springs. Chalcedony and Na-K-Ca (Mg corrected when applicable) temperatures average 77°C. Coyote Spring and Twin Peaks Spring, about 20 km to the southwest, form Group B2. This water is similar to Meadow and Hatton water although samples are more varied (figure 13). Temperatures range between 20° and 31°C, with Twin Peaks being hotter, and pH is slightly basic ranging from 7.4 to 8.8. The Twin Peaks samples (nos. 19, 101, and 115) are of Na-Ca Cl character. There is no agreement between the average chalcedony geothermometer temperature of 86°C, and the Na-K-Ca Mg corrected geothermometer average temperature of 56°C. The Coyote Springs and vicinity samples vary in character from Na-Ca Cl-SO_4-HCO_3 to Ca-Na Cl-SO_4-HCO_3. Average chalcedony and Na-K-Ca geothermometer temperatures are 67°C with the chalcedony temperature being consistently higher.

Samples from southern Pavant Valley outside of a thermal area (Groups B1 and B2) form Group B3 plotted on figure 12. The low-temperature sam-

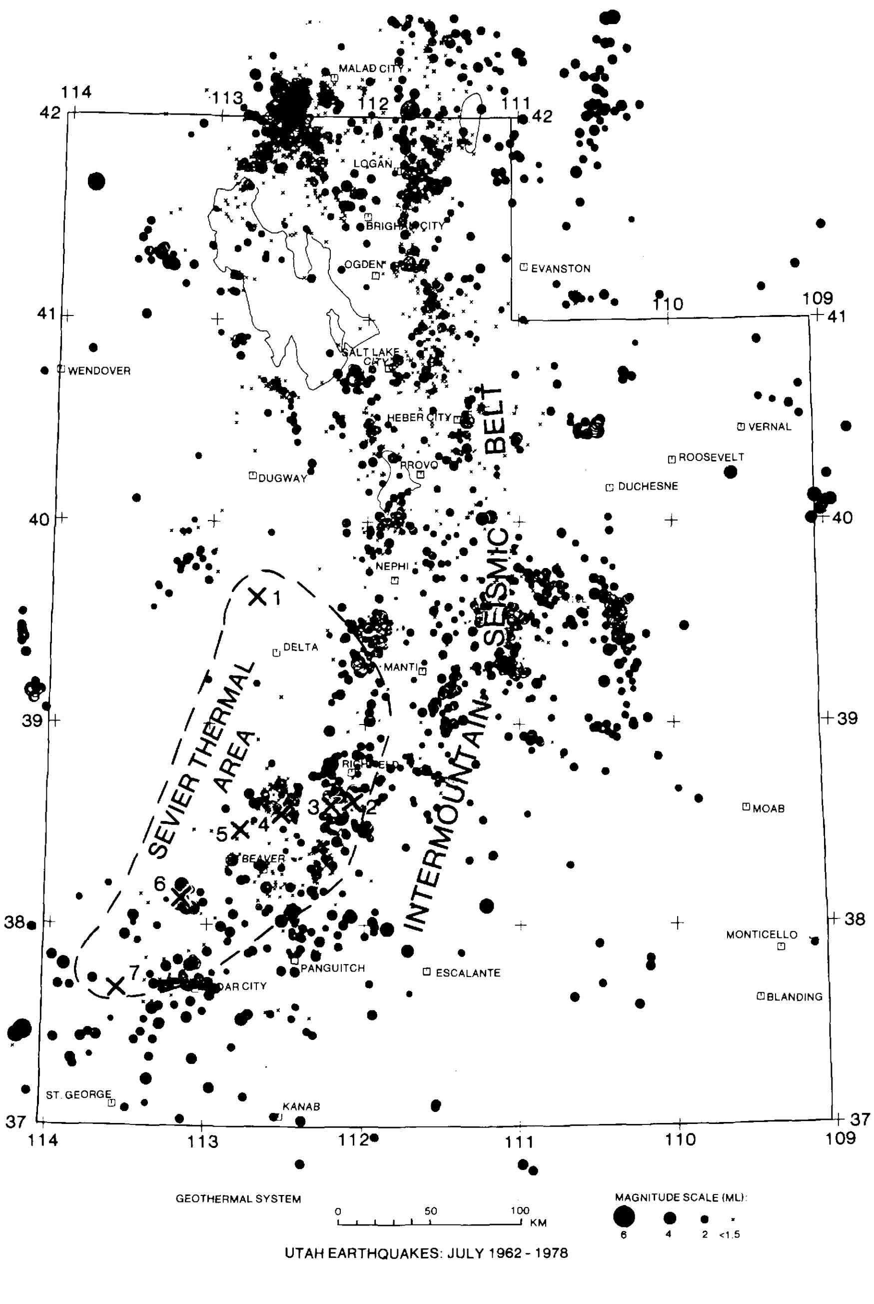


Figure 8. Epicenters of Utah earthquakes from July 1962 to June 1978 with magnitudes 2 or greater from Richins (1979). Also shown are the high-temperature geothermal systems, the Sevier thermal area, and the Intermountain seismic belt. Geothermal areas: (1) Abraham, (2) Monroe-Red Hill, (3) Joseph, (4) Cove Fort, (5) Roosevelt, (6) Thermo, (7) Newcastle.

TABLE 2. GEOCHEMICAL DATA ON THERMAL WATER SAMPLES - SEVIER AREA

Trilinear Plot Group	Sample No.-Description	TDS ppm Meas.	TDS ppm Calc.	T°C	pH	Qtz Conductive	Geothermometer °C Chalcedony	Na-K-Ca	Na-K-Ca Mg Corrected
A1	58-Abrah HS	3630	3621	82°	7.3	110	80	156	73 R=23.13
Abraham	94-Baker	—	3742	85°	7.4	115	86	163	91 R=18.6
	107-Crater	3692	3606	84°	6.5	117	89	164	90 R=18.8
A2	59	308	304	21.5°	7.5	74	42	37	—
Sevier Desert	60	209	215	20°	7.7	78	47	not calc	not calc
by Delta	61	202	201	20°	7.5	70	39	29	—
	62	492	491	21°	7.8	82	51	not calc	not calc
	63	262	262	23.5°	8.2	82	51	not calc	not calc
	64	363	363	26.5°	7.9	82	51	not calc	not calc
	65	1760	1758	29°	8.0	93	62	not calc	not calc
	66	508	501	26.5°	8.5	79	48	not calc	not calc
	67	230	229	28°	7.3	78	47	not calc	not calc
	68	292	292	26°	8.1	85	54	not calc	not calc
	69	339	339	24°	7.8	94	63	not calc	not calc
	70	248	247	25°	7.8	79	48	not calc	not calc
	71	281	293	26.5°	8.2	86	55	not calc	not calc
	72	461	437	18°	8.7	79	48	not calc	not calc
	73	834	833	21°	7.7	82	51	not calc	not calc
	74	2250	2244	25.5°	8.0	87	56	not calc	not calc
B1	21 Meadow?	—	4772	29°	6.6	104	74	203	90 R=20.8
Meadow/	86 Meadow	—	4681	41.1°	—	96	66	210	78 R=24.3
Hatton	87 Hatton	—	4663	37.8°	6.5	96	66	not calc	not calc
	95 Meadow	—	4808	41°	7.5	99	69	68	—
	113 Meadow	—	4824	33°	7.1	124	96	199	78 R=23.7
	114 Meadow	—	5621	24°	7.0	114	85	208	80 R=23.7
	116 Warm Spr	—	4779	35°	—	not calc	not calc	not calc	not calc
B2	2	—	1400	20.5°	7.6	114	85	67	—
Twin Peaks/	3	—	642	23.5°	7.8	94	63	45	—
Coyote Hills	19	—	7169	25°	8.8	98	68	94	63 R=29.7
	101 Tw Pk	—	4350	28°	7.6	111	81	96	50 R=35.8
	102 Cudahy	—	912	32°	7.7	111	81	71	Mg corr.<0
	109 Coyote Spr	—	460	20°	6.8	110	80	45	—
	115 Tw Pk	—	4076	31°	7.4	137	110	92	55 R=34
	117	—	558	20°	7.8	99	69	56	—
B3	15	—	291	20°	7.8	102	72	45	—
Southern	18	—	352	22°	8.7	77	45	21	—
Pavant	20	—	5047	21°	7.6	45	13	196	29 R=41.9
Valley	23	—	191	20°	7.3	45	13	not calc	not calc
	24	—	253	21°	7.7	51	18	20	—
C1	90 Sulphurdale	8816	8339	82°	—	150	124	not calc	not calc
Cove Fort	118 Well 31-33	7600	7187	59°	8.8	123	95	290	285 R=1.7
	119 Well 42-7	5200	5160	86°	8.5	143	116	415	404 R=1.8
	120 Well 14-29	4776	5121	69°	7.4	133	105	137	43 R=35

TABLE 2. Continued

Trilinear Plot Group	Sample No.-Description	TDS ppm Meas.	TDS ppm Calc.	T°C	pH	Geothermometer °C Qtz Conductive	Chalcedony	Na-K-Ca	Na-K-Ca Mg Corrected
D1	12 Mon	–	2870	82°	7.3	100	70	62	–
Monroe/Red	77 Mon	2700	2697	76°	7.6	105	76	184	111 R=15.1
Hill/Joseph	78 RH	2630	2619	76.5°	7.8	103	73	188	93 R=19.3
	79 Mon(CU)	2680	2675	42°	7.6	104	74	183	145 R=8.8
	80 Johnson	428	427	25°	7.4	82	51	15	–
	81 Joe	4970	4960	64°	6.6	128	100	139	83 R=20.2
	103 RH	3019	2787	76.5°	6.3	109	79	180	111 R=14.9
	104 Mon	2948	1820	70°	6.2	110	80	179	108 R=15.3
	105 Joe	5230	4997	63°	6.5	131	104	142	83 R=20.3
D2	16	–	481	21°	8.3	94	63	42	–
Sevier River	17	–	4526	21°	7.9	70	39	71	51 R=40.9
Valley	22	–	344	22°	7.7	51	18	6	–
	76 Richfield	307	302	22°	8.3	42	10	29	–
	82 Redmond	599	597	21°	8.0	92	61	77	41 R=45.6
E1	13	–	202	23°	8.1	51	18	28	–
Beaver Basin	14	–	306	20°	7.8	100	70	42	–
	25	–	342	21°	7.4	113	84	29	–
	88	–	253	20°	7.9	117	89	not calc	not calc
	89	–	712	21.1°	7.9	117	89	not calc	not calc
F1	11	–	375	20°	8.1	87	56	35	–
Parowan Valley	112	210	206	20°	8.0	109	79	50	–
G1	10	–	947	33.5°	7.7	82	51	81	Mg corr.<0
Minersville	48 Dotsons	1030	1006	32.5°	8.3	81	49	87	77 R=24.9
	49	1030	1013	33.5°	7.7	82	51	not calc	not calc
	50	1020	828	33°	7.4	39	7	185	72 R=24.6
	51	475	574	21.5°	8.3	92	61	not calc	not calc
	52	291	279	21°	7.5	99	69	55	–
	56	268	268	21°	8.2	86	55	25	–
	85	1020	864	13.5°	7.7	85	54	not calc	not calc
G2	4	–	5887	24°	6.0	98	68	237	137 R=14
Roosevelt	41 RHS-McKean	7040	6932	85°	–	234	222	293	284 R=2
	91 Seep	–	5764	25°	5.6	167	144	233	142 R=13
	92 Steam Well	–	6506	–	5.8	268	265	297	not calc
	96 Well 14-2	–	2861	–	–	116	87	291	not calc
	97 Well 54-3	6700	6462	–	6.7	209	193	296	not calc
	98 Well 72-16	6752	6740	92°	5.0	276	275	274	not nec. R<0.5
	99 Well 3-1	7067	7712	>205°	6.3	263	259	292	not calc
	100 Well 52-21	5727	5620	–	7.3	158	134	219	210 R=2.8
	108 Well 9-1	–	5890	225°	7.3	229	216	278	276 R=0.6
G3	5	–	525	20°	7.8	74	42	19	–
Milford Valley	26	–	3254	26.5°	9.5	70	39	116	R>50*
	27	–	316	21°	7.7	119	90	66	–
	28	–	611	20°	8.0	72	40	32	–
	29	–	465	20.5°	8.1	78	47	40	–
	30	–	170	20°	8.0	82	51	41	–

TABLE 2. Continued

Trilinear Plot Group	Sample No.-Description	TDS ppm Meas.	TDS ppm Calc.	T°C	pH	Qtz Conductive	Chalcedony	Na-K-Ca	Na-K-Ca Mg Corrected
	31	—	193	20.5°	8.1	78	47	54	—
	42	316	335	27°	7.9	118	90	90	63 R=30.3
	43	253	253	25.5°	8.2	86	55	62	—
	44	224	221	20.5°	8.0	75	44	32	—
	45	248	278	22°	—	86	55	not calc	not calc
	46	561	523	20°	7.9	97	67	49	—
	47	720	655	22°	7.7	111	82	8	—
	57	249	248	20.5°	7.7	89	59	not calc	not calc
H1	6	—	1207	66°	7.6	120	91	203	120 R=14.4
Thermo	40	1495	1464	78°	7.0	128	100	201	117 R=14.8
	53 Well-KGRA	253	249	22.5°	8.2	98	68	68	—
	54 Thermo	1500	1492	—	8.1	143	116	199	127 R=12.9
	55 Thermo	1490	1490	82.5°	7.8	148	122	213	130 R=13.5
	106 Thermo	1700	1509	89.5°	8.0	144	118	202	120 R=14.4
H2	83	1040	1029	30.5°	7.7	122	94	not calc	not calc
Newcastle	84 Chris. Bros.	1120	1112	95.5°	7.6	137	110	166	Mg corr.<0
H3	1	—	2486	23°	7.7	111	81	84	54 R=36.3
Escalante Valley	7	—	224	20°	8.1	104	74	62	—
	8	—	622	20°	7.9	109	79	54	—
	9	—	683	24°	7.8	55	23	40	—
	32	446	404	60°	9.1	105	76	94	not calc Mg<0
	33	1760	1646	27°	7.2	101	71	169	103 R=15.9
	34	1730	1592	20°	7.1	104	74	170	120 R=12.4
	35	482	470	20°	7.1	94	63	not calc	not calc
	36	724	626	20°	7.1	83	52	not calc	not calc
	37	304	349	20°	7.9	109	79	58	—
	38	672	554	22°	7.6	98	68	52	—
	39	1556	1563	28°	7.1	96	66	100	Mg corr.<0
	110 Junes Well	653	651	20°	7.9	74	42	38	—

*R>50 Therefore, underground water temperature probably equals measured temperature.

Geochemical data on thermal samples in the Sevier thermal area; not calc indicates no trilinear plot because Na reported as Na + K; dash under measured TDS, T°C, and pH columns indicates no data; geothermometer not calculated if data are lacking; dash under Na-K-Ca Mg corrected column indicates Mg correction does not apply; Mg<0=Mg less than defection limit.

ples range from 20° to 22°C with a slightly basic pH range from 7.3 to 8.7. With the exception of sample no. 20, the water is Ca-Na HCO_3-Cl-SO_4 in nature. The Na-K-Ca geothermometer temperatures exhibit greater agreement than the chalcedony geothermometer and average 29°C. The fifth sample (no. 20) is no warmer at 21°C but is chemically distinct from other samples in Group B3 with a very high TDS content. This water is Na-Ca Cl in character.

Cove Fort samples (Group C1), plotted on figure 14, show scatter due to different amounts of SO_4, increasing amounts of which change the water from Na Cl to Na-Ca Cl in character. Water in three exploratory wells drilled by Union Oil Company range from 59° to 86°C with a slightly basic pH range of 7.4 to 8.8. The quartz conductive geothermometer, the most appropriate one to use in this high-temperature system, averaged 137°C for all samples. This contrasts with a measured subsurface temperature of 178°C in an exploration test well (Ross and Moore, 1985).

Two high-temperature thermal areas in the Sevier River Valley, Monroe-Red Hill and Joseph, have been combined with Johnson Warm Spring (south of Monroe) to form Group

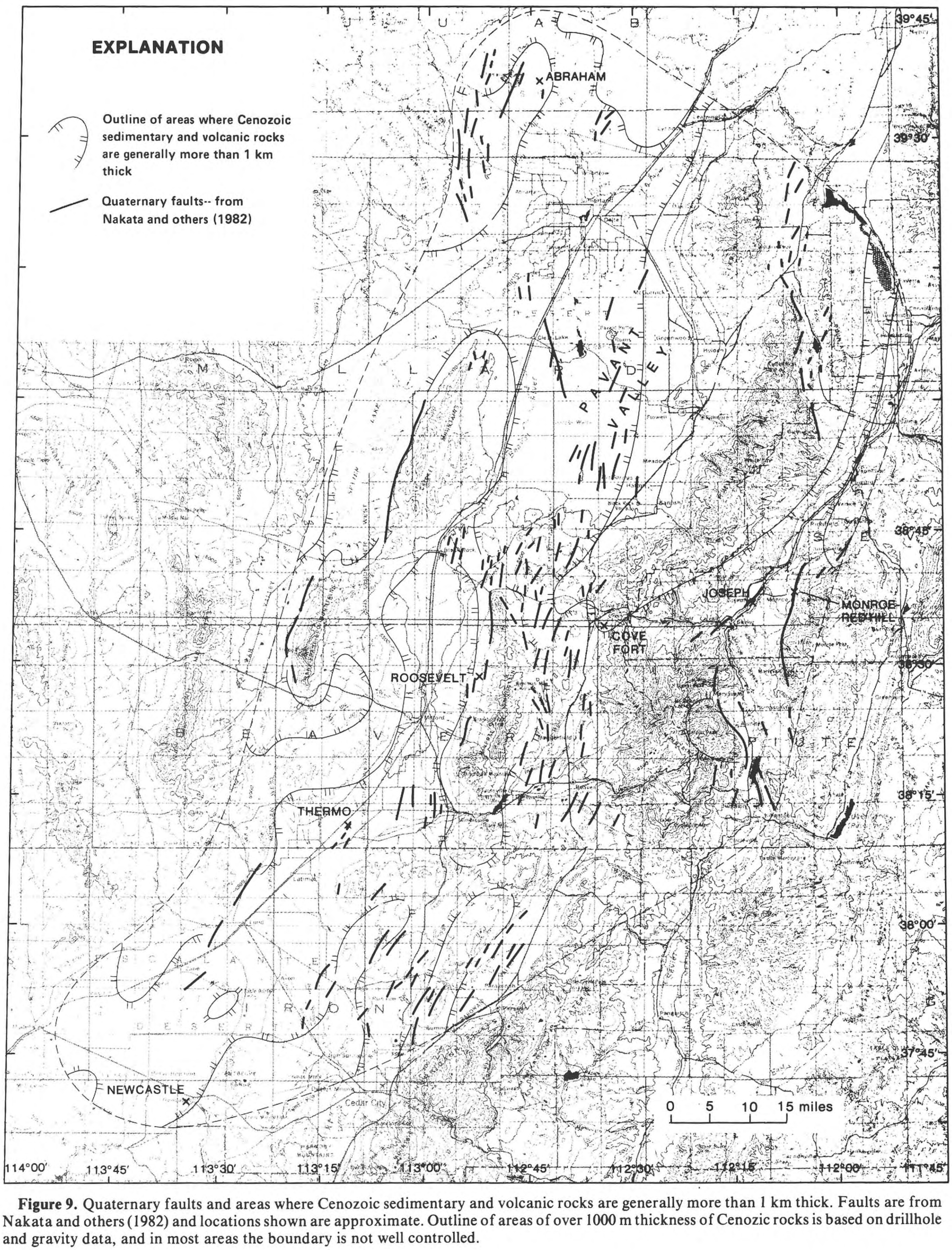


Figure 9. Quaternary faults and areas where Cenozoic sedimentary and volcanic rocks are generally more than 1 km thick. Faults are from Nakata and others (1982) and locations shown are approximate. Outline of areas of over 1000 m thickness of Cenozic rocks is based on drillhole and gravity data, and in most areas the boundary is not well controlled.

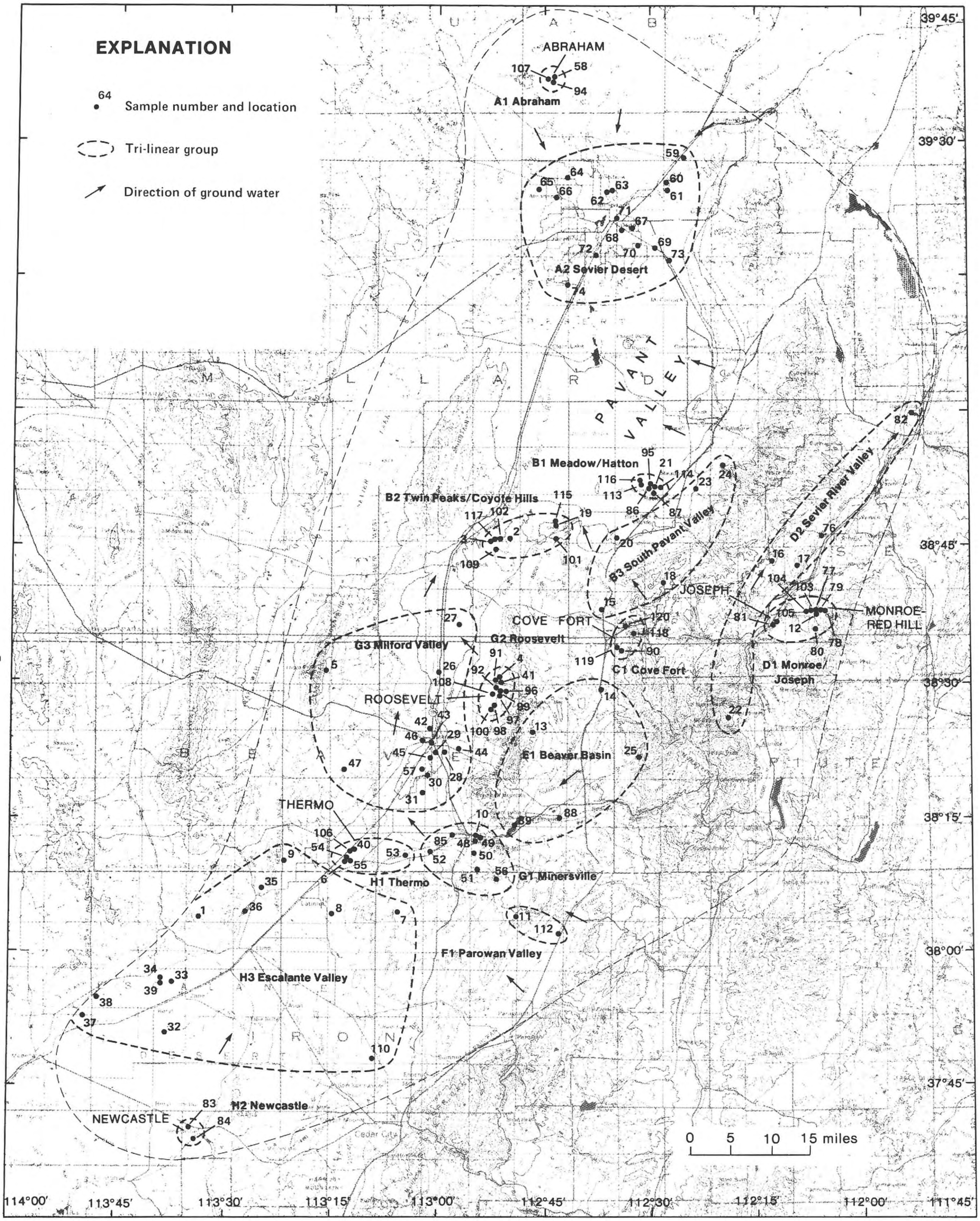


Figure 10. Map of the Sevier thermal area showing water sample numbers and locations, trilinear groups, and directions of ground-water movement. Data are from Bliss, 1983; Capuano and Cole, 1982; Cole, 1983; Goode, 1978; Klauk and Gourley, 1983; Lee, 1908; Mariner and others, 1983; McHugh and others, 1980, 1981; McHugh and Miller, 1981; Moore, 1980; Mower, 1965; Mower and Cordovas, 1974; Mundorff, 1970; Sandberg 1963, 1966; and Union Oil Company, 1978d, 1978e, 1979b.

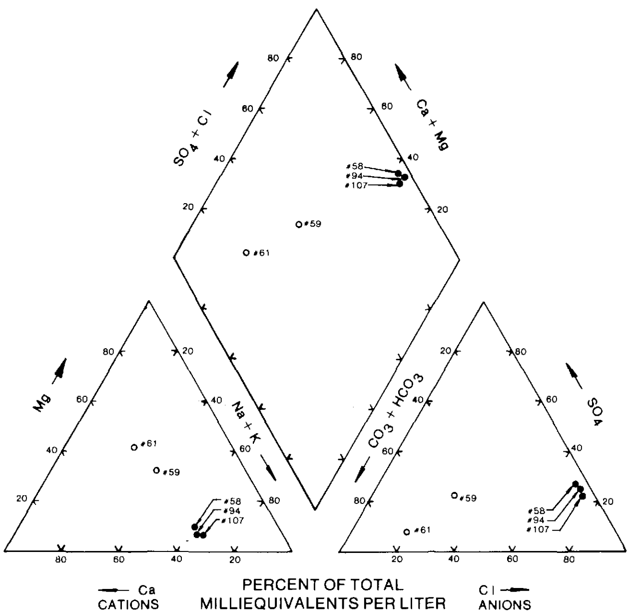


Figure 11. Piper diagram of common ions in Group A1-Abraham (●) and Group A2-Sevier Desert near Delta (o) in the Sevier thermal area.

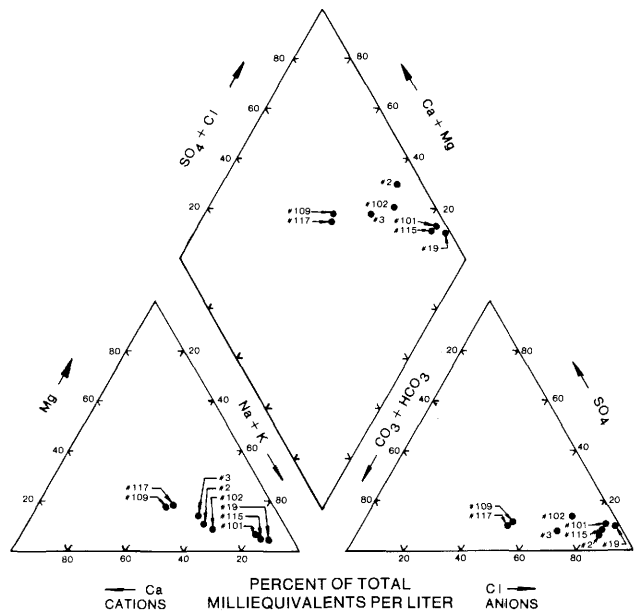


Figure 13. Piper diagram of common ions in Group B2-Twin Peaks/Coyote Hills (●) in the Sevier thermal area.

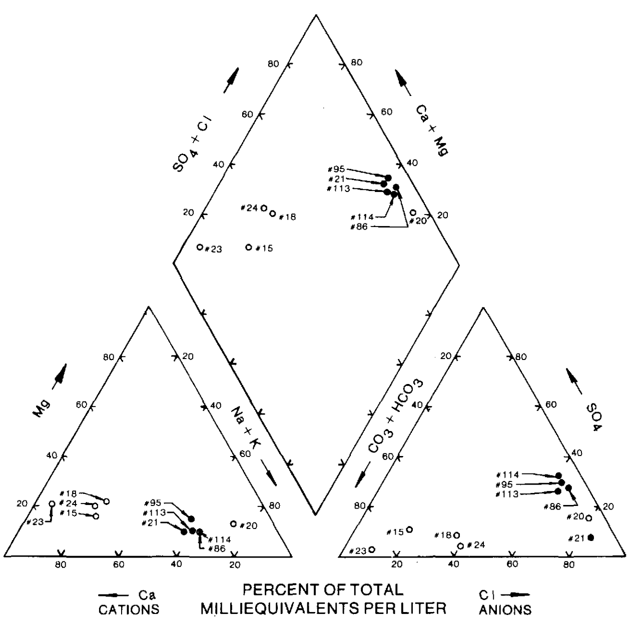


Figure 12. Piper diagram of common ions in Group B1-Meadow/Hatton (●) and Group B3-southern Pavant Valley (o) in the Sevier thermal area.

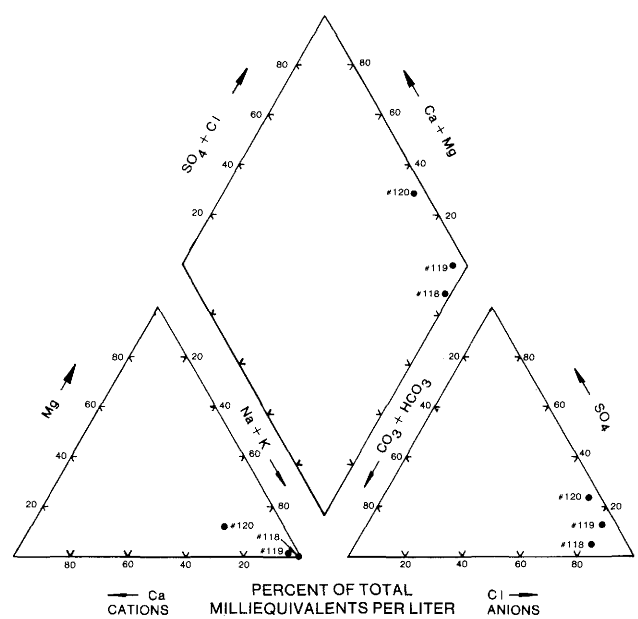


Figure 14. Piper diagram of common ions in Group C1-Cove Fort (●) in the Sevier thermal area.

D1. Joseph water is Na-Ca Cl, Monroe-Red Hill water is Na-Ca Cl-SO_4-HCO_3, and Johnson Warm Springs is Ca-Na Cl-SO_4-HCO_3 (figure 15). Temperatures for Monroe-Red Hill and Joseph range from 42° to 82°C, with all but one sample warmer than 62°C. The range for pH is 6.2 to 7.8. The average of the applicable quartz conductive or chalcedony geothermometers is 107°C which is similar to the average Na-K-Ca geothermometer temperature of 100°C. Johnson Hot Spring is cooler (25°C) and geothermometers are low and inconsistent—15°C (Na-K-Ca) and 51°C (chalcedony). The other Sevier River water samples plotted on figure 15 exhibit a large scatter, are low temperature (20° to 22°C), and slightly basic with a pH of 7.7 to 8.3. Geothermometer temperatures vary but are low.

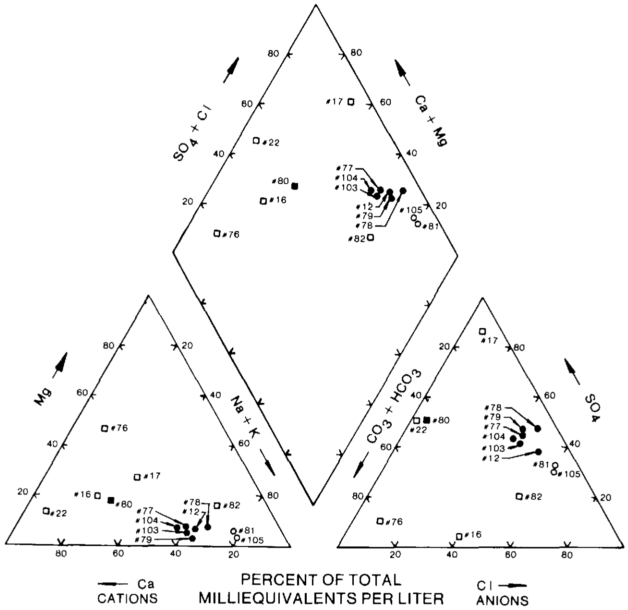


Figure 15. Piper diagram of common ions in Group D1-Monroe/Red Hill (●), Joseph (o), and Johnson (■), and Group D2-Sevier River Valley (□) in the Sevier thermal area.

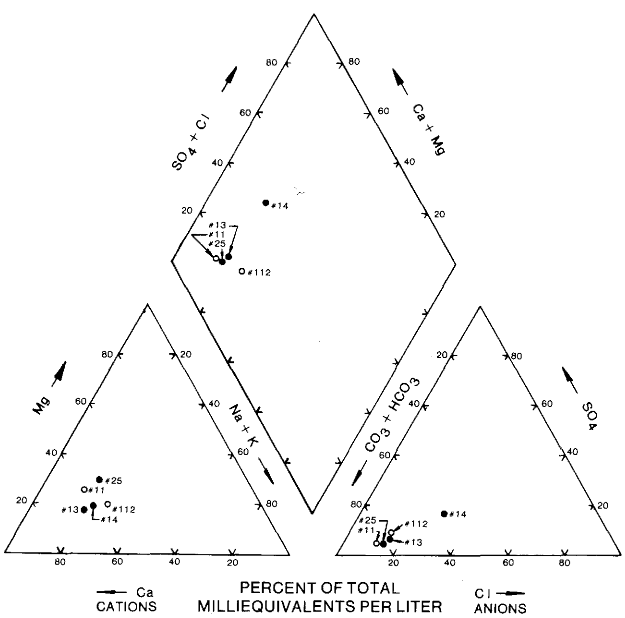


Figure 16. Piper diagram of common ions in Group E1-Beaver Basin (●) and Group F1-Parowan Valley (o) in the Sevier thermal area.

Beaver Basin water, which collects in Minersville Reservoir, and two Parowan Valley samples (nos. 11 and 112) are plotted on figure 16. Group E1 and F1 waters are Ca-Na HCO_3-Cl-SO_4, low temperature (20° to 23°C), and slightly basic (pH 7.4 to 8.1). There is no agreement between the Na-K-Ca and the chalcedony geothermometers; however, all calculated temperatures are less than 90°C.

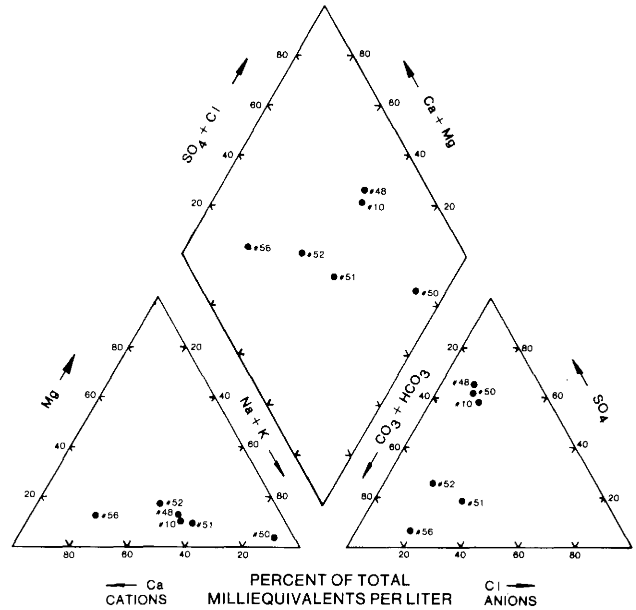


Figure 17. Piper diagram of common ions in Group G1-Minersville (●) in the Sevier thermal area.

In the vicinity of Minersville (Group G1) there are eight low-temperature (less than 90°C) thermal springs and wells, including Dotsons (Radium) Warm Spring (Murphy, 1980). Water is primarily Na-Ca Cl-SO_4-HCO_3 (figure 17) in character (sample nos. 52 and 56 are Ca-Na HCO_3-Cl-SO_4). Temperatures range from 13.5° to 33.5°C with pH from 7.4 to 8.3. The average temperature for the chalcedony geothermometer is 56°C, disregarding the temperature of 7°C for sample no. 50.

The Milford Valley samples (Group G3) are grossly similar to those in the Minersville area as is evident on Figure 18. Temperatures in the Milford Valley range from 20° to 27°C; the majority cluster near 20°C. A number of low-temperature wells (less than 90°C) are present in and around Milford. The pH values are consistently within 7.7 and 8.0 with the exception of sample no. 26 (pH 9.5). Sample no. 26 has a high TDS concentration and is discussed in a later section. Three samples (nos. 27, 42, and 47) have chalcedony geothermometer temperatures between 82° and 90°C. All other chalcedony reservoir temperatures are lower than 67°C. Na-K-Ca geothermometer temperatures are not consistent. All Roosevelt geothermal system samples (Group G2) plot in the Na Cl corner of figure 18. The temperature data for these samples is diverse: a bottom-hole temperature of 225°C (sample no. 108) in one exploratory well and 25°C (sample no. 91) at a seep in the area. The water for Group G2 is slightly acidic with a pH range of 5 to 7.3. The temperatures calculated with the Na-K-Ca geothermometer are fairly consistent and average a high 271°C for the resource being produced to generate electricity, while the quartz conductive geothermometer temperatures are less consistent and average 202°C. A maximum temperature of 254°C was measured in one of the wells tapping the resource (Ross, Nielson, and Moore, 1982).

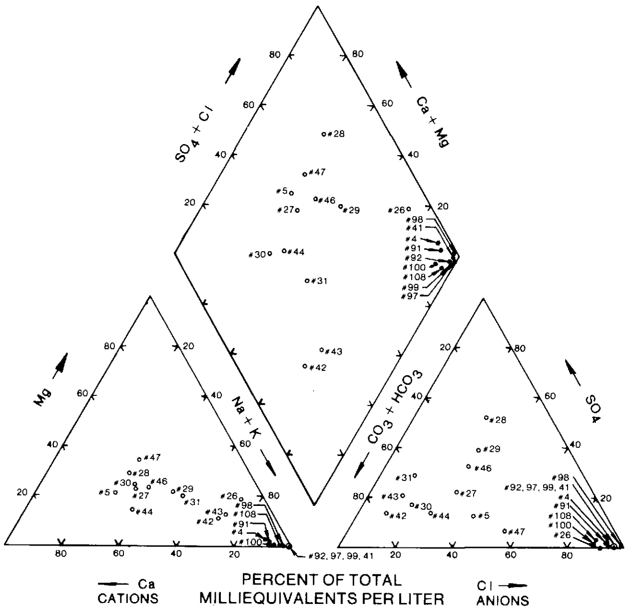


Figure 18. Piper diagram of common ions in Group G3-Milford Valley (o) and Group G2-Roosevelt (•) in the Sevier thermal area.

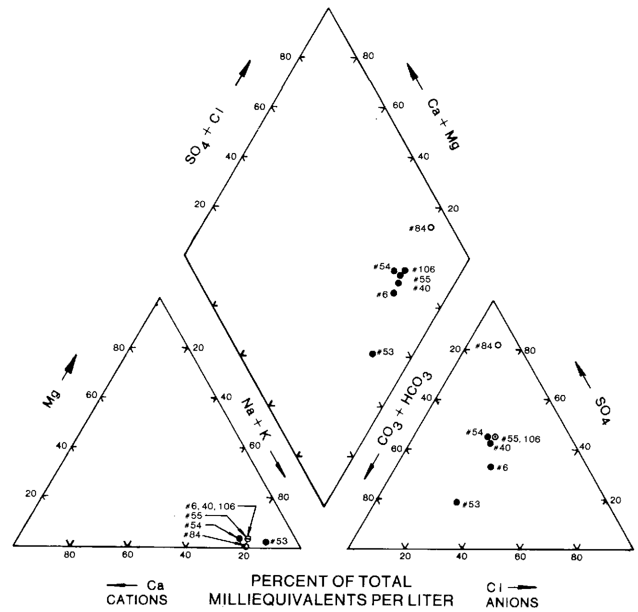


Figure 19. Piper diagram of common ions in Group H1-Thermo (•) and Group H2-Newcastle (o) in the Sevier thermal area.

The thermal areas in the southern portion of the study area, Thermo and Newcastle, make up Groups H1 and H2 in figure 19. Thermo Hot Springs water is Na-Ca Cl-SO_4-HCO_3 in character and groups closely with the exception of sample no. 53 from a well in the eastern part of the thermal area. Water temperatures at the hot springs range from 66° to 89.5°C and are slightly basic with a pH range of 7.0 to 8.2. The average Mg corrected Na-K-Ca geothermometer temperature of 123°C is similiar to the average quartz conductive temperature of 137°C for Thermo. The calculated reservoir temperature for sample no. 53 is 68°C for both Na-K-Ca and chalcedony geothermometers. Sample no. 84 from a well near Newcastle is Na-Ca Cl water at 95.5°C. A nearby well (sample no. 83), however, only has a temperature of 0.5°C. The slightly basic water has a pH range of 7.6 to 7.7. Quartz conductive geothermometer temperatures for the two samples are 122° and 137°C.

The remaining samples in the Escalante Valley (Group H3) plotted in figure 20 show data scatter similar to that of water samples from Milford Valley. The data can be grouped by water type as follows:samples 1, 8, 9, 36, 38, and 100 are Ca-Na Cl-SO_4-HCO_3; samples 33, 34, and 39 are Na-Ca Cl-SO_4-HCO_3; samples 35 and 37 are Ca-Na HCO_3-Cl-SO_4; sample 7 is Na-Ca HCO_3-Cl-SO_4; and sample 32 is Na HCO_3-Cl-SO_4. The only correlation found between location and water type is for samples 33, 34, and 39 northwest of Zane, which, along with sample no. 1, have anomalously high TDS values and are discussed in a later section. The slightly basic (pH range of 7.1 to 8.1) Escalante Valley samples are between 20° (most samples) and 28°C, with the exception of sample 32—a well sample of 60°C (pH 9.1). The Na-Ca-K geothermometer temperature for this sample is 94°C and the chalcedony temperature is somewhat lower at 76°C. The high TDS

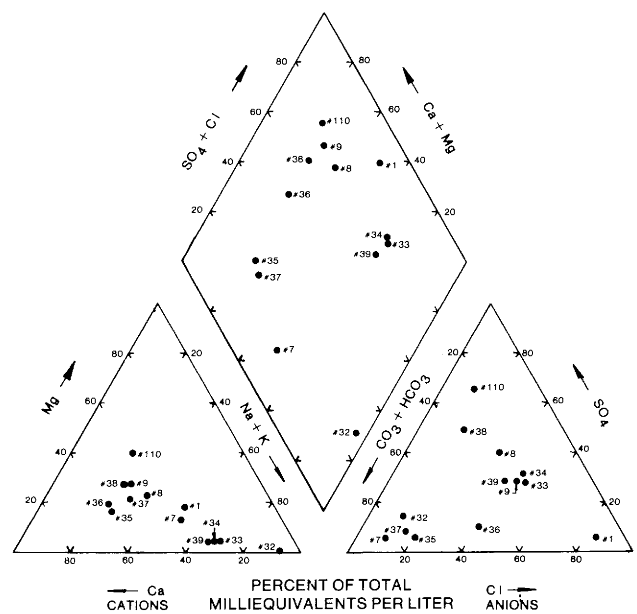


Figure 20. Piper diagram of common ions in Group H3-Escalante Valley (•) in the Sevier thermal area.

samples 33, 34, and 39 have an average Mg corrected Na-K-Ca geothermometer temperature of 108°C and an average quartz conductive temperature of 100°C, both fairly high reservoir temperatures. The remaining Escalante Valley samples have geothermometer temperatures of 51° (Na-K-Ca) and 62°C (chalcedony). Temperatures from the latter geothermometer exhibit more variation.

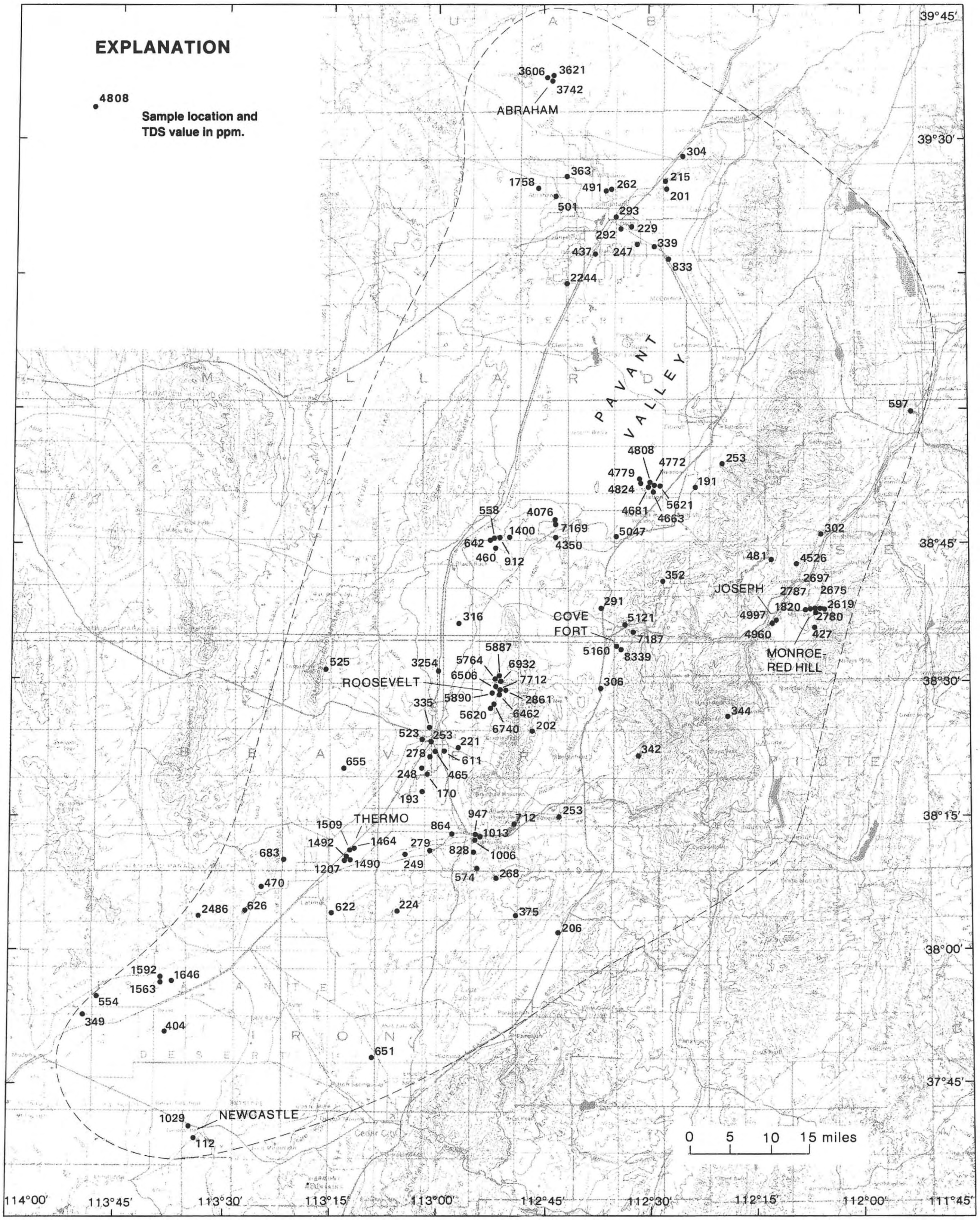


Figure 21. Map showing calculated total dissolved solids concentrations in ppm for water samples in the Sevier thermal area.

Total Dissolved Solids—The TDS concentrations are plotted on figure 21. TDS is calculated by the ELE program from the common ions with all HCO_3 calculated as CO_3. Measured TDS values, which measure all carbonate as CO_3, were not available for all samples. The water samples were collected at various depths, so represent TDS values from different aquifers causing some variation in the values from a single area.

Each thermal area—Abraham, Meadow, Hatton, Coyote Spring, Twin Peaks Spring, Monroe-Red Hill, Joseph, Cove Fort, Roosevelt, Thermo, Dotsons (Radium) Warm Spring, and Newcastle—is clearly depicted with TDS values greater than 1,000 ppm reflecting the high dissolved constituents typical of thermal water. Most TDS values are in the range of 1,000 to 7,000 mg/l, classifying these waters as slightly to moderately saline according to Hem (1970). There is no clear-cut linear relationship between increasing TDS content and increasing temperature for the thermal areas. This may be due to different rock and soil types, water chemistry of each thermal system, different flow paths sampled, or different depths of sample collection.

Most samples collected from non-thermal areas have TDS concentrations values ranging from 200 to 600 ppm. Some areas, however, have anomalously high TDS values:1) northwest and southwest of Delta, 2) southwest of Kanosh, 3) north of Joseph, 4) north of Milford, and 5) northwest of Zane.

The two samples near Delta with TDS values of 1,758 and 2,244 ppm have high concentrations of Na, Cl, and SO_4, a characteristic of thermal water. Both are well samples—sample 65 with a temperature to 29°C was taken at a depth of 304 m, whereas sample 74 is from a depth of 183 m at a temperature of 25.5°C (Goode, 1978). The chalcedony geothermometer does not indicate reservoir temperatures of any significance as a resource for these two samples. The high Na and Cl contents may be supplied from lake deposits into which both wells have been drilled (Mower and Feltis, 1968).

Southwest of Kanosh, a TDS of 5,047 ppm was calculated for sample 20. The chemistry of the sample is similar to that of Twin Peaks Spring (samples 19, 101, and 115) about 12 km to the west that also has elevated Na, Cl, and SO 4 concentrations. The surface temperature of sample 20 is only a few degrees below that of sample 19, however, geothermometer temperatures for sample 20 are considerably lower than those calculated for the other three samples (table 2). The sample is, however, in the immediate vicinity of Coyote and Twin Peaks Spring, Meadow and Hatton Hot Springs, and Cove Fort. In this area of apparent high geothermal potential the anomalous water chemistry should not be discounted.

Sample 17, north of Elsinore in the Sevier River Valley, is anomalous with a TDS concentration of 4,526 ppm primarily due to high SO_4. The sample is located near an area of strong limonite anomaly on a LANDSAT image map of limonitic rocks in the Richfield 1 x 2 degree quandrangle (Podwysocki and Segal, 1985). Limonitic rocks are possible indicators of hydrothermally altered rocks and contain various ferric iron oxide, oxyhydride, and sulfate minerals. The Na-K-Ca Mg corrected and chalcedony geothermometers, however, only calculate temperatures of 51° and 39°C, respectively.

Sample 26, from a spring north of Milford, has a TDS value of 3,254 ppm primarily due to a high Cl and Na content, a water chemistry markedly different from other samples in the Milford Valley. Sample 26 is located down gradient to the west of Roosevelt Hot Springs and the chemistry may reflect this influence. The chalcedony geothermometer temperature is 39°C.

Four samples northwest of Zane have anomalous water chemistry. Samples 33, 34, and 39 have high TDS values ranging from 1,592 to 1,646 ppm and are enriched in Na, Ca, SO_4, and Cl as compared to other Escalante Valley samples exclusive of thermal areas. Klauk and Gourley (1983) report that Li versus B plots for these samples were similiar to Li/B plots for Thermo Hot Springs samples, suggesting a geothermal anomaly in the area north of Zane. They applied the mixing model of Truesdell and Fournier (1977) to sample 33 and derived a temperature of 128°C, which is reasonable because the Na-K-Ca temperature of 103°C is probably low due to mixing. This temperature is comparable to the Na-K-Ca temperatures they calculated for Thermo Hot Springs (117°-124°C) which may be a non-mixing environment.

Sample 1, about 15 km northwest of samples 33, 34, and 39, has an even higher TDS value (2,486 ppm) due largely to Cl content. Other chemical parameters and geothermometer temperatures (table 2) are similiar for the four samples, thus increasing the size of the geothermal anomaly described by Klauk and Gourley (1983).

INDIVIDUAL GEOTHERMAL SYSTEMS

METHOD OF STUDY

The seven known high-temperature geothermal systems in the Sevier thermal area (table 1) have been evaluated individually, along with a geothermal system in an area of special geothermal interest in Pavant Valley. In evaluating these systems all available information was considered, nearly all of which is from reports available to the public. However, a minor amount of confidential data on the Cove Fort system that has not been released to the public was made available to the UGMS. Some temperature and geochemical data were obtained as part of this study and are also reported here. The purpose of the evaluation was to make an estimate of the quality and quantity of the resource in each system and to determine what techniques are likely to produce additional information useful in further defining and developing the resources, as well as in exploring for undiscovered systems.

Because most of the information used as the basis for this study is from published sources, not all of the data relating to the individual systems are repeated here; however, we have attempted to reference the most important studies. Examples of one or more applications of techniques that have proven most useful are included.

Case studies for three geothermal systems have been published by ESL/UURI (Hulen and Sandberg, 1981; Ross, Moore, and Christensen, 1982; Ross, Nielson, and Moore, 1982), and Rush (1983) presents moderately detailed data on five systems. The geothermal system at Roosevelt Hot Springs has been studied intensely and is currently being developed. The case study by Ross, Nielson, and Moore (1982) provides a good description of this system. No significant new data are available to us and we have not attempted to expand on their study. We did examine the regional setting of the Roosevelt system and have expanded on studies of other systems.

The resource assessment of Brook and others (1979) has been taken as a starting point for each of the seven known high-temperature systems. One of us (Mabey) was a member of the team that made that assessment and thus is familiar with the data and assumptions that went into the assessment. The same definitions and methodology used by Brook and others (1979) and Muffler and Guffanti (1979) are used in this report. Resource (or useful accessible resource base) refers to "the thermal energy that could be extracted and used at some reasonable future time." As in these early studies, this report considers only the resource within 3 km (about 10,000 ft) of the surface, accepting the rationale that this is approximately the limit of practical development with existing technology and that little information is available for greater depths. Energy is also reported in units of 10^{18} joules (J) which is approximately equal to 10^{15} British thermal unit (BTU) or one quad (a quadrillion BTU).

The names for most geothermal systems have traditionally been taken from the name of the hot springs that are usually the most apparent surface expression of the system. In this report the name of the hot springs refers only to the springs (example:Thermo Hot Springs), the term thermal area refers to the extent of a geothermal system (example:Thermo thermal area), and geothermal system refers to the entire system (example:Thermo geothermal system).

Local Geophysical and Geochemical Surveys

Several geophysical techniques have been tested or applied to investigations of local geothermal systems in the study area. At Roosevelt geothermal area nearly all of the standard techniques of geophysical exploration have been used; less work has been done on other systems. The local geophysical surveys can be classified as direct if the objective of the survey was to detect the resource, or indirect when obtaining information on some aspect of the geology relating to the resource. Direct methods may involve measuring temperature to detect heat escaping from the system, or one of several methods that might detect the presence of hot water or steam. Increasing temperature and alteration of rock normally increases its electrical conductivity. Also hot water is normally high in dissolved solids and is thus more conductive than cooler waters. A variety of geophysical methods have been used to detect electrical conductivity anomalies associated with geothermal systems. The movement of fluids within some geothermal systems produces self potential, and ground noise anomalies have been used to directly study the systems.

Geochemical techniques applied to geothermal exploration can also be classified as direct or indirect. Direct techniques supply data on the system or detect anomalies that are directly related to a geothermal system, whereas indirect methods relate to geologic controls. Geochemical techniques commonly involve analysis of water, soil, or gas samples. Some surveys detect components that have leaked from the system and others involve analysis of water or gas sampled from the system. Of particular importance are geochemical thermometers to estimate reservoir temperatures. In general this study relies on the geochemical thermometer work reported in Brook and others (1979) which is based directly on work by Mariner and others (1978). The estimated reservoir temperatures reported in these studies are conservative, primarily because mixing of thermal and nonthermal waters was not assumed. The reservoir temperatures estimated by Rush (1983) are generally higher because of assumed mixing. The water sample collected at Hatton Hot Springs was the only new chemical analysis relating to reservoir temperatures that was made as part of this study.

FUMAROLE BUTTE GEOTHERMAL SYSTEM

Crater Bench, about 30 km northwest of Delta, is a mass of basalt flows with a general surface elevation about 75 m higher than the surrounding floor of the Sevier Desert (figure 22). Fumarole Butte, a volcanic vent north of the center of the bench, is about 120 m higher than the desert floor. The black, fine-grained, vesicular to non-vesicular basalt flows that make up the main mass of Crater Bench are about one million years old. North of Crater Bench, The Hogback is composed of basalt and rhyolite flows about six million years old (Galyardt and Rush, 1981). Although the dates determined for the volcanic rock surrounding Fumarole Butte do not indicate any recent eruptions, during historic time vents of gas have been reported from the butte. The Old River Bed which lies immediately east of Crater Bench is the channel through which water from Sevier Lake overflows into the Great Salt Lake basin. About 15,000 years ago the entire area except for the top of Fumarole Butte was inundated by Lake Bonneville.

Several hot springs issue from a low mound east of Crater Bench. Topographic maps published by the USGS label these springs "Baker Hot Springs." However, USGS Professional Paper 1044-H (Rush, 1983) and USGS Map I-1297 (Galyardt and Rush, 1981) label them "Crater Hot Springs." USGS Circular 790 (Brook and others, 1979) lists them as "Abraham (Baker, Crater) Hot Springs." The Utah State Committee on Place Names has recommended that the name "Abraham Hot Springs" be used (J.M. Haymond, oral communication, 1985) and that name is used in this report. Because we conclude that the relationship of Abraham Hot Springs to the major geothermal system in the area may be different from that proposed by recent investigators, we use the name Fumarole Butte to refer to the geothermal system.

Rush (1983) inventoried about 40 spring orifices and estimated the total flow in February 1976 as 90 l/s. He estimated seepage as about 45 l/s and concluded that maximum discharge at that time was about 140 l/s. The maximum water temperature he measured was 87°C. Using geochemical geothermometers Rush calculated a reservoir temperature between 100° and 140°C. Using a mixing model he estimated a reservoir temperature of 140°C. Brook and others (1979) obtained

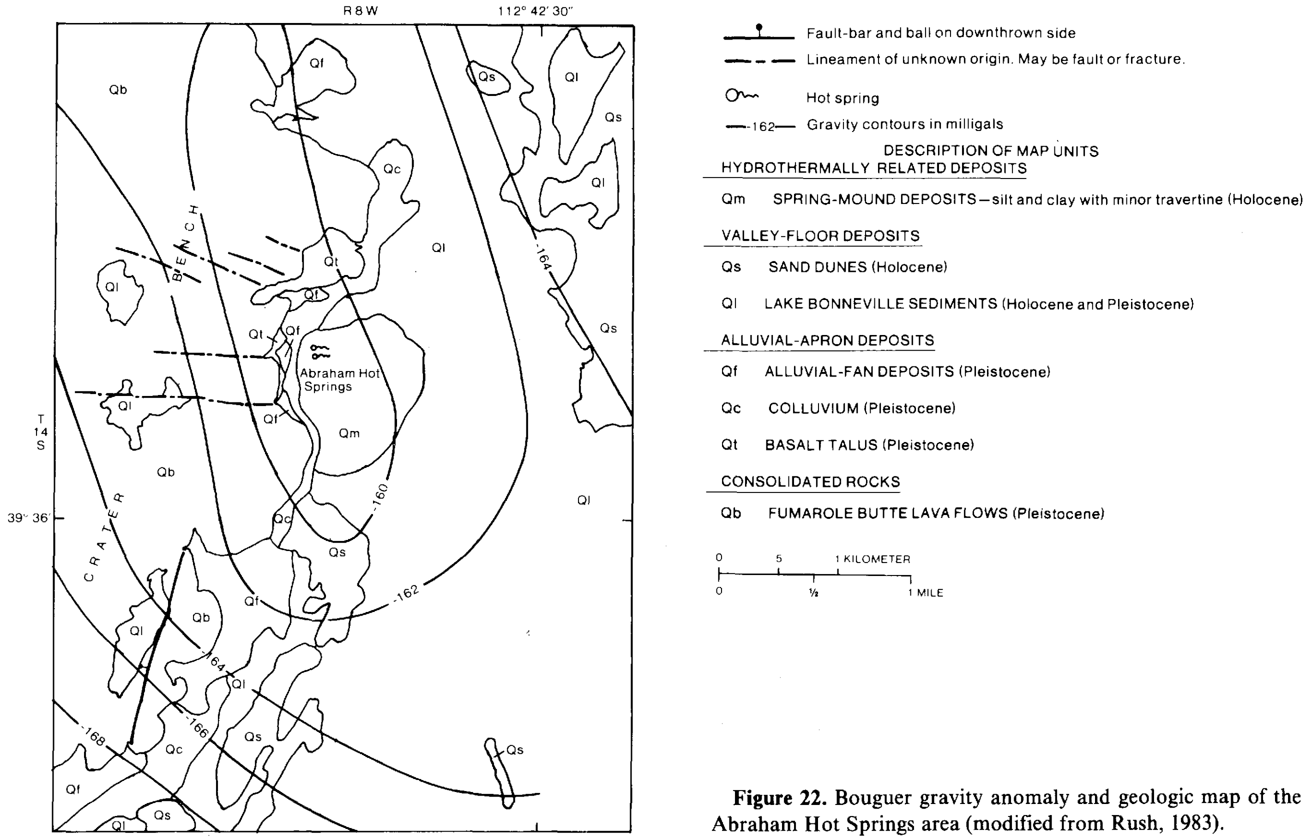


Figure 22. Bouguer gravity anomaly and geologic map of the Abraham Hot Springs area (modified from Rush, 1983).

calculated reservoir temperatures of 86°C (Na-K-Ca, Mg corrected), 89°C (chalcedony), and 117°C (quartz conductive) and calculated a mean reservoir temperature of 97 ± 7°C. They report a discharge of only 1000 l/min (16 l/s) and a measured temperature of 76°C. Rush estimated the heat flow in four water wells 5 to 10 km southeast of the hot springs as ranging from less than 42 to 120 mW/m^2. He assumed heat flow of 84 mW/m^2 and a thermal conductivity of 1.13 W/m/K to determine that the water discharging in the hot springs could have been heated to the measured temperature by circulating to a depth of 1,300 to 1,700 m. Rush concluded that, "the vertical flow of hot water to the land surface is assumed to be through a fault-controlled permeable zone that generally underlies hot springs in the Basin and Range province. However, no fault cutting the spring mound could be located during geologic field mapping."

Callaghan and Thomas (1939) report that in 1929 and 1930, 715 tons of ore were produced from shallow pits in the hot spring deposit, averaging about 20 percent manganese. They report an analysis of an ore sample that was 42.92 percent MnO, 17.93 percent Fe_2O_3, and 4.28 percent BaO. They concluded that the restriction of manganese to a single bed or horizon indicated that the extensive deposition of manganese was related to a distinct episode of the spring.

Fumarole Butte and Crater Bench, upon which it stands, appear to be primarily a constructural landform built by basalt flows erupting from the general area of the butte. Several normal faults and fissures have been mapped on the bench trending a few degrees east of north. Most of the faults are downthrown to the west. There are no faults mapped at the surface or inferred from the geophysical data that suggest the east side of Crater Bench is fault controlled. The hot springs do not appear to be aligned in a pattern that suggests a fault. The "spring-mound crest deposits" form a nearly circular outcrop while the outcrop of "spring-mound deposits" is slightly elongated north-south (Rush, 1983). The gravity data across the east side of the bench do not define a gradient that suggests a fault on this side of the bench.

In addition to the water discharged at Abraham Hot Springs there is evidence of thermal activity at Fumarole Butte. Gilbert (1890) wrote:

> *Before visiting this butte I had listened with incredulous interest to the statement that smoke or steam was sometimes seen to rise from it, but personal observation subsequently removed all doubt. About the outer edge of the summit are thirty or forty crevices from which warm, moist air gently flows. The permanence of the phenomenon is attested by the verdure lining the openings—a deep green moss glistening with moisture and vividly contrasting alike with the somber rocks and the sparse, ashen vegetation without. In different openings I found the temperatures 62°, 70°, 72°, and 73.5°F, all above the atmospheric mean for the locality, which is approximately 55°. At the time of the observation the outer air had a temperature of 30°, and was dry. A little mist formed over some of the openings, but was reevap-*

orated within a few feet. On days that are moist, cool and still, a conspicuous cloud must arise. It can hardly be doubted that this thermal manifestation testifies to a residium of volcanic heat in the old flue.

A group of hot springs at the southeastern base of the mesa may have the same significance. Their temperatures range from 110°to 178°F.

Ives (1947) examined the fissures in 1946 and reported the warm, moist air currents from the fissures were slightly acidic in odor and nauseating when inhaled for extended periods. He measured an air temperature of 110°F (43°C) in one fissure (ambient temperature at the time was 40°F). Smith and others (1978) state, "On top of the volcanic neck, warm moist air is still issuing forth from several cracks in the rock, possibly indicating residual heat." Murphy (1980) indicates a thermal spring at the approximate site of the butte labeled Fumarole Butte with an indicated water temperature of 23°C. No source is cited for this information. No spring at this location is listed in any of the standard lists of thermal springs, and we have been unable to find evidence supporting the existence of a spring at the butte.

In September, 1985 gas flowing from the fissures in several places along the summit rim of the butte was sampled. On the southeastern side of the summit an opening about 0.5 m long, and varying in width from about 5 cm at one end to about 25 cm at the other, was filled with green moss and was emitting warm, moist gas. No odor was associated with the vapor. A temperature of 23°C was measured in the opening about 1 m below the surface with an ambient temperature of 15°C. Approximately 6 m below the summit on the eastern side of the butte two more fissures were found. Warm, moist gas also rose from these moss-filled cavities although with not as much velocity. The two openings were wider, about 15 cm by 30 cm, and seem to be related to the same joint system at depth that can be seen along the eastern, massive, side of the butte having the most well-developed cooling joints (figure 23). A temperature measured in these fissures about 3 m below the surface was 20.5°C. Fissures mapped to the south and west of the butte (Galyardt and Rush, 1981) were examined in hopes of detecting discharging gas but none was found.

Two gas samples were collected from the fissure on the summit of the butte. Analyses were made by Anatec Laboratories, Inc. and are given in table 3, along with the other gas samples. Samples 1-4 were collected 300 ml, stem-type gas bombs and samples 5-7 collected in 600 ml, stem-type gas bombs. All samples were analyzed by Lucas cell scintillation counting to determine radon-222 content. Observed activities were adjusted to correct for decay which occurred between sampling and analysis. Other sample headspace gases (argon, oxygen, nitrogen, methane, helium, hydrogen, and carbon monoxide) were quantified by gas chromatography with thermal conductivity and flame ionization detection. Carbon dioxide, concentrated in the sample alkaline aqueous phase, was determined by acid evolution-infrared absorption. Ammonia and hydrogen sulfide, also contained in the aqueous phase, were quantified using an ammonia-specific electrode and alkaline oxidation-turbidimetry, respectively. The only

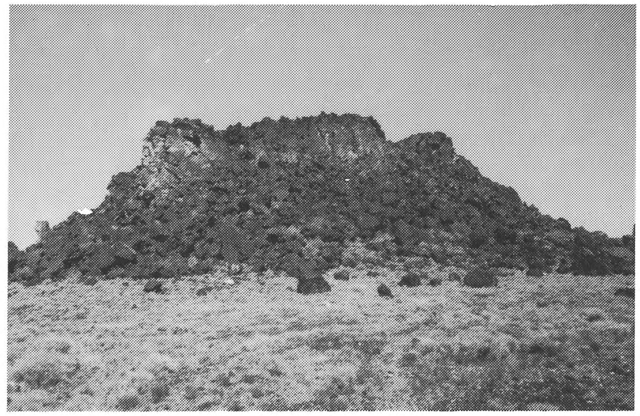


Figure 23. Photograph of the southeast side of Fumarole Butte; note cooling joints in massive area.

constituents present in detectable levels in samples 1 and 2 are carbon dioxide and carbon monoxide. The field configuration of the fumarole imposed difficult sampling procedures, and a very high air content, about 99 percent, was present in the initially evacuated sample flasks. The lack of detectable amounts of the other gases may reflect inefficient sampling but does not negate the possibility that they are components in the gas. The radon level is very high in the samples. Anomalous radon concentrations have been used to locate covered faults related to a reservoir in areas of known geothermal potential (Nielson, 1978). Radon is released from rock as a product of the U 238 decay series and radon flux is increased in areas of fracturing. The Fumarole Butte area is surrounded by north-trending faults (Galyardt and Rush, 1981) and the basalt cone exhibits columnar jointing. Radon in the two samples may be moving to the surface through these faults and joints.

Samples 3 and 4 were taken from Abraham Hot Springs to the east of Fumarole Butte, temperature 83°C at the sampling site. Carbon dioxide, ammonia, argon, nitrogen, helium (sample 4), and carbon monoxide were detected in these samples. The radon level is much lower than in the Fumarole Butte samples.

The primary source of subsurface information in the immediate area of the Fumarole Butte geothermal system is geophysical. The regional gravity map (figure 6) defines a gravity high trending a little west of north with the axis passing directly through Abraham Hot Springs. On the more detailed 2 mgal contour map (figure 22) the anomaly has 10 mgals of closure and is 13 km long and 6 km wide. Rush (1983) suggests the anomaly may be caused by a "shallow body of volcanic rock or hot water deposits of relatively high density." Smith and others (1978) present two density models to explain the measured anomaly. Both models assume the unconsolidated sediments underlying the Sevier Desert are about 200 m thick and that most of the gravity high is produced by a dike-like mass within the underlying bedrock. In both models the mass extends from a depth of about 300 to 1200 m with a density of 2.96 or 2.97 g/cc. One mass is about 1.5 to 2.5 km wide and the other 1.5 to 6 km wide. Both models also show the basement (Precambrian?) stepping about 2,000 m down to the west

TABLE 3. GAS ANALYSES - FUMAROLE BUTTE, ABRAHAM HOT SPRINGS, COVE FORT

SAMPLE NO. 1 — FUMAROLE BUTTE

Sample gas/steam ratio (ft^3/lb):	3.32E-02
Sample gas/steam ratio (moles/1000 moles steam):	1.67
Sample gas/steam ratio ($g/10^6$ grams steam):	3939.67
Percent Air	98.98
Total weight of condensate (grams):	1.10
Initial headspace pressure (psi):	11.90
Radon (pCi/L):	32008

Gas	Mole % (w/o H_2O)	Moles per 1000 moles H_2O	ppm (with H_2O)
Water Vapor	N/A	N/A	9.96E+05
Carbon Dioxide	9.30E+01	1.55E+00	3.77E+03
Total Sulfur (as H_2S)	<5.23E-01	<8.71E-03	<1.64E+01
Ammonia	<2.95E+00	<4.92E-02	<4.63E+01
Argon	<1.31E+00	<2.18E-02	<4.81E+01
Nitrogen	<1.31E+00	<2.18E-02	<3.37E+01
Methane	<7.27E-01	<1.21E-02	<1.07E+01
Helium	<9.01E-02	<1.50E-03	<3.32E-01
Hydrogen	<7.27E-02	<1.21E-03	<1.35E-01
Carbon Monoxide	1.7 E-02		

SAMPLE NO. 2 — FUMAROLE BUTTE

Sample gas/steam ratio (ft^3/lb):	2.70E-02
Sample gas/steam ratio (moles/1000 moles steam):	1.36
Sample gas/steam ratio ($g/10^6$ grams steam):	3207.03
Percent Air	99.04
Total weight of condensate (grams):	1.30
Initial headspace pressure (psi):	11.70
Radon (pCi/L):	58990

Gas	Mole % (w/o H_2O)	Moles per 1000 moles H_2O	ppm (with H_2O)
Water Vapor	N/A	N/A	9.97E+05
Carbon Dioxide	9.27E+01	1.26E+00	3.06E+03
Total Sulfur (as H_2S)	<5.45E-01	<7.40E-03	<1.40E+01
Ammonia	<3.08E+00	<4.18E-02	<3.93E+01
Argon	<1.37E+00	<1.87E-02	<4.12E+01
Nitrogen	<1.37E+00	<1.87E-02	<2.89E+01
Methane	<7.65E-01	<1.04E-02	<9.21E+01
Helium	<9.47E-02	<1.29E-03	<2.85E-01
Hydrogen	<7.64E-02	<1.04E-03	<1.16E-01
Carbon Monoxide	1.7 E-02		

SAMPLE NO. 3 — ABRAHAM HOT SPRINGS

Sample gas/steam ratio (ft^3/lb):	6.45E-01
Sample gas/steam ratio (moles/1000 moles steam):	32.39
Sample gas/steam ratio ($g/10^6$ grams steam):	69001.03
Percent Air	58.60
Total weight of condensate (grams):	2.26
Initial headspace pressure (psi):	9.25
Radon (pCi/L):	3147

Gas	Mole % (w/o H_2O)	Moles per 1000 moles H_2O	ppm (with H_2O)
Water Vapor	N/A	N/A	9.35E+05
Carbon Dioxide	6.46E+01	2.09E+01	4.78E+04
Total Sulfur (as H_2S)	<1.34E-02	<4.35E-03	<7.69E+00
Ammonia	1.36E-01	4.41E-02	3.90E+01
Argon	4.42E-01	1.43E-01	2.97E+02
Nitrogen	3.48E+01	1.13E+01	1.64E+04
Methane	<1.43E-02	<4.63E-03	<3.86E+00
Helium	<1.77E-03	<5.74E-04	<1.19E-01
Hydrogen	<1.43E-03	<4.63E-04	<4.85E-02
Carbon Monoxide	1.9 E-03		

SAMPLE NO. 4 — ABRAHAM HOT SPRINGS

Sample gas/steam ratio (ft^3/lb):	1.82E+00
Sample gas/steam ratio (moles/1000 moles steam):	91.23
Sample gas/steam ratio ($g/10^6$ grams steam):	176456.57
Percent Air	20.55
Total weight of condensate (grams):	1.46
Initial headspace pressure (psi):	7.90
Radon (pCi/L):	6559

Gas	Mole % (w/o H_2O)	Moles per 1000 moles H_2O	ppm (with H_2O)
Water Vapor	N/A	N/A	8.50E+05
Carbon Dioxide	4.21E+01	3.84E+01	7.97E+04
Total Sulfur (as H_2S)	<7.27E-03	<6.63E-03	<1.07E+01
Ammonia	1.27E-01	1.16E-01	9.31E+01
Argon	1.21E+00	1.10E+00	2.08E+03
Nitrogen	5.65E+01	5.15E+01	6.81E+04
Methane	<7.05E-03	<6.43E-03	<4.87E+00
Helium	9.35E-02	8.53E-02	1.61E+01
Hydrogen	<7.05E-04	<6.43E-04	<6.13E-02
Carbon Monoxide	6.1 E-03		

TABLE 3. Continued

SAMPLE NO. 5 — WELL 66-28, COVE FORT

Sample gas/steam ratio (ft^3/lb):	3.81E+00
Sample gas/steam ratio (moles/1000 moles steam):	191.27
Sample gas/steam ratio ($g/10^6$ grams steam):	464491.91
Percent Air	0.00
Total weight of condensate (grams):	9.90
Initial headspace pressure (psi):	0.65
Radon (pCi/L):	50

Gas	Mole % (w/o H_2O)	Moles per 1000 moles H_2O	ppm (with H_2O)
Water Vapor	N/A	N/A	6.83E+05
Carbon Dioxide	9.81E+01	1.88E+02	3.13E+05
Total Sulfur (as H_2S)	1.44E+00	2.75E+00	3.55E+03
Ammonia	9.83E-03	1.88E-02	1.21E+01
Argon	<1.20E-04	<2.30E-04	<3.48E-01
Nitrogen	3.18E-01	6.08E-01	6.46E+02
Methane	<7.26E-04	<1.39E-03	<8.44E-01
Helium	<8.99E-05	<1.72E-04	<2.61E-02
Hydrogen	1.25E-01	2.40E-01	1.84E+01
Carbon Monoxide	<1.2 E-02		

SAMPLE NO. 6 — WELL 47-6, COVE FORT

Sample gas/steam ratio (ft^3/lb):	1.26E+00
Sample gas/steam ratio (moles/1000 moles steam):	63.16
Sample gas/steam ratio ($g/10^6$ grams steam):	154013.42
Percent Air	79.67
Total weight of condensate (grams):	1.30
Initial headspace pressure (psi):	11.20
Radon (pCi/L):	1287

Gas	Mole % (w/o H_2O)	Moles per 1000 moles H_2O	ppm (with H_2O)
Water Vapor	N/A	N/A	8.67E+05
Carbon Dioxide	9.97E+01	6.30E+01	1.33E+05
Total Sulfur (as H_2S)	<2.31E-02	<1.46E-02	<2.40E+01
Ammonia	1.96E+01	1.24E-01	1.01E+02
Argon	<3.24E+02	<2.05E-02	<3.94E+01
Nitrogen	<3.24E+02	<2.05E-02	<2.76E+01
Methane	<1.80E-02	<1.14E-02	<8.79E+00
Helium	<2.24E-03	<1.41E-03	< 2.72E-01
Hydrogen	<1.80E-03	<1.14E-03	< 1.11E-01
Carbon Monoxide	1.9 E-02		

SAMPLE NO. 7 — WELL 47-6, COVE FORT

Sample gas/steam ratio (ft^3/lb):	4.14E-01
Sample gas/steam ratio (moles/1000 moles steam):	20.77
Sample gas/steam ratio ($g/10^6$ grams steam):	49370.30
Percent Air	96.29
Total weight of condensate (grams):	0.53
Initial headspace pressure (psi):	9.50
Radon (pCi/L):	1681

Gas	Mole % (w/o H_2O)	Moles per 1000 moles H_2O	ppm (with H_2O)
Water Vapor	N/A	N/A	9.53E+05
Carbon Dioxide	9.51E+01	1.98E+01	4.60E+04
Total Sulfur (as H_2S)	<1.73E-01	<3.59E-02	<6.46E+01
Ammonia	3.80E+00	7.89E-01	7.10E+02
Argon	<3.47E-01	<7.21E-02	<1.52E+02
Nitrogen	<3.47E-01	<7.21E-02	<1.07E+02
Methane	<1.93E-01	<4.01E-02	<3.40E+01
Helium	<2.39E-02	<4.97E-03	<1.05E+00
Hydrogen	< 1.93E-02	<4.01E-03	4.28E-01
Carbon Monoxide	2.1 E-02		

ANALYTICAL PRECISION

Following are data obtained on replicate analysis of one gas sample for which results are reported. Data are relative standard deviations (RSD) (or average relative standard deviations) calculated using data reported as "ppm with water."

Analyte	Relative Standard Deviation (%)	Matrix of Determination
Carbon Dioxide	0	Condensate/NaOH Solution
Total Sulfur (as H_2S)	NA[1]	Condensate/NaOH Solution
Ammonia	0	Condensate/NaOH Solution
Argon	6.4	Residual Gas Phase
Nitrogen	5.3	Residual Gas Phase
Methane	NA	Residual Gas Phase
Helium	NA	Residual Gas Phase
Hydrogen	NA	Residual Gas Phase
Air	1.8	Residual Gas Phase

NA[1]—Not applicable if analyte not detected in replicated sample.

AIR CORRECTION PROCEDURE

Analytical results on the preceding pages are adjusted without oxygen present in sources. Oxygen quantitated in samples is therefore considered to represent sample contamination by atmospheric air. Concentrations of argon, nitrogen, and oxygen measured in samples have been adjusted by the following procedure:

(1) Total moles of O^2 determined in sample are multiplied by 1.00 and subtracted from total moles of O^2.
(2) Total moles of O^2 determined in sample are multiplied by 3.73[a] and subtracted from total moles of N.
(3) Total moles of O^2 determined in sample are multiplied by 0.0446[a] and subtracted from total moles of Ar.

[a]Molar argon: oxygen: nitrogen ratios are derived from, "Components of atmospheric air," *in* CRC handbook of chemistry and physics, 65th ed.

under the west side of Crater Bench. The model with the smaller mass assumes that part of the gravity high is produced by a bedrock high.

The regional aeromagnetic data (figure 7) define a prominent arcuate high extending east-west with the axis passing approximately through both Crater Bench and Abraham Hot Springs. There is no apparent correlation between the gravity and magnetic anomalies. A correlation would be expected if the dike inferred from the gravity data were igneous rock. Part of the magnetic anomaly has a near-surface source and likely relects the basalts in Crater Bench. There is also a deeper regional component that cannot be explained by surface geology. The anomaly is the most southern of a complex of magnetic anomalies that make up the Park City-Bingham magnetic zone, which is the northernmost of the three west-trending zones of magnetic anomalies in western Utah reflecting belts of Cenozoic igneous rocks (Mabey and others, 1978). A narrow magnetic high extends south from the Fumarole Butte area. This high is parallel to but offset from a north-trending magnetic high in the area of abundant basalt flows west of Fillmore. The position of these two comparable magnetic anomalies suggests that two similar but not continuous zones of basaltic eruption occur in the Sevier Desert.

Shawe (1972) presents evidence that three Tertiary calderas are aligned in an east-trending zone north of Fumarole Butte. Parts of the boundaries of the western and central calderas are defined by the surface geology. The boundary of the eastern caldera and much of the boundaries of the other two are covered by post-caldera sedimentary and volcanic rock, and thus only the approximate location can be inferred from surface geology. Shaw inferred that the southeast boundary of the central caldera, which he named the Keg caldera, is near the northwest side of Crater Bench.

A large regional aeromagnetic high occurs in the area of the three calderas proposed by Shawe (1972). The western, northern, and eastern boundaries as inferred by Shawe encompass the areas of highest magnetic intensity, but on the south the zone of high magnetic intensity extends beyond the boundaries inferred by Shawe in an area of complete post-caldera cover. The magnetic anomaly suggests that the calderas may extend farther south and that the Fumarole Butte geothermal system may be located where a north-trending zone of Quaternary volcanism intersects the south edge of an older caldera complex.

The numerous Quaternary faults in the Fumarole Butte area and to the west suggest that this is an active seismic area, but historic seismicity near the butte is low (figure 8). The butte lies about 50 km west of the Intermountain seismic belt and no earthquakes have been recorded in the immediate vicinity of the butte. Two earthquakes of magnitude four or greater have been reported a few kilometers to the west and in 1974, 1975, and 1976 a swarm of small earthquakes occurred 30 to 50 km northwest of the butte (Richins, 1979).

Several attempts have been made to map the resistivity structure in the area. Johnson (1975) interprets resistivity and induced polarization surveys as indicating that basalt flows of Crater Bench are 90 to 120 m thick and rest on low-resistivity sediments. Rush (1983) presents an interpretation by A. A. R. Zohdy of a resistivity sounding across the bench (figure 24). This interpretation shows basalt about 90 m thick above interbedded basalt and alluvium or weathered basalt about 130 m thick. Underlying this is about 1,000 m of low-resistivity sediments underlain by basement (pre-Tertiary). Both interpretations agree that high-resistivity basalt flows are underlain by low-resistivity material which both presume to be sediments. Johnson indicates a resistivity of 1 to 3 ohm-m for these sediments. A profile of audiomagnetotelluric soundings was made through Abraham Hot Springs (Senterfit and Bedinger, 1976). Pseudosections based on these soundings show a surface layer about 100 m thick with an apparent resistivity ranging from 4 to 160 ohm-m. This is underlain by a layer at least 300 m thick with a resistivity between 1 and 4 ohm-m. There is evidence in the audiomagnetotelluric data of a higher resistance to one side of the profile. The easternmost stations show this effect the least, suggesting that the resistive mass may be basalt flows underlying Crater Bench to the west.

If Zohdy's interpretation is correct and the general level of bedrock under Crater Bench is about 1,200 m below the surface, it is possible to explain all of the gravity high as being produced by a buried bedrock ridge trending parallel to the general basin and range features of the region. Because there is no magnetic expression of the postulated ridge, it is likely to be composed of pre-Tertiary sedimentary rocks rather than igneous rocks. Total relief on the ridge would be about 1000 m if the density contrast between the ridge and enclosing low-density sediments is 0.4 g/cc.

Basalt is generally believed to originate at great depths, migrate to the surface in relatively small conduits, and dissipate most of its original heat as it cools on the surface. Normally the one million year old basalt flows in Crater Bench would not suggest that a near-surface heat reservoir underlies the vent. However, the venting of warm, moist gas in the vicinity of Fumarole Butte is strong evidence that a geothermal system underlies the butte. This could involve a reservoir of heat within the volcano or a convecting system controlled by Quaternary faults. The presence of the hot springs 6.5 km to

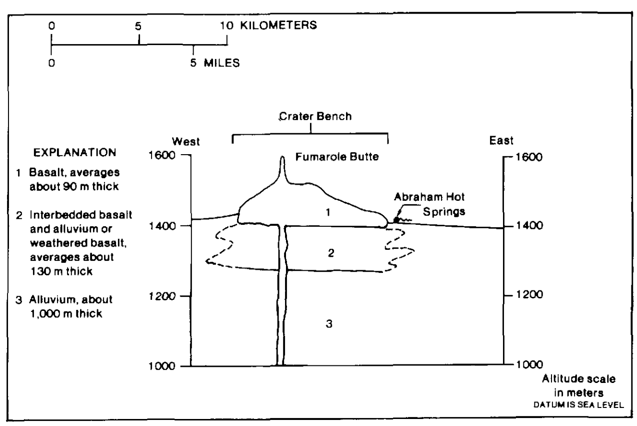


Figure 24. Generalized cross section through Fumarole Butte and Abraham Hot Springs based on Schlumberger resistivity soundings from A.A.R. Zohdy (modified from Rush, 1983).

the east, with no evidence of the structural control that would provide the deep circulation required to heat the water being discharged, suggests that the hot springs and the venting at Fumarole Butte may be manifestations of the same geothermal system.

At least two basic models for the Fumarole Butte geothermal system should be considered, one is the fault-controlled system proposed by Rush (1983). In this model, an inferred normal fault provides a conduit for the upward leakage of hot water from a reservoir underlying the area of the spring. The inferred fault or a similar fault is part of a deep circulation system wherein the water is heated by a temperature gradient that is near the regional normal and no local heat source is required. Models similar to this can explain the occurrence of most hot springs along range fronts in the Basin and Range Province.

A second model that is consistent with all available data is shown in figure 25. In this model the venting at Fumarole Butte is assumed to reflect a local reservoir of heat relatively near the surface. Water heated by this source migrates upward, moves to the east (down gradient), and is forced to the surface as it crosses the buried ridge indicated by the gravity data. Deep circulation may or may not occur in this model.

Brook and others (1979), using a reservoir temperature of 97°C and a volume of 6.1 km^3 obtained a reservoir thermal energy of 1.36 x 10^{18} J for their Abraham (Baker, Crater) Hot Springs system.They based the volume estimate primarily on the extent of the geophysical anomalies in the vicinity of the hot springs, and they assumed that the thermal water sampled in the spring had not been mixed with non-thermal ground water. This estimate is still valid if the assumption that the hot springs are the primary expression of the system is correct. The alternate model proposed here for the Fumarole Butte geothermal system suggests the possibility of a substantially larger reservoir volume, although meaningful quantitative estimates of the volume cannot be made with existing data. Also, if the thermal water has flowed a horizontal distance of several kilometers in the near surface, more mixing with non-thermal water is likely and thus the reservoir temperature may be significantly higher than the 97°C calculated by Brook and others.

The geological, geochemical, and geophysical data in and around the Fumarole Butte geothermal area do not provide any direct evidence of a large, very high-temperature geothermal system. However, a system with significantly higher energy content than that computed by Brook and others (1979) may exist. Thermal gradient-heat flow holes drilled to depths of about 100 m on Crater Bench west of Abraham Hot Springs would be a logical first step in evaluating the westward extent of the geothermal system.

MONROE-RED HILL AND JOSEPH GEOTHERMAL SYSTEMS

Three hot springs extending over a distance of approximately 10 km in the south end of Sevier Valley have calculated mean reservoir temperatures of 101° and 107°C (Brook and others, 1979). These springs (Monroe, Red Hill, and Joseph Hot Springs) were included in the Monroe-Joseph Known Geothermal Resource Area (KGRA). Brook and others considered Monroe-Red Hill Hot Springs as a single geothermal system and Joseph Hot Springs as a separate system. Rush (1983) considered all three springs under the general heading of the Monroe Joseph KGRA but recognized that Monroe-Red Hill and Joseph Hot Springs probably were parts of two separate systems. They appear to be two separate but similar geothermal systems and this study uses the names Monroe-Red Hill and Joseph geothermal systems.

Because these geothermal systems are in a more populated area than other systems in the Sevier thermal area, they have been of greater interest as a source of water for space heating. Under contracts from the Department of Energy, the faculty and students at the University of Utah Department of Geology and Geophysics have obtained geophysical data in the area of the Monroe-Red Hill and Joseph geothermal system (Mase and others, 1978; Halliday and Cook, 1978). The University of Utah Department of Geology and Geophysics and Terra Tek Inc. completed an evaluation of the Monroe-Red Hill geothermal system, supported by the Department of Energy, to determine if the resource would support a major space-heating development. This evaluation included geological, geophysical, and geochem-

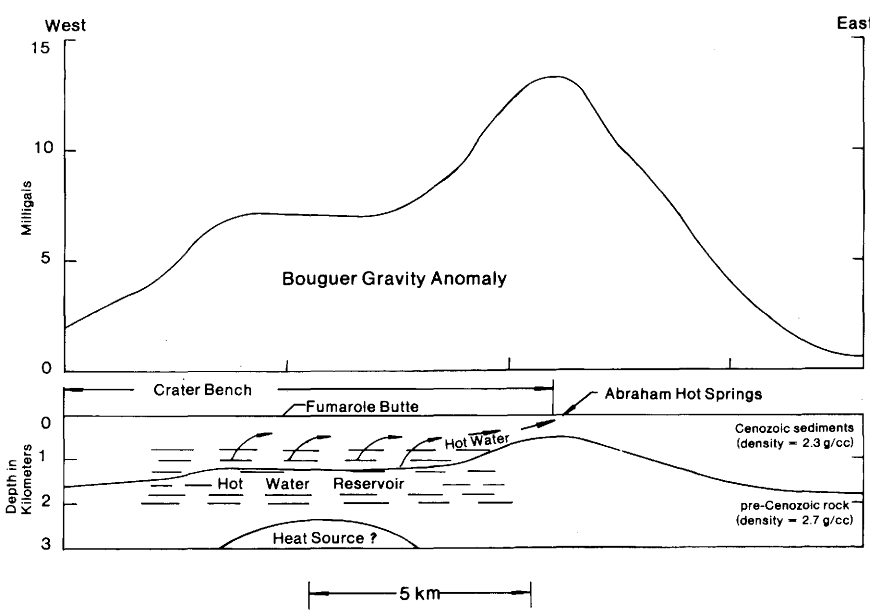


Figure 25. Residual Bouguer gravity anomaly profile and proposed geothermal model for Fumarole Butte and Abraham Hot Springs. Gravity data from Smith, 1974.

ical surveys and drilling of two test wells and one production well.

Both the Joseph and Monroe systems are in the area covered by a series of Miscellaneous Investigation Series maps of the Tushar Mountains and adjoining areas published by the USGS. Of particular importance to the geothermal studies are the geologic map (Cunningham and others, 1983), the gravity map (Cook and others, 1984), and the magnetic map (Campbell and others, 1984). Audiomagnetotelluric soundings were obtained as part of the evaluation of the KGRA by the USGS (Gardner, Williams, and Brougham, 1976).

The Monroe-Red Hill and Joseph geothermal systems are associated with Tertiary volcanic rocks. The hot springs lie near the contact between valley fill and mountainous volcanic terrain (figure 26). The volcanic rocks are Oligocene to Miocene age (30-19 Ma) basaltic andesite, andesite, rhyodacite, and quartz latite lava flows and volcanic breccia, and quartz latite and alkali rhyolite ash-flow tuffs which erupted from the Monroe Peak and Mount Belknap calderas (figure 3) and from sources to the west of the study area in the Needle Range (Cunningham and others, 1983).

The three hot springs discharge from travertine mounds along or near two northeast-trending normal faults. Joseph Hot Springs is near the Dry Wash fault, which parallels the Sevier River along the northwest edge of a group of hills that are an appendage of the Antelope Range. Monroe and Red Hill Hot Springs are in or near the Sevier fault zone, which parallels the western margin of the Sevier Plateau. In the vicinity of the hot springs, steep gravity gradients associated with both fault zones are interpreted as indicating a considerable thickness of low-density sedimentary rocks northwest of the fault zones. Tertiary volcanic rocks are the dominant rock on the upthrown side of the fault. Both hot springs systems lie near the north edge of the Marysvale-Pioche igneous belt as defined by the regional magnetic data (figure 7) and near the southeast edge of the Sevier thermal area.

Flow from the springs varies considerably with time. Rush (1983) made ten measurements at seven sites in Monroe-Red Hill Hot Springs and twelve sites at Joseph Hot Springs from March 16, 1976 to March 26, 1977. The total flow at Monroe-Red Hill ranged from 11.0 l/s on May 11, 1976 to a maximum of 21.1 l/s on January 1, 1977. At Joseph the total flow declined from 2.7 l/s for the first two measurements to 1.3 l/s for the final measurements. Water temperature variations up to 14°C were also measured during the period. The maximum temperatures Rush measured were 75°C at Monroe-Red Hill and 65°C at Joseph. Brook and others (1979) estimate the flows for Monroe-Red Hill at more than 20 l/s with a temperature of 76°C and Joseph at 1.7 l/s with a temperature of 59°C. Mase and others (1978) estimate a total discharge from Red Hill Hot Springs at 12.8 l/s and that at Monroe 5.8 l/s with discharge temperatures of 75°C for Red Hill and 70°C at Monroe.

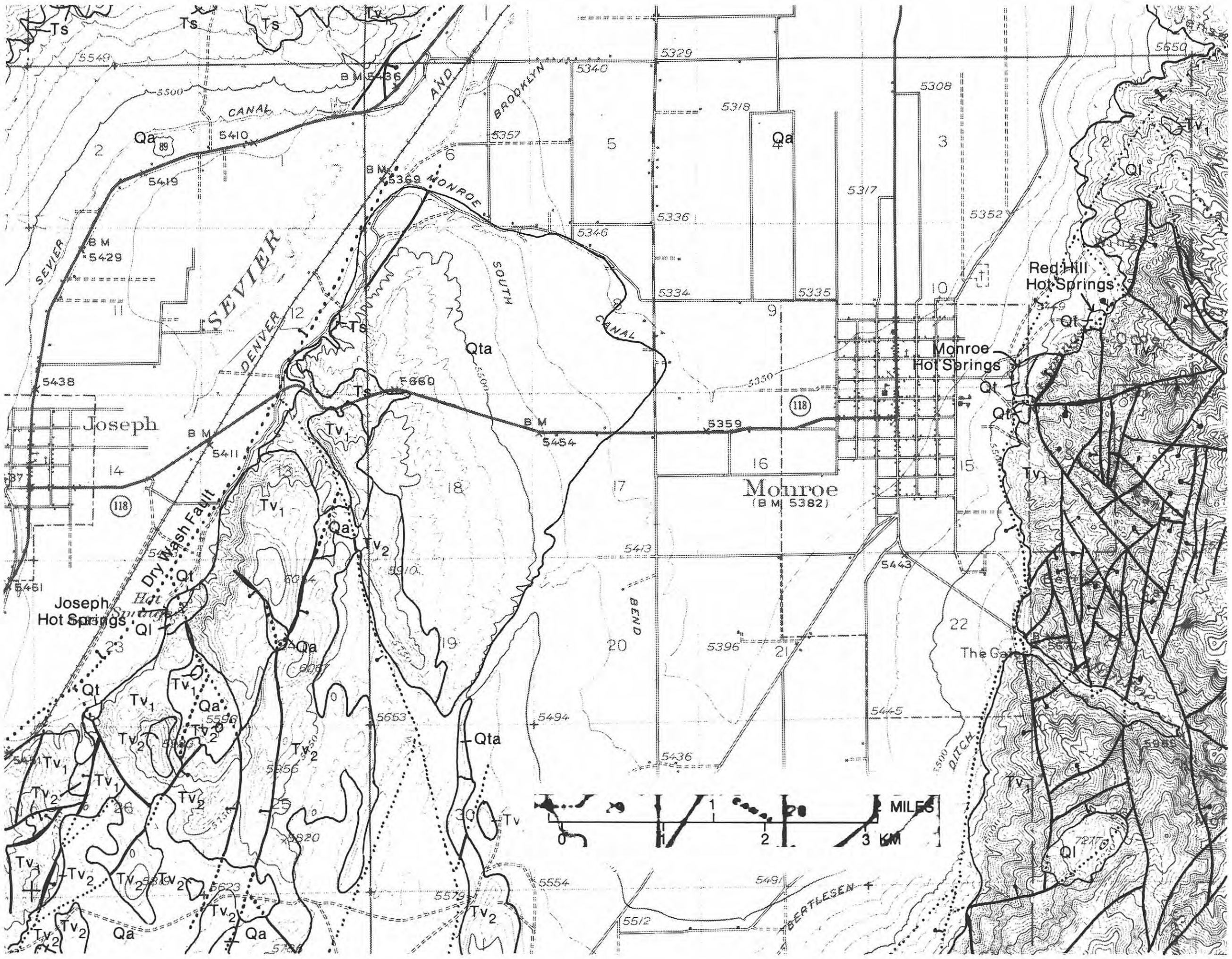
Rush (1983) assumed that the water discharged in the springs was about an equal mixture of thermal and non-thermal water and estimated the reservoir temperature as 100° to 160°C. Rush determined heat flow in 13 holes in the area ranging from less than 4 to 750 mW/m^2 in the immediate area of Joseph Hot Springs. He concluded that regional conductive heat flow was about 82 mW/m^2 and calculated that the water was probably heated to this temperature by circulating to depths of 2 to 4 km.

From approximately 6 km southwest of Joseph Hot Springs, the Sevier River flows northeast then north in an arcuate valley for a distance of about 100 km. In the southern part of the valley the topographic relief from the valley floor to the adjoining highlands is over 1500 m. A gravity low is approximately coincident with the valley for most of its extent (figure 6); however, the gravity anomaly is probably not caused entirely by low-density valley fill underlying the valley. The anomaly appears to be caused by a combination of (1) valley fill, (2) Tertiary sediments that are more extensive than the valley fill, and (3) Mesozoic sediments perhaps including salt structures involving the Jurassic Arapien Formation. Using gravity data alone it is not possible to isolate the effects of these sources; however, the gravity data provide a good indication of the location of, and some indication of the vertical displacement on, the larger faults.

The interpretation of the gravity data is further complicated by the large regional gravity gradient across Sevier Valley. A part of this gradient is produced by a thickening of the crust under the high topography of the Colorado Plateau relative to the Basin and Range Province to the west. However, the regional gradient is not a simple isostatic anomaly reflecting the difference in regional elevation across the valley. The Bouguer gravity anomaly over the Pavant Range west of the valley is significantly higher than normal for the regional elevations of the range.

Halliday and Cook (1978) have modeled gravity profiles across the faults near Joseph and Red Hill Hot Springs using the relatively detailed gravity data obtained in their survey (figure 27). Near Joseph Hot Springs they inferred a fault with a vertical throw of 800 m located about 500 m northwest of the location assumed for the Dry Wash fault and concluded, "it should be noted that the lack of an anomaly corresponding with the Dry Wash fault is explained as due to relatively little displacement along the fault." It appears that Halliday and Cook assumed the Dry Wash fault was at the contact between the volcanic rock and the alluvium. The geologic mapping of Cunningham and others (1983) does not support this conclusion. Their geologic map shows the Dry Wash fault, in the area of Joseph Hot Springs, as an inferred fault passing several hundred meters northwest of the springs. The gravity data suggest that the Dry Wash fault lies in the approximate position indicated on the geologic map and that it has about 1000 m of vertical displacement. Joseph Hot Springs is thus offset a few hundred meters southeast from the main trace of the Dry Wash fault. A normal fault that trends a few degrees west of north is mapped in the bedrock immediately south of the hot springs and is projected to pass through the area of the springs. The distribution of gravity stations does not define any anomaly that may be produced by the north-trending fault.

Halliday and Cook (1978) infer a fault in the immediate vicinity of Red Hill Hot Springs with a displacement of about



Qa	ALLUVIAL DEPOSITS (Quaternary)
Ql	LANDSLIDE DEBRIS (Quaternary)
Qt	TRAVERTINE (Quaternary)
Qta	OLDER ALLUVIUM (Pleistocene to Miocene?)
Ts	SEVIER RIVER FORMATION (Pliocene and Miocene) fluviatile and lacustrine sedimentary rocks
Tv_1	TERTIARY VOLCANICS (Miocene and Oligocene) - intermediate - composition assemblage, 30-22 Ma
Tv_2	TERTIARY VOLCANICS (Miocene) - rhyolitic composition, 19 Ma
	Fault-dotted where concealed, bar and ball on downthrown side

Figure 26. Geologic map of the Joseph-Monroe area (generalized from Cunningham and others, 1983).

300 m. The main trace of the Sevier fault is several hundred meters to the west. This interpretation is consistent with the geology as mapped by Cunningham and others (1983). A detailed gravity profile was not obtained across Monroe Hot Springs but the gravity data available suggest that the Sevier fault zone lies west of the springs.

The aeromagnetic map (figure 28) based on an aeromagnetic survey flown about 300 m above the surface (Campbell and others, 1984) shows that the Dry Wash and Sevier faults in the vicinity of the hot springs are generally coincident with magnetic gradients. At Joseph Hot Springs the magnetic anomaly pattern is similar to the gravity anomaly, with a high over the hills to the southeast and a low over the valley. In the area of Monroe and Red Hill Hot Springs the pattern is reversed with higher magnetic intensity over the valleys than over the topographic highs to the east. Detailed magnetic data (figure 29) obtained along ground profiles show magnetic lows coincident with part of the spring areas at Monroe and a strong magnetic gradient at Red Hill (Halliday and Cook, 1978).

Mase and others (1978) prepared a heat flow contour map of the Monroe-Red Hill area (figure 30). The map shows an extensive heat flow high, with more local heat flow highs, over the two hot spring systems. The extensive heat flow high and

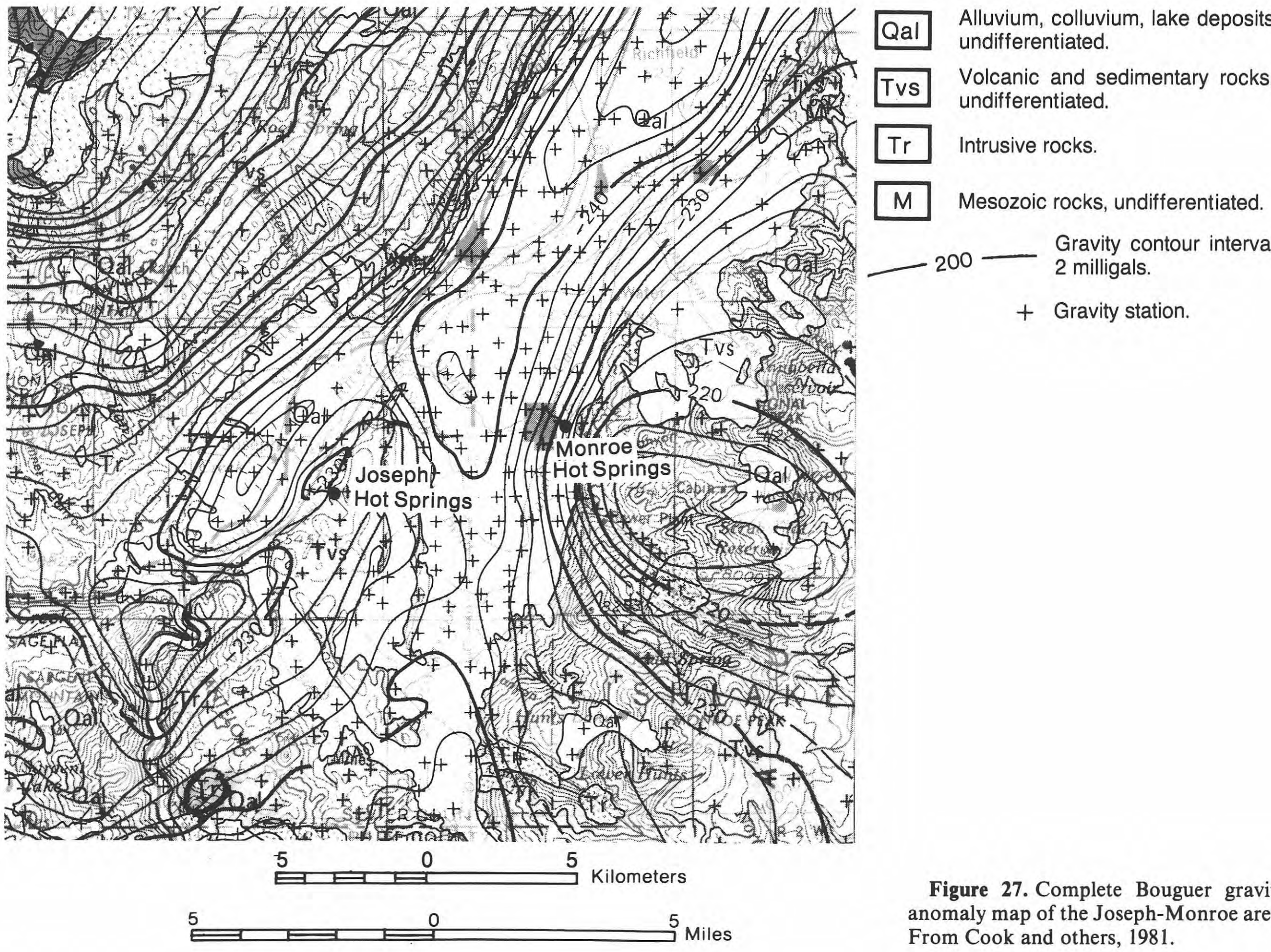


Figure 27. Complete Bouguer gravity anomaly map of the Joseph-Monroe area. From Cook and others, 1981.

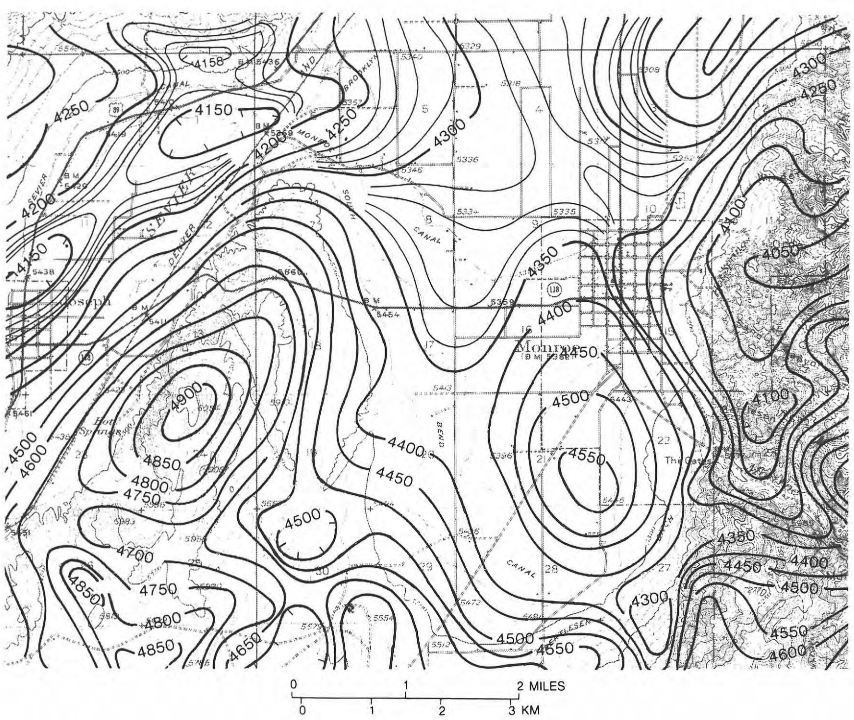


Figure 28. Residual aeromagnetic map of the Joseph-Monroe area. Contour interval is 10 and 50 gammas and values are relative to an arbitrary datum. Flight level was approximately 300 m above mean terrain. Flight paths are north-south approximately 0.8 km apart. From Campbell and others, 1984.

the two local highs are elongated approximately parallel to the Sevier fault zone. The map indicates that over an area of about 4.5 km^2 the near-surface heat flow is more than five times the regional average. They calculated that 2.9 MW (megawatts) of heat is lost from the Monroe-Red Hill system by conductive

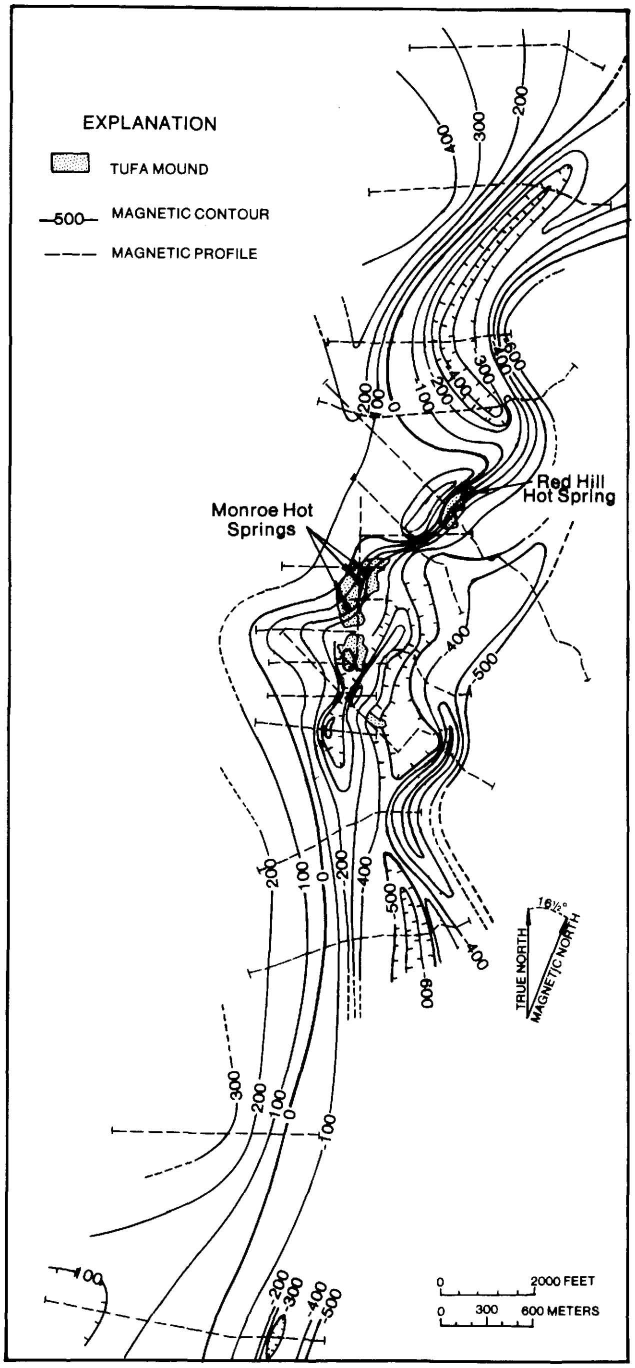


Figure 29. Total magnetic intensity based on surface magnetometer survey of the Monroe-Red Hill Hot Springs area. Contour interval 100 gammas; values relative to an arbitrary datum. Magnetometer stations were 65 and 164 feet (20 and 50 m) apart along profiles shown. From Halliday and Cook, 1978.

heat flow and that convective loss through the hot springs is 4.9 MW for a total of 7.8 MW of net heat discharge.

The dipole-dipole survey (figure 31) of the Monroe-Red Hill area defines a resistivity low approximately parallel to the Sevier fault zone but offset a few hundred meters to the east (Mase and others, 1978). The lowest resistivity values were measured in the immediate vicinity of the hot springs. At Monroe Hot Springs a near-vertical zone, about 75 m wide and with considerable depth extent, is modeled to have an apparent resistivity of 4 ohm-m superimposed on a zone about 500 m wide with a resistivity of 7 ohm-m (figure 32). At Red Hill Hot Spring a somewhat wider zone of 4 ohm-m resistivity is indicated with a 7 ohm-m zone confined to the northwest.

The audiomagnetotelluric survey (figure 33) that the USGS made of the Monroe Joseph KGRA consisted of 11 soundings distributed over an area of about 100 km^2 (Gardner, Williams, and Brougham, 1976). The soundings at Monroe Hot Springs indicated an apparent resistivity of about 4 ohm-m at all frequencies between 27 and 7.5 Hz (the lowest frequency measured). At Joseph Hot Spring the apparent resistivity increased from about 4 ohm-m at 27 Hz to about 10 ohm-m at 7.5 Hz suggesting an increase in resistivity with depth that is not apparent at Monroe Hot Springs. Elsewhere in the valley the apparent resistivities were greater than 20 ohm-m in the 27 to 7.5 Hz range.

The southern part of the Sevier Valley is one of the most seismically active areas in Utah. In 1901 one of the largest (perhaps the largest) earthquakes in Utah's historic record occurred in the area. The exact epicenter of the earthquake is not known but it caused considerable damage in Joseph and Monroe. Numerous other earthquakes have been felt and the seismicity map (figure 8) shows the concentration of earthquakes in this area. Nakata and others (1982) identify the Dry Wash fault and part of the Sevier fault zone as suspected Quaternary faults but do not identify evidence of late Quaternary movement.

The structural setting in the south end of the Sevier Valley is not typical of either the Basin and Range Province or the Colorado Plateau, and the character of the recent faulting is uncertain. It is possible that the normal faults that have been mapped or inferred along the margins of the valley do not have the listric style that is common to many active faults in western Utah. The Sevier Valley faults may be relatively high-angle faults that penetrate to considerable depth and thus provide better conduits for deep circulation of ground water than do listric faults that decrease in dip with depth. The data obtained from test drilling have been interpreted as indicating that the Sevier fault dips about 67 degrees to the west in the Monroe area (Mase and others, 1978). The geophysical data have been interpreted as indicating a somewhat steeper dip. Temperatures observed in the test holes indicate that convection at Monroe Hot Springs is confined to the fault zone.

The hot springs in the Joseph and Monroe-Red Hill geothermal systems occur near the intersection of normal faults, that displace Oligocene and Miocene volcanic rocks, with the Dry Wash and Sevier fault zones. The zones of intersection of

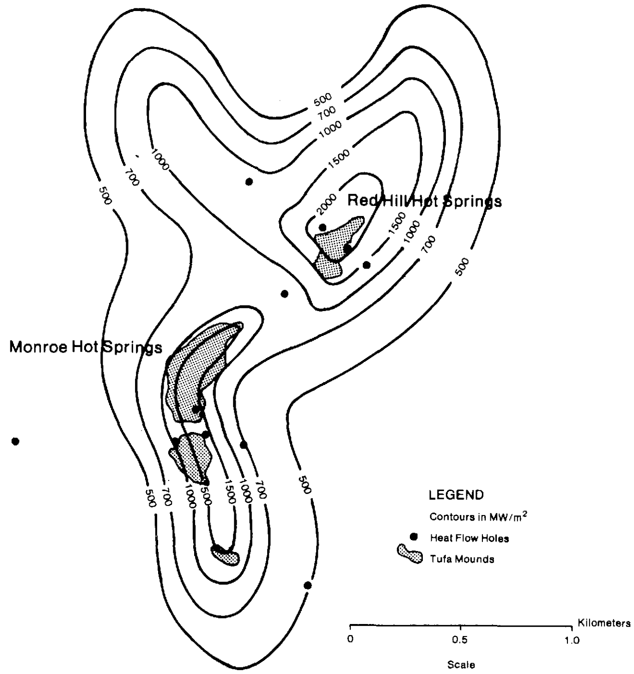


Figure 30. Heat flow map of the Monroe-Red Hill Hot Springs area. From Mase and others, 1978.

these two fault systems may contain fracture patterns that provide for increased convection within the geothermal systems.

The lack of any evidence of recent volcanic activity, suggesting a local heat source for the Joseph and Monroe-Red Hill geothermal systems, and the apparent association of the hot springs with active faults indicate that the systems are fault controlled. The calculated reservoir or estimated reservoir temperatures could be obtained by ground water circulating in the fault zones to depths of 2 to 4 km.

Water from the Monroe-Red Hill system is currently being used for bathing and swimming. The investigation of the feasibility of using the resource for a major space-heating project led to the conclusion that the project was not practical at that time (Hulen and Sandberg, 1981). The 457 m-deep production well flowed 17.7 l/s of 75°C water. Pumping 38 l/s for 30 hours produced a drawdown of about 110 m in the production well and significant drawdown in test wells within 76 m of the production well. During a 70-hour pumping test at about 17 l/s, flow in all the hot springs and seeps in the Monroe travertine mound declined and some stopped flowing. Flows at Red Hill were not affected.

The area of high, near-surface heat flow at the Joseph geothermal system extends for about 8 km but this likely reflects the distribution of water that has convected into the shallow subsurface. No drilling or geophysical data exist to indicate the extent of a reservoir of hot water within the system. Brook and others (1979) assigned the system the minimum volume of 3.3 km^3 they assumed for small systems. There is no basis to revise either the volume or reservoir temperature of 107°C assumed by Brook and others, or the estimate of total thermal energy of 0.83 x 10^{18}J computed by them.

All of the data available for the Joseph geothermal system are consistent with the hot water in the system being confined to fractured rock in or adjacent to the Dry Wash fault zone. A more extensive reservoir may exist at depth in fractured vol-

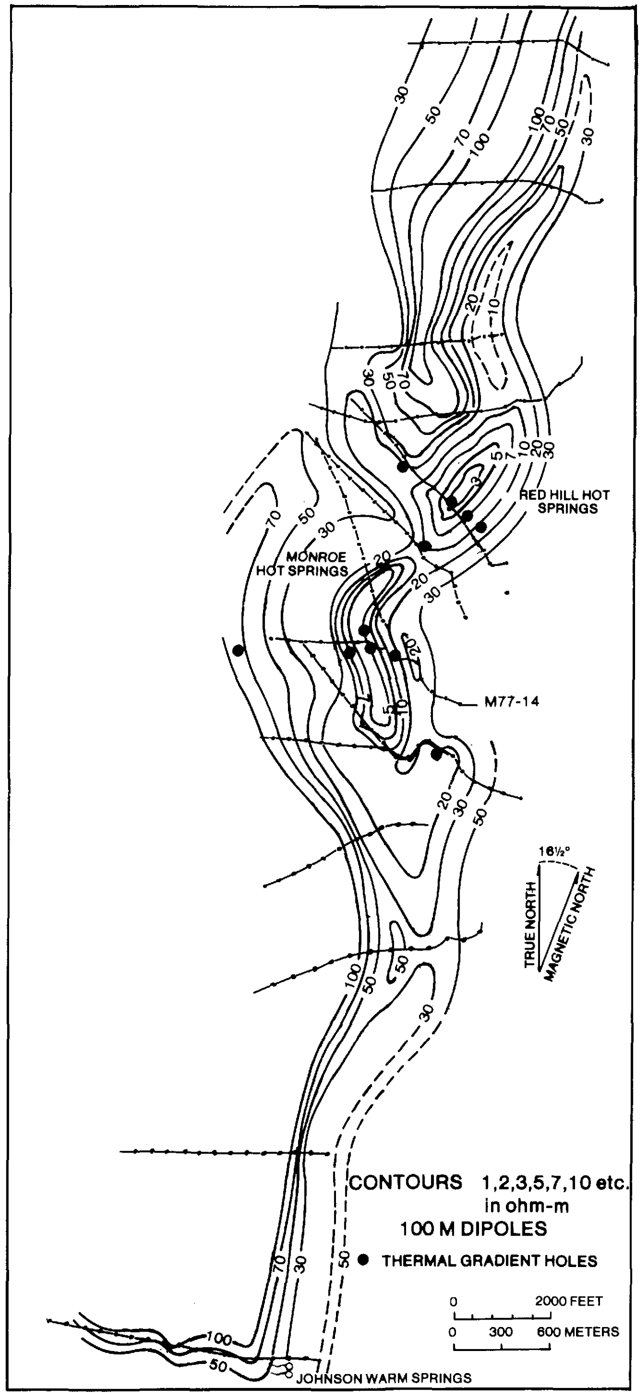


Figure 31. Dipole-dipole apparent resistivity map of the Monroe-Red Hill Hot Springs area. From Mase and others, 1978.

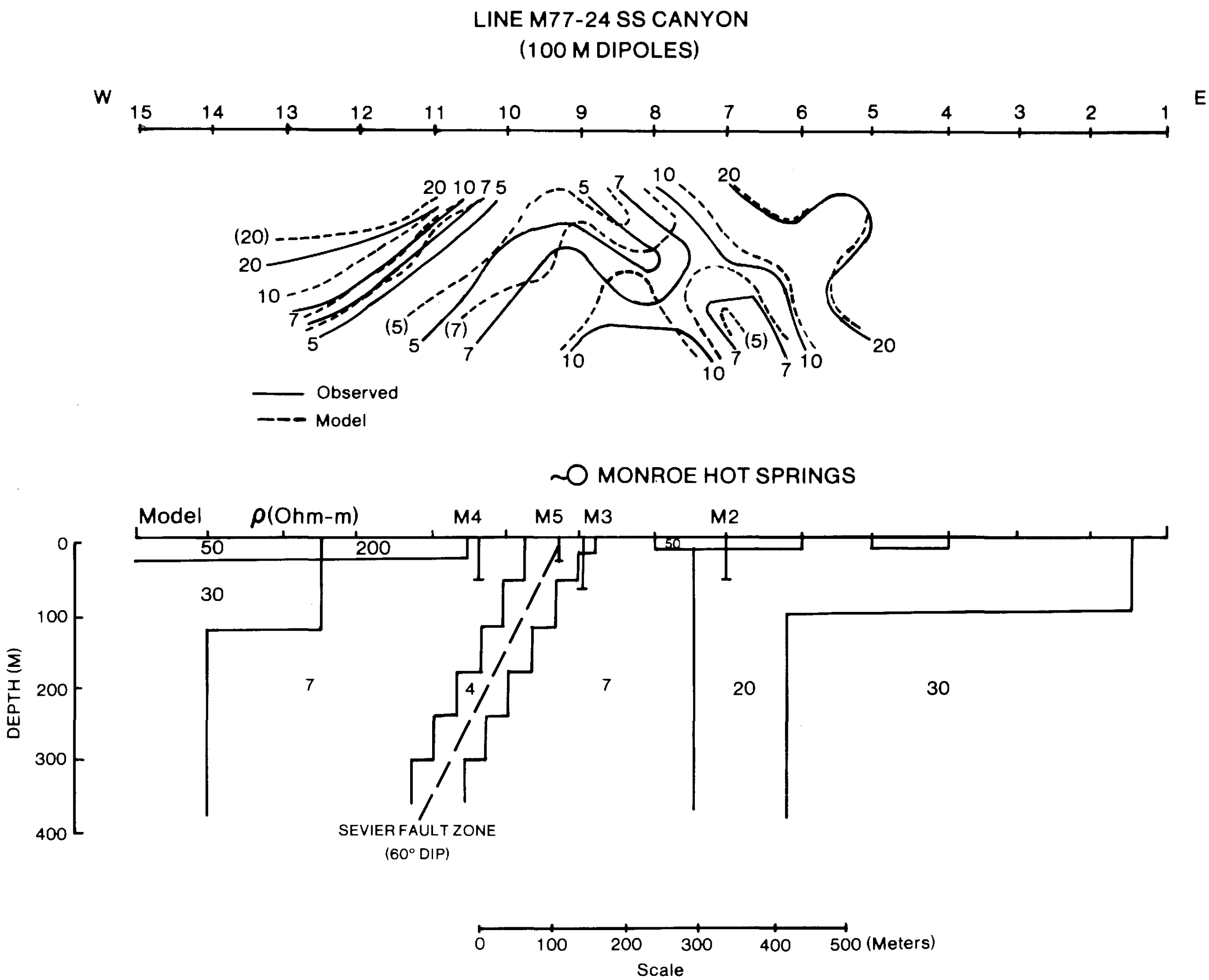


Figure 32. Electrical resistivity model through Monroe Hot Springs along line M77-14 (figure 31). From Mase and others, 1978.

canic rock or perhaps in older rock, but evidence of such a reservoir has not been reported. Based on the limited data that currently exist for this system it does not appear promising as an exploration target for high-temperature water.

The geophysical and temperature data indicate that the Monroe-Red Hill geothermal system extends along the Sevier fault zone for a distance of several kilometers and suggest a somewhat greater area for the reservoir than the 4.7 km^2 assumed by Brook and others (1979). Based on these data we increase the reservoir volume to 6.6 km^3. There are no new data that suggest revision of the 101°C mean reservoir temperature assumed by Brook and others. With the larger reservoir volume, the method of Brook and others (1979) yields an energy content of 1.53×10^{18}J for the system.

The subsurface data for the Monroe-Red Hill geothermal system indicate that the hot water is convecting into the near surface within the fractured rock of the fault zone. A more extensive reservoir at depth is likely to also be in fractured volcanic rock. The Monroe-Red Hill system is a promising source of moderate-temperature water up to temperatures of about 100°C, but the existing data do not indicate any great promise for the production of large volumes of higher temperature water.

COVE FORT GEOTHERMAL SYSTEM

The Cove Fort geothermal system, as used here, is the same as the Cove Fort-Sulphurdale system of Brook and others (1979) and is the second largest known geothermal system in Utah; however, some important features of the system are poorly understood. An area of about 130 km^2 was included in the Cove Fort-Sulphurdale KGRA. The surface manifestations of the system are sulfur deposits, altered ground, and gaseous emissions; no thermal springs have been reported. Several companies have conducted exploration in the area that included drilling three deep holes and several shallow or intermediate-depth holes. The area was included in an industry-

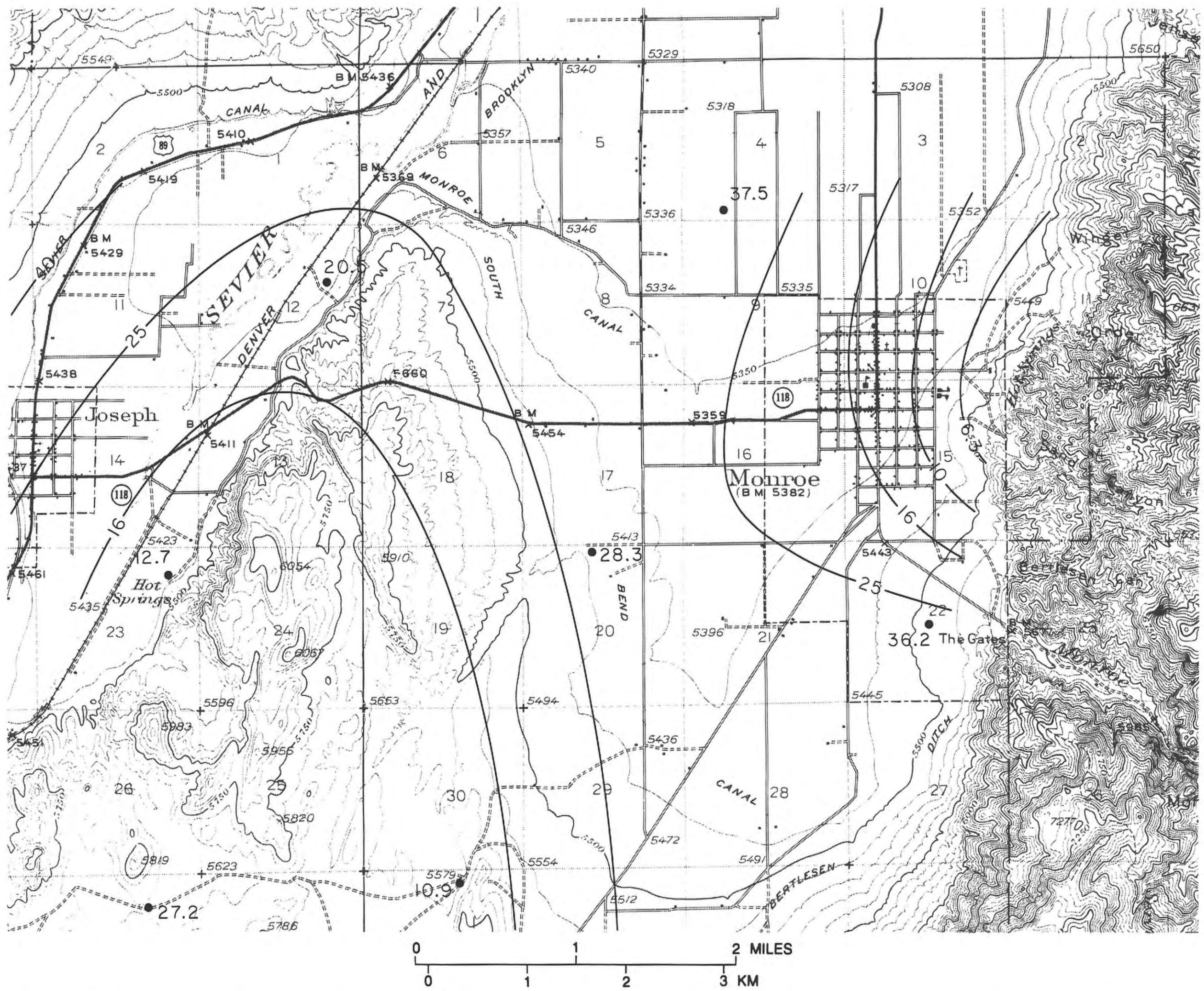
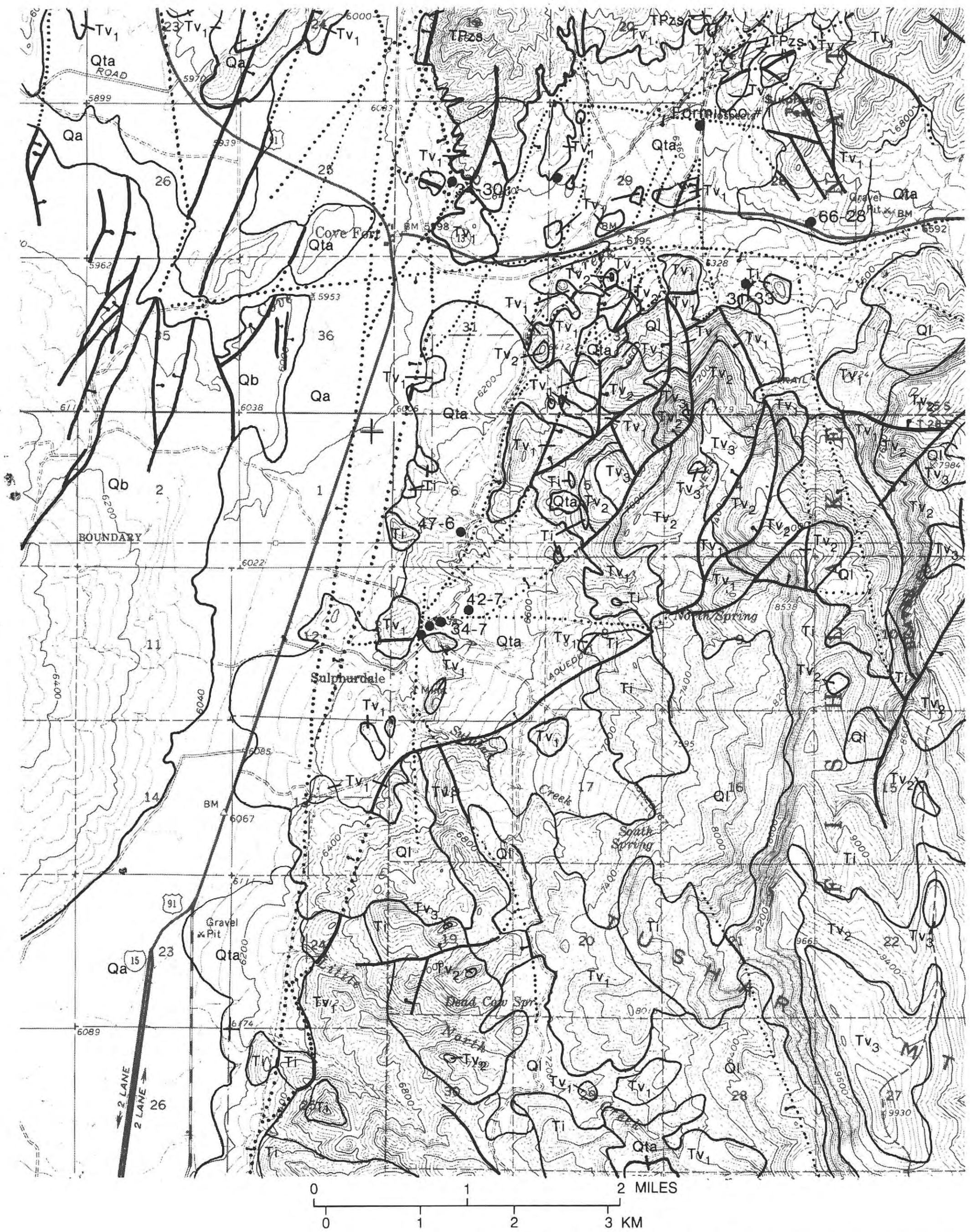


Figure 33. Apparent resistivity map of the Joseph-Monroe area from audiomagnetotelluric measurements at 7.5 Hz frequency. Values are ohm-meters obtained with north-south and east-west orientation of telluric line. Data from Gardner, Williams and Brougham (1976).

coupled project that resulted in considerable data, obtained by private industry, being made available to the public. In 1985, commercial production of electricity was begun by Mother Earth Industries Inc. with the electrical power being purchased by, the City of Provo.

The Cove Fort geothermal area lies between the Tushar Mountains on the south and the Pavant Range on the north (figure 34). The northern part of the Tushar Mountains is volcanic and intrusive igneous rock of Tertiary age. Eruptions from the Three Creeks, Monroe Peak, and Mount Belknap calderas (figure 3) produced the quartz latite and alkali rhyolite ash-flow tuffs (27 to 19 Ma) in the area. Shallow monzonite-latite plutons were emplaced sometime between 27 and 22 Ma and cut much of the ash-flow tuff in unit Tv_1. The Pavant Range is a block of pre-Tertiary sedimentary rock locally overlain on the south by Tertiary volcanic rocks and on the east by Tertiary sedimentary rocks. West of the thermal area is a large flow of Quaternary (Pleistocene) basaltic andesite. Much of the geothermal area is covered with alluvium, some perhaps as old as Miocene. Tertiary volcanic rocks cropping out in the southern part of the area appear to be in a large gravitational slide block. All of the deep drill holes penetrated pre-Tertiary sedimentary rocks and one drill hole (42-7) encountered quartz monzonite in four zones below a depth of 1,185 m (Moore and others, 1979). North- to northeast-trending normal faults are common. A few east-trending faults have been mapped which may be part of a major east-trending lineament approximately coincident with the south edge of the Sevier oval. The geothermal area is a few kilometers east of a zone of numerous north-trending Quaternary faults that extend north from Beaver Valley (Steven and Morris, 1983a).

The Energy Research and Development Administration (the predecessor of DOE) and Union Oil Company entered into a cost-sharing exploration program in the Cove Fort thermal area. Through this program the data from three deep exploration holes and other surface and subsurface data



- **Qa** ALLUVIAL DEPOSITS (Quaternary)
- **Q1** LANDSLIDE DEBRIS (Quaternary)
- **Qb** BASALTIC ANDESITE OF COVE FORT (Pleistocene)
- **Qta** OLDER ALLUVIUM (Pleistocene to Miocene?)
- **Tv_3** TERTIARY VOLCANICS, BIMODAL ASSEMBLAGE—(Miocene) alkali rhyolite ash-flow tuff, 19 Ma
- **Tv_2** TERTIARY VOLCANICS, INTERMEDIATE COMPOSITION (Miocene)—ash-flow tuff, 22 Ma
- **Ti** TERTIARY INTRUSIVE (Miocene?)—shallow monzonite porphyry plutons
- **Tv_1** TERTIARY VOLCANICS, INTERMEDIATE COMPOSITION (Miocene and Oligocene)—lava flows, flow breccia, and ash-flow tuff
- **TPzs** TERTIARY, MESOZOIC, AND PALEOZOIC SEDIMENTARY ROCKS
- Fault - Dotted where concealed. Bar and ball on downthrown side
- ● Geothermal Exploration Wells

Figure 34. Generalized geologic map and principal geothermal wells of the Cove Fort area. Generalized from Cunningham and others, 1983 and Steven and Morris, 1983a.

obtained by Union Oil Company were made available to the public. The results of this program are in reports by Union Oil Company (1973, 1978a, 1978b, 1978c, 1978d, 1978e, 1979a, 1979b). The deep wells were 799, 1,591 and 2,358 m deep. Maximum temperatures of 91°, 146°, and 178°C respectively were measured. Depth to the regional water table ranged from 409 to 427 m. In addition ESL/UURI has made geological, geophysical, and geochemical studies of the area, and Ross, Moore, and Christensen (1982) and Ross and Moore (1985) have reported case histories which include data obtained by both Union Oil Company and ESL/UURI.

In 1983 Mother Earth Industries encountered steam in a relatively shallow drill hole about 1 km northeast of Sulphurdale. Subsequently, two production wells have been completed in the immediate area and the steam produced is being used to power a 3.2 MW electrical generating system. When the wells were first vented the steam was mixed with a considerable amount of noncombustible gas that declined as the wells produced. Only a limited amount of information on these wells has been made public. Apparently the steam reservoir is at a depth of 350 m. Ross and Moore (1985) report a well-head temperature of about 200°C but we cannot confirm this temperature. The extent of the reservoir has not been determined. Mother Earth Industries drilled three additional holes outside of the immediate area of steam production; however, no information regarding drilling has been released. Analyses of gas samples taken from the two accessible drill holes (66-28 and 47-6, figure 34) are presented in table 3. Sample 5 was collected at a well head where a temperature of 32°C was measured; carbon dioxide, sulfur, ammonia, nitrogen, and hydrogen were detected. Samples 6 and 7 were collected from a pipe venting into the atmosphere and gave a temperature of 32°C with the well-head thermometer measuring 30°C. Carbon dioxide, ammonia, and carbon monoxide were detected in these samples. The radon concentration varies between the two wells, but studies show that radon concentrations in geofluids from geothermal reservoirs vary significantly with time (Stoker and Kruger, 1975).

The Cove Fort geothermal area lies in a zone with a steep regional gravity gradient between the high on the east side of the Sevier Desert and a gravity low over the Tushar Mountains (figure 6). Superimposed on this gradient are local anomalies present in the geothermal area (figure 35). A gravity low, produced in part by sedimentary and volcanic rocks underlying a northward extension of Beaver Valley, extends along the west side of the geothermal area. Steep, local, linear gradients along the east side of this low have been interpreted by Ross and Moore (1985) as indicating faults that may be the principal conduits that tap a deep geothermal reservoir. A gravity high extending south, from outcrops of Paleozoic rock northeast of Cove Fort, suggests that a buried ridge of Paleozoic rock lies near the east edge of the geothermal area. Ross and Moore have used the gravity data to infer several faults in the geothermal area.

A regional magnetic low extends along the north edge of the Marysvale-Pioche magnetic zone (figure 7). This low is generally assumed to be a polarization low as would be expected along the north edge of a mass magnetized in the direction of the earth's magnetic field. The Cove Fort geothermal area lies

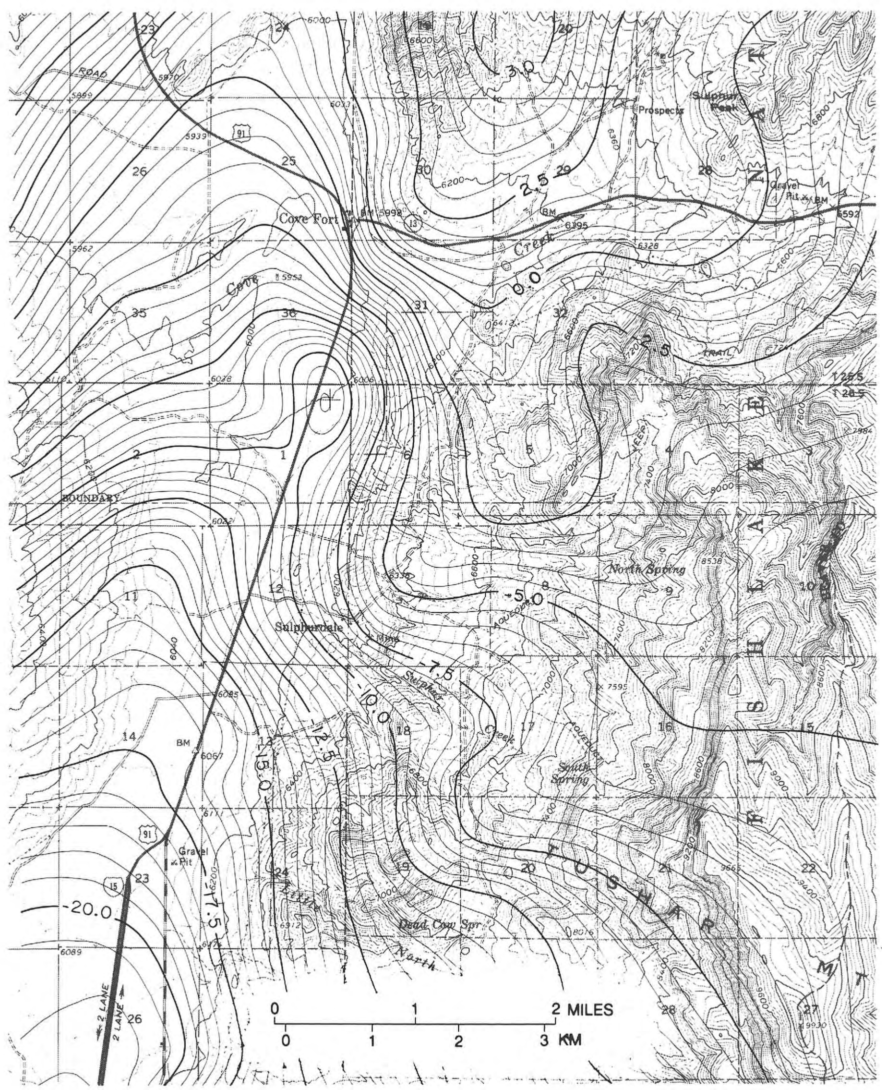


Figure 35. Residual Bouguer gravity anomaly map of the Cove Fort area. Contour interval is 0.5 mgal. Density assumed in Bouguer correction is 2.4 g/cc. From Union Oil Company, 1978a.

within this magnetic low and the area of lowest intensity, for a distance of about 40 km, is coincident with the geothermal area. D. L. Campbell (personal communication, 1985) has suggested that this very low intensity may reflect an underlying pluton that is magnetized in a direction approximately opposite to the current geomagnetic field. We prefer the interpretation that the low intensity is almost entirely a polarization low produced by the strongly magnetized pluton to the south.

The detailed magnetic survey defines a low of about 20 gammas in the Cove Fort thermal area (figure 36). The cause of this low is not known but could be produced by alteration of the volcanic rocks, by variations in the thickness or magnetization of the volcanic rocks, or by structure in the volcanic rocks.

The thermal gradient map (figure 37) of the Cove Fort thermal area is based on widely-spaced shallow drill holes. The thermal data can be interpreted as indicating an extensive thermal anomaly with gradients of over 300°C/km extending from about 3 km north of Cove Fort to Sulphurdale, or as an anomaly northeast of Cove Fort and another anomaly at Sulphurdale. Ross and Moore (1985) report that the normal thermal gradient in Cenozoic rocks of the type penetrated by these drill holes is about 50°C/km. Thus the thermal gradient over a large area is well above normal. They also report that this is part of a 200 to 300 km^2 anomalous thermal area that extends north into Dog Valley.

Dipole-dipole resistivity profiles in the area of the Cove Fort thermal anomaly (Ross, 1979) did not define any deep resistivity anomaly clearly associated with the geothermal system. Several areas of low resistivity were interpreted as being caused by zones of hydrothermal alteration; several north-trending zones of low resistivity were interpreted as defining zones of upward migration of thermal fluids along normal faults. Ross concluded that high resistivities at depths exceeding 610 m were not encouraging for the presence of a deep geothermal reservoir. Drilling subsequent to the work reported by Ross has revealed that part of the reservoir in the Cove Fort geothermal system contains steam rather than hot water. The resistivity of a steam reservoir could be substantially higher than a reservoir containing hot brine.

Ross, Moore, and Christensen, (1982) report the results of a geochemical survey based on cuttings from shallow thermal gradient holes. Anomalous concentrations of As, Hg, Pb, and Zn occurred in some of the holes but the distribution of these elements does not appear to define an anomaly that indicates the position of the geothermal system at depth.

A telluric current survey (figure 38) (O'Donnell and Stanley, 1987), consisting of 13 stations and a base station, reveals a deep resistivity low, but the form of the anomaly is not well defined by the few stations. Major uncertainties are inherent in the interpretation of telluric current data, particularly in areas of complex geology and high surface relief as exist at Cove Fort. However, the telluric current data are consistent with the suggestion of Ross, Moore, and Christensen, (1982) and Ross and Moore (1985) that the area of most intensive exploration to date may be on the periphery of a large convective system located to the northwest.

More earthquake activity has been recorded at Cove Fort than any other thermal area studied in Utah. Cove Fort lies near the northwest edge of the most active part of a segment of the Intermountain seismic belt that trends southwest across southwest Utah. A concentration of epicenters has been located in and around the Cove Fort area. A detailed earthquake study of the Cove Fort geothermal area (Olson

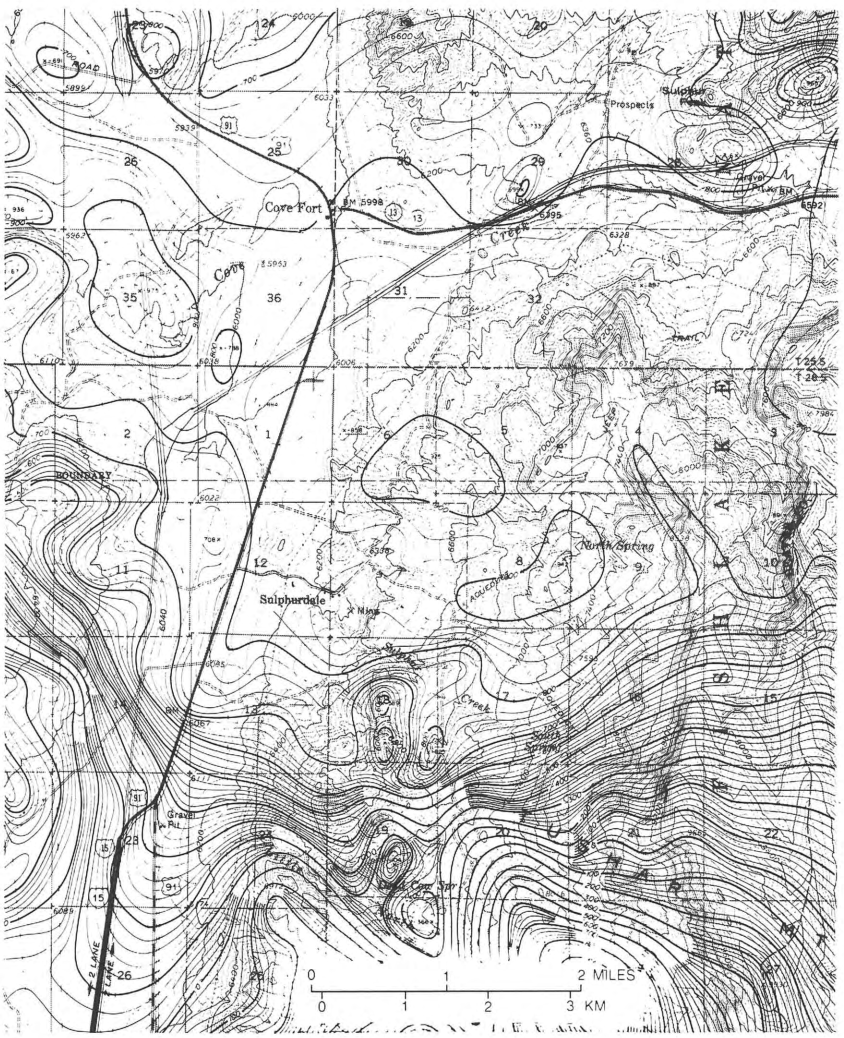


Figure 36. Residual aeromagnetic map of the Cove Fort area. Contour interval is 20 to 100 gammas. Flight lines are north-south, 1640 feet (0.5 km) apart and 1,000 feet (305 m) above mean terrain. (From Earth Science Labratory, 1978).

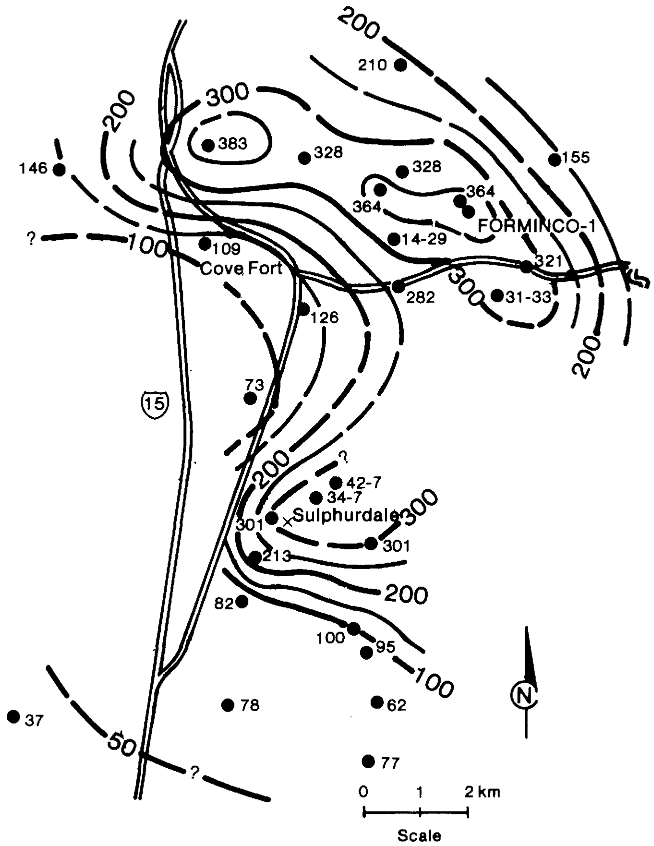


Figure 37. Thermal gradient map of the Cove Fort thermal area. Values are °C/km in the interval 98 to 250 feet (30 to 76 m) below the surface. From Union Oil Company, 1978a.

and Smith, 1976) revealed that 75 percent of events recorded during the monitoring period had focal depths of less than 5 km, with the shallowest events clustered northeast of Cove Fort and the deeper events located farther north in the area of Dog Valley. Olson and Smith conclude that the shallow events may be related to a north-trending fault dipping about 70 degrees west. Fault-plane solutions for events in the Cove Fort area indicate primarily normal faulting with east-west extension.

The Cove Fort geothermal system is not well understood despite considerable drill hole and geophysical data. The data suggest a complex system probably consisting of a deeper water-dominated system and one or more vapor-dominated systems at shallow depths. In the area northeast of Sulphurdale where steam is being produced, a cap over the shallow parts of the geothermal system appears to have formed from a large gravity slide block off the northwest part of the Tushar Mountains. Northwest-trending normal faults that are currently active appear to control the conduits through which thermal fluids move into the shallower parts of the system and may partially control the deeper parts of the geothermal system. Several investigators have proposed that the heat source for the Cove Fort geothermal system is directly related to the Quaternary basalt flows to the west. As pointed out by Moore and others (1979) the distribution of young hydrothermal alteration and the current seismicity cannot be definitely related to the basaltic volcanism. No evidence has been presented that indicates the basalt is derived from a magma chamber in the upper crust that could provide heat to the geothermal system. Existing information is compatible with the heat being derived from deep circulation of water in a complex active structural zone; however, the possibility of a local heat source in the area should not be ignored.

The production of steam from the Cove Fort geothermal system is from two wells about 350 m deep and 90 m apart. A third well that first encountered this vapor-dominated portion of the system was about the same depth and very near the production wells. The drilling has not been directed toward determining the limits of this resource and there are no drill holes in the area that indicate the extent of this part of the geothermal system. Based on unpublished flow testing of the production wells, Geothermal Resources Council Bulletin (1985) reported "MEI's (Mother Earth Industries) reservoir engineers, Therma Source, Inc., have prepared an economic evaluation which details the proven field as 25 MW, with the possibility of 125 MW of potential. The lifetime expectation of the field is 20 years minimum."

The resource being produced by the two production wells can best be defined by a combination of geophysical surveys and test drilling. A detailed temperature gradient and/or heat flow survey would likely indicate the extent of the near-surface parts of this reservoir. Because the structure is likely to be complex with extensive zones of alteration and the reservoir partly vapor dominated, the application of electrical methods to explore the deeper parts of the system may be limited. A refraction seismic survey should be able to map the interfaces between (1) alluvium, (2) displaced Tertiary volcanic rock, (3) in-place Tertiary volcanic rock, and (4) pre-Tertiary rock. The survey might also indicate the structural control on the deeper part of the system. A few additional gravity stations are needed to supplement the existing gravity data to provide a more complete definition of the gravity field. The refraction seismic data and upgraded gravity survey, when combined with the existing geophysical information and the information from the production wells, could be used to construct a model of the deeper part of the geothermal system that could be tested by drilling.

Evaluating the total resource in the Cove Fort system is a complex problem. The existing geological and geophysical data can be used to develop a regional geologic model of the subsurface, which can be tested and refined by more detailed geophysical and perhaps geochemical surveys. Deep drilling will ultimately be required to adequately define the resource. We propose a preliminary model that includes the following major elements: (1) an east-west regional structure of undetermined width but with the south edge about 2 km south of Sulphurdale; (2) a north-trending fault zone through Cove Fort with the Sulphurdale fault a key member; (3) a pluton lying southeast of (1) and (2) of undetermined age, but possibly significantly younger than the volcanic rocks of the Tushar Mountains, that may be related to a north-trending zone of active volcanism; and (4) a possible very young intrusive mass centered a kilometer west of Cove Fort.

The east-west regional feature through the Cove Fort area is probably a very deep-seated structure that is expressed in the near-surface geology in several ways. To the east, the Clear Creek downwarp trends west into the Cove Fort thermal area. The southern edge of the Sevier oval passes through the Cove Fort area, and the northern edge of the Marysvale-Pioche volcanic belt, as defined by the magnetic anomalies, lies immediately south of the area. A westward offset of the area of most intense earthquake activity in the Intermountain seismic belt occurs in this general location. How these west-trending features interrelate is not known nor is it clear what the fundamental nature of the regional structure is.

Most young north-trending faults in this region, including those through the Cove Fort thermal area, are normal faults related to east-west extension. Where dips on these faults have been determined, most are to the west. To the north these basin and range faults often have listric form. In the absence of any direct measurement the westward dip of 70 degrees indicated by the shallow earthquakes might be assumed, but the possibility of lower dips particularly at depth should be considered.

The primary evidence for a pluton southeast of the intersection of the east-west feature and the north-trending fault zone is a prominent magnetic high. Several outcrops of Miocene intrusive rock occur in the area of this high, but the distribution of these outcrops relative to the magnetic high does not indicate that the rocks at the surface are the primary cause of the magnetic anomaly. Because igneous rock covers the entire area of the magnetic high, any detailed analysis of the anomaly is subject to large uncertainties. The major features of the anomaly suggest the top of the main magnetic mass is within 2 km of the surface. The possibility that this pluton is younger than the Miocene rocks at the surface should be considered.

The evidence of a young intrusive mass west of Cove Fort is weak. This area is within a zone of active volcanism which extends from the central Mineral Mountains to Pavant Butte. A telluric current low is evidence of a conductive body but provides little control on the size or vertical position of the mass. Poorly defined magnetic and gravity lows of low amplitude are approximately coincident with the telluric current low. The shallow thermal anomaly is offset to the east but is poorly defined in the key area.

The size and temperature of the geothermal resource in the Cove Fort geothermal system are a major uncertainty. Considerable information has become available since Brook and others (1979) made their assessment of the Cove Fort system. This additional information has provided further evidence that the system is complex but does not provide a basis to revise either the calculated mean reservoir temperature of 167°C, mean reservoir volume of 39 km^3, or the mean reservoir thermal energy of 16.0 x 10^{18}J. However, the resource in the Cove Fort system may be both larger and higher in temperature than these values indicate.

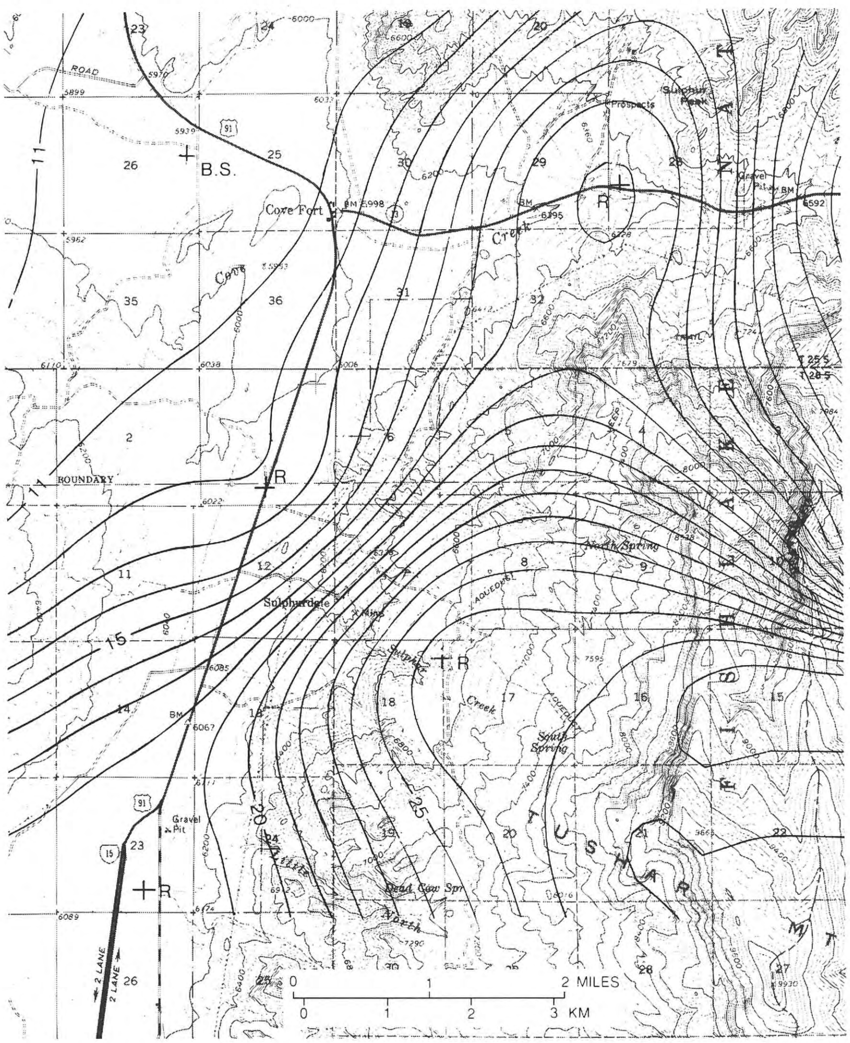


Figure 38. Telluric current map of the Cove Fort area. Contours are K values which are proportional to the ratio of the conductance of the base station (labeled BS) to the conductance of the rover stations (labeled R) (lower values indicate lower resistivity). From O'Donnell and Stanley, 1987.

ROOSEVELT GEOTHERMAL SYSTEM

The Roosevelt geothermal system is by far the most explored geothermal system in Utah and appears to be substantially larger and hotter than other known systems in the state. Water produced from the system is being used to generate about 20 MW of electric power and the system appears capable of much greater production. Ward and others (1978) and Ross, Nielson, and Moore (1982) have summarized the geological, geochemical, and geophysical data relating to the

Roosevelt geothermal system. In this report their conclusions are accepted and all of the detailed data they analyzed are not presented but the regional setting of the system is explored.

The Roosevelt geothermal system is on the east side of Milford Valley at the front of the Mineral Mountains (figure 39). Milford Valley is a major north-trending valley that, in topographic form, is typical of the basin and range valleys of western Utah. The Mineral Mountains are similar to numerous other ranges in the region in size and form but the geology is unusual. Most of the range is composed of a composite Tertiary batholith in an uplifted structural block. Sibbett and Nielson (1980) describe the block as an igneous diapir. The granitic batholith, which is the largest exposed in Utah (Nielson and others, 1978), consists of numerous plutons that are equivalent to volcanic rocks of the Mount Belknap volcanics in the Tushar Mountains to the east (Steven and Morris, 1984). The older, small hornblende quartz monzonite and hornblende granodiorite plutons were emplaced during the calc-alkalic igneous activity in the Tushar Mountains. Metamorphic rocks, 1.7 Ga, are exposed in a band along the west side of the Mineral Mountains (Aleinikoff and others, in press). Paleozoic and Mesozoic sedimentary rocks occur at the

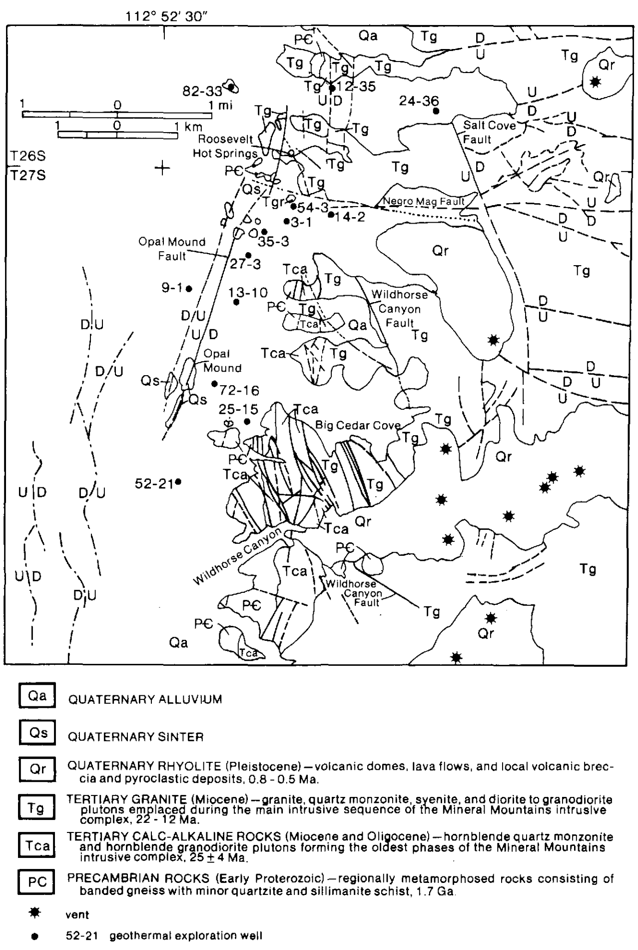


Figure 39. Geologic map of the Roosevelt geothermal area. Generalized from Nielson and others, in press.

north and south ends of the range. In the central part of the range is a series of rhyolite domes and flows about 500,000 years old (Lipman and others, 1978).

The surface expression of the Roosevelt geothermal system is a zone of opal mounds, cemented alluvium, and Roosevelt Hot Spring. Flow from the spring in historic time has been low, generally less than 1 l/s. Water temperatures of 88° and 85°C were measured in 1908 and 1950, respectively, but in 1966 the spring was dry (Mundorff, 1970). The high silica content of the spring waters focused attention on this system as a probable high-temperature system.

Although Roosevelt Hot Spring occurs near the east edge of the Milford Valley, the springs and the related zone of mounds and cemented alluvium are not controlled by a normal fault zone at the edge of the thick prism of valley fill. In the area of the thermal system, bedrock consisting of Tertiary granitic rock and older metamorphic rock is within a few hundred meters of the surface and the Opal Mound fault, which is interpreted as being the western boundary of a fracture zone about 500 m wide that is a major control of the system, is inferred to be a steeply dipping normal fault down to the east. West of the geothermal system the valley fill thickens over a distance of about 5 km probably by a combination of step faults and a dipping bedrock surface.

The regional gravity anomalies (figure 6) in the area of the Roosevelt geothermal system are not typical of those in the Basin and Range Province. A major gravity low occurs over the central part of Milford Valley. It has been interpreted as indicating up to 3,000 m of low-density sediments underlying the central part of the valley. The lowest anomaly values and presumably the thickest low density sediments are approximately opposite the geothermal system. The gravity high to the east is not centered over the Mineral Mountains but lies near the mountain front with the axis of the high approximately coincident with the surface expression of the geothermal system and outcropping pre-Tertiary rock to the north and south. The Bouguer anomaly values increase north and south from a low in the central part of the range a few kilometers south of the geothermal system. Quaternary and Tertiary rhyolite domes are present in the area of the gravity low. Three possible causes for all or part of the gravity low in the range should be considered: (1) the root zone of the batholith, (2) a lower density phase of the batholith and, (3) a zone of molten or partly molten rock at depth.

A regional magnetic high (figure 7) is produced by the Mineral Mountains batholith. This high is part of the Marysvale-Pioche magnetic zone and is consistent with the Mineral Mountains batholith being an uplifted segment of a more regional batholith. The magnetic high is coincident with the range and, in part, reflects the surface relief of the exposed batholith. The magnetic intensity is highest in the northern and southern parts of the Mineral Mountains. The relatively low area in the middle is south of the area of low Bouguer gravity anomaly values. This magnetic low likely reflects a zone of the batholith with lower magnetic susceptibility.

Detailed gravity and magnetic surveys of the Roosevelt geothermal area are reported by Ward and others (1978). They found the detailed gravity data useful in modeling some of the detailed structure and the magnetic data useful in the study of areal geology.

Geoelectric surveys have defined a resistivity low approximately coincident with the Roosevelt geothermal area. Resistivities of 5 ohm-m or less are produced by the fractured rock containing hot water. Saturated alluvium and altered rock has a resistivity of 5 to 12 ohm-m. Two-dimensional modeling of the geoelectric sections suggests a steeply dipping zone of fractured rock about 300 to 400 m wide in the first few hundred meters below the surface.

Self-potential anomalies associated with parts of the Roosevelt geothermal system have been reported by Corwin and Hoover (1979) and Sill and Johng (1979). Ross, Nielson, and Moore (1982) concluded that a 100 millivolt low over the Opal Mound

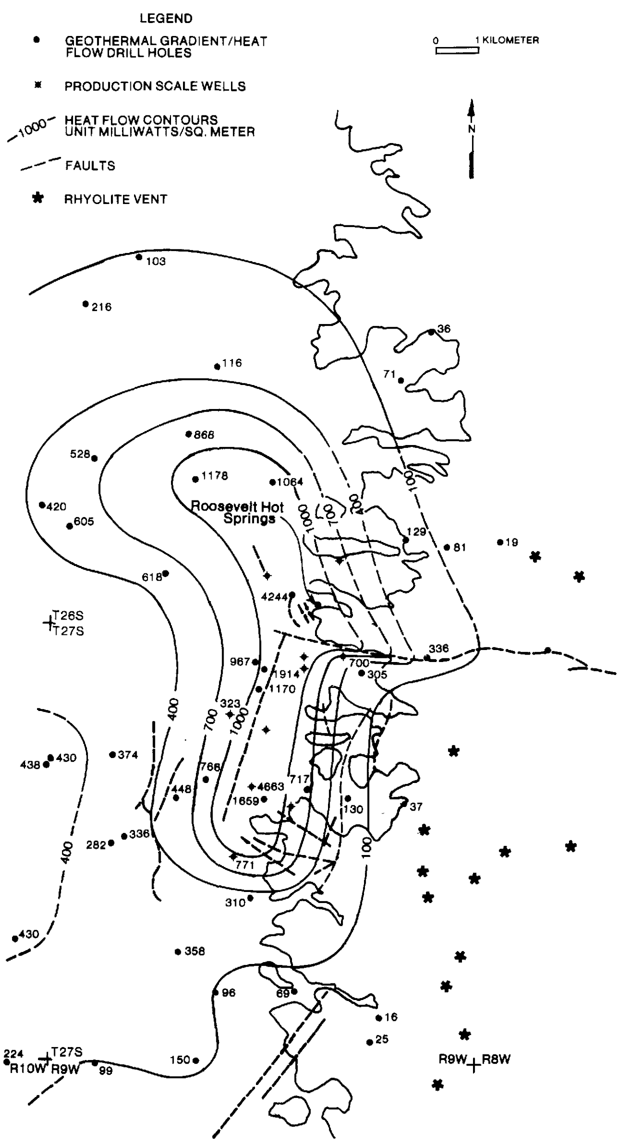


Figure 40. Conductive heat flow map of the Roosevelt geothermal area. From Wilson and Chapman, 1980.

fault was the most unambiguous expression of the geothermal system.

A heat flow high about 5 km wide and 20 km long occurs over the Roosevelt geothermal area (figure 40). In an area about 2 km wide and 8 km long heat flow values in excess of 1,000 mW/m^2, or about ten times the regional normal, were calculated. This area of highest heat flow is approximately coincident with the near-surface parts of the geothermal system. The heat flow map appears to be a good indicator of the extent of the geothermal system.

The Roosevelt geothermal system lies along the northwest edge of the band of more intense seismicity in the Intermountain seismic belt. There is no concentration of earthquakes in the vicinity of the geothermal area but a system of north-trending Quaternary faults extends through the area. Based on a microearthquake survey of the area involving 49 days of recording, Ward and others (1978) concluded: "These earthquake data suggest that the Roosevelt Hot Springs thermal area is only slightly active and at a much lower level than the earthquake zones 60 km east near Richfield and Marysvale, Utah." They also report P-wave details for raypaths beneath the Mineral Mountains show a delay that might be caused by a low-velocity zone related to a partially molten mass. A seismic refraction profile across the Roosevelt geothermal area (Gertson and Smith, 1979) indicated that both the depth to bedrock and the easternmost basin-bounding fault with large displacement were at least 1 km west of the Opal Mound fault. A seismic reflection profile (Ross, Nielson, and Moore, 1982) across the area defined numerous reflecting horizons in the basin fill under Milford Valley and fewer, generally discontinuous, reflectors within the older rocks. The reflection data suggest that the north-trending faults both east and west of the Opal Mound dip to the west.

Seismic emission (seismic noise) surveys have been attempted in the Roosevelt geothermal area (Katz, 1977). Ross, Nielson, and Moore (1982) concluded that while these surveys may indicate areas of geothermally induced noise they are imprecise in defining geothermal conduits. They further concluded that it was unlikely that the usefulness of the method justified inclusion in the exploration of a geothermal system in a complex geologic setting.

Regional studies of ground-water geochemistry have revealed anomalous concentrations of total dissolved solids, B, and Cl (figure 41) associated with the Roosevelt geothermal system (Ross, Nielson, and Moore, 1982). These anomalies probably reflect leakage from the geothermal system into the shallow aquifer. Ross, Nielson, and Moore conclude that Hg and As anomalies in the soils are associated with hot spring deposits and faults related to the geothermal reservoir.

Production from the Roosevelt geothermal system is from fractured Tertiary granite and older metamorphic rocks. The fracturing that produced the reservoir is probably controlled by the intersection of a system of north-trending Quaternary normal faults with more westerly-trending older faults that do not show evidence of Quaternary movement. Isotopic studies of the thermal water indicate that it could have originated as precipitation in the ranges to the east (Ross, Nielson, and

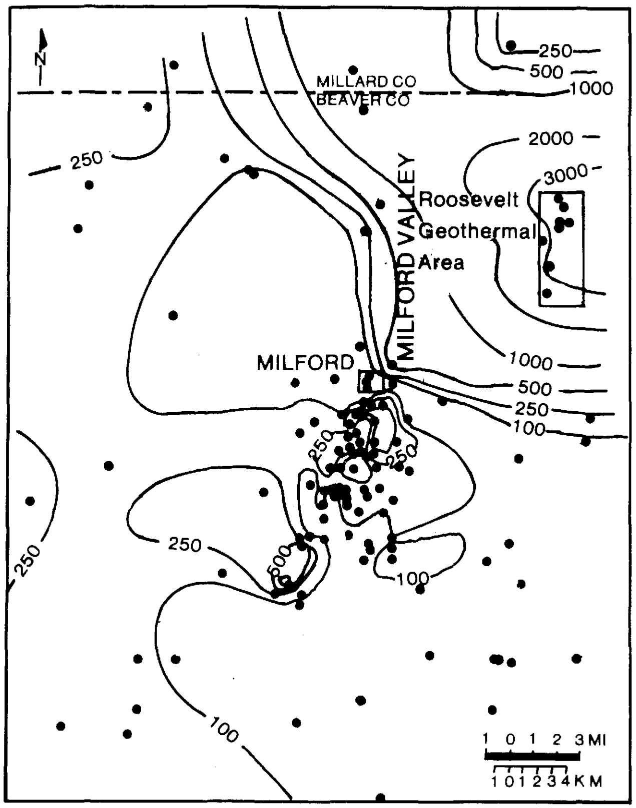


Figure 41. Concentration of chloride in ground water in the Roosevelt geothermal area and adjacent parts of Milford Valley. Values are parts per million. From Ross, Nielson, and Moore, 1982.

Moore, 1982). The water probably circulates through fractures to unknown depths underneath the Mineral Mountains and then rises into the reservoir along the west side of the range. The presence of young rhyolite domes and flows near the geothermal area in the Mineral Mountains is evidence of a hot intrusive mass under the range that is the source of heat for the system.

Ross, Nielson, and Moore (1982) summarize the data for 14 wells in the area of the Roosevelt geothermal system. These wells range in depth from 382 to 2,286 m. Seven of these wells ranging in depth from 382 m to 2,232 m were classified as producers. Maximum measured well temperature was 254°C. Ross, Nielson, and Moore calculate reservoir temperatures of 217° to 277°C. Brook and others (1979) calculate a mean reservoir temperature of 265°C and calculate that the seven wells could produce 4.5 x 10^5 kg/hr total mass flow at 260°C. The mean reservoir thermal energy calculated by Brook and others is 32 x 10^{18}J. Their total for all other Utah systems with temperatures greater than 90°C is 24 x 10^{18}J.

THERMO GEOTHERMAL SYSTEM

Thermo Hot Springs discharge from two spring mounds near the axis of a narrow part of the Escalante Desert between the Shauntie Hills and the Black Mountains (figure 42). The mounds are composed primarily of siliceous sinter and wind-blown quartz sand and silt. East of the hot springs is a young, 10.3 Ma, alkali ryholite flow or dome which is part of an east-trending alignment of small plugs, domes, and lava flows. The dacitic Horse Valley Formation to the southeast of the hot springs was erupted from many clustered central vents. It and the volcanic rocks of Shauntie Hills to the northwest of the hot springs are the product of a complex of stratovolcanos which erupted between 29 and 19 Ma. There are relatively low-volume Pleistocene?, Pliocene, and Miocene basalt lava flows southeast of the hot springs which may be transitional in origin and age to the Horse Valley Formation (Rowley, 1978). Maximum measured water temperature in the springs is 89.5°C (Mariner and others, 1978) and estimates of the discharge range from 0.5 to 2 l/s. Brook and others (1979) estimate the mean reservoir temperature as 142°C and Rush (1983) estimates the reservoir temperature to be between 140° and 200°C. Klauk and Gourley (1983) report the analysis of four water samples from the Thermo area. These samples yielded quartz conductive temperatures ranging from 128° to 131°C.

The Thermo Hot Springs occur in a north-northeast-trending fault zone that crosses the axis of the Escalante Desert. The most recent movement on the faults passing through the spring mounds is down on the east but movement on some faults in the zone is down to the west (Rush, 1983). To the southeast, a system of west-northwest-trending faults displaces Tertiary volcanic rocks and, locally, Quaternary alluvium. Rowley and Lipman (1975) concluded that the hot springs were controlled by the intersection of these two fault zones.

A regional gravity high (figure 6) extends north across the Escalante Desert immediately east of the hot springs and the gradient on the west of this high may reflect the fault zone that controls the springs. However, the net displacement across the zone indicated by the gravity data is down to the west whereas the most recent movement on half of the faults, including those through the spring mounds, is down to the east. The detailed gravity data (figure 43) indicate that a fault with several hundred meters of net vertical displacement down to the west passes through the area of the hot springs. Sawyer and Cook (1977) interpret a detailed gravity profile across the springs as indicating "bedrock" about 150 m below the surface on the upthrown side of the fault. The thermal area is on the northwest edge of an active part of the Intermountain seismic belt and in an area of abundant local earthquakes.

A hole drilled by Republic Geothermal, Inc. about 1.6 km west-southwest of the springs was in alluvium to a depth of about 350 m. Volcanic rock extended from 350 to 960 m and sedimentary and metamorphic rocks extended from 960 to 1,500 m where granite was encountered. The hole was bottomed in granite at 2,220 m. The maximum temperature measured in the hole was 174°C at 2,000 m (Republic Geothermal, Inc., written communication, 1985).

East of the hot springs is a low-amplitude gravity low. Within the area of this low is the 10.3 Ma rhyolite dome or flow which Rowley (1978) called the rhyolite of the Thermo Hot Springs area. This rhyolite is much younger than the volcanic rocks that make up most of the mountains farther to the northwest and southeast. The gravity low may reflect a large mass of similar rock.

Qa ALLUVIUM (Holocene and Pleistocene)

Qf PAN AND PEDIMENT DEPOSITS (Holocene and Pleistocene) - includes Lake Bonneville deposits.

Qh HOT SPRING DEPOSITS (Holocene and Pleistocene) - siliceous sinter with windblown quartz sand and silt.

Tr RHYOLITE OF THERMO HOT SPRINGS (Miocene) - lava flows or dome of alkali rhyolite, 11.6 Ma.

Thr HORSE VALLEY FORMATION (Miocene) - rhyodacite to dacite lava flows, volcanic mudflow breccia, plugs, minor ash-flow tuff, and mafic rocks, 22-19 Ma. Brown Knoll and Old Rambler Hill are composed of dacitic to andesitic(?) volcanic mudflow breccia and subordinate lava flows, and belong to the mafic member of the formation.

Tsh VOLCANIC ROCKS OF SHAUNTIE HILLS (Miocene) - dacitic lava flows, volcanic mudflow breccia, and ash-flow tuff

High-angle fault of Cenozoic age - most are normal faults. Dashed where approximate, dotted where concealed, queried where uncertain. Bar and ball on downthrown side, bar and ball not shown where concealed.

Figure 42. Geologic map of the Thermo geothermal area. Generalized from Rowley, 1978 and showing Republic Geothermal Inc. well.

The Thermo geothermal area is along an east-trending magnetic gradient that is approximately coincident with the Blue Ribbon lineament as defined by Rowley and others (1978). The magnetic gradient probably defines the south edge of a pluton of batholithic size. A detailed ground magnetic survey (figure 44) (Sawyer, 1977) defines a local magnetic low approximately coincident with the gravity high in the area of the hot spring mounds. This low might reflect alteration associated with the hot springs system as suggested by Sawyer or it may reflect a different magnetization of the rocks uplifted here.

A thermal anomaly defined by the shallow heat flow map and 30 m-depth temperature map (Rush, 1983, figures 9 and 10) of the Thermo geothermal area is approximately coincident with the hot springs mounds. Rush calculated 15 x 10^{13} cal/yr as the probable heat discharge from the entire Thermo geothermal system. In addition to the heat flow high over the hot springs, two areas to the southeast and east have values in excess of 250 mW/m^2 (Rush, 1983). A temperature gradient map (figure 45) defines an extensive thermal area covering most of the width of the Escalante Desert in the area of the hot springs.

An audiomagnetotelluric survey (Gardner, Williams, and Long, 1976) defines a northeast-trending regional apparent resistivity low through the Thermo Hot Springs (figure 46). The station spacing does not provide any detail of the anomaly in the vicinity of the hot springs mounds, but the lowest apparent resistivity (about 2 ohm-m) at the lower frequencies was found at the west mound. The more extensive anomaly appears to reflect, at least in part, low-resistivity sediments along the axis of the Escalante Desert.

The Thermo geothermal system occurs in a north-northwest-trending fault zone, along the axis of a narrow part of the Escalante Desert, in a major east-trending lineament and near a 10.3 Ma rhyolite dome. The rhyolite

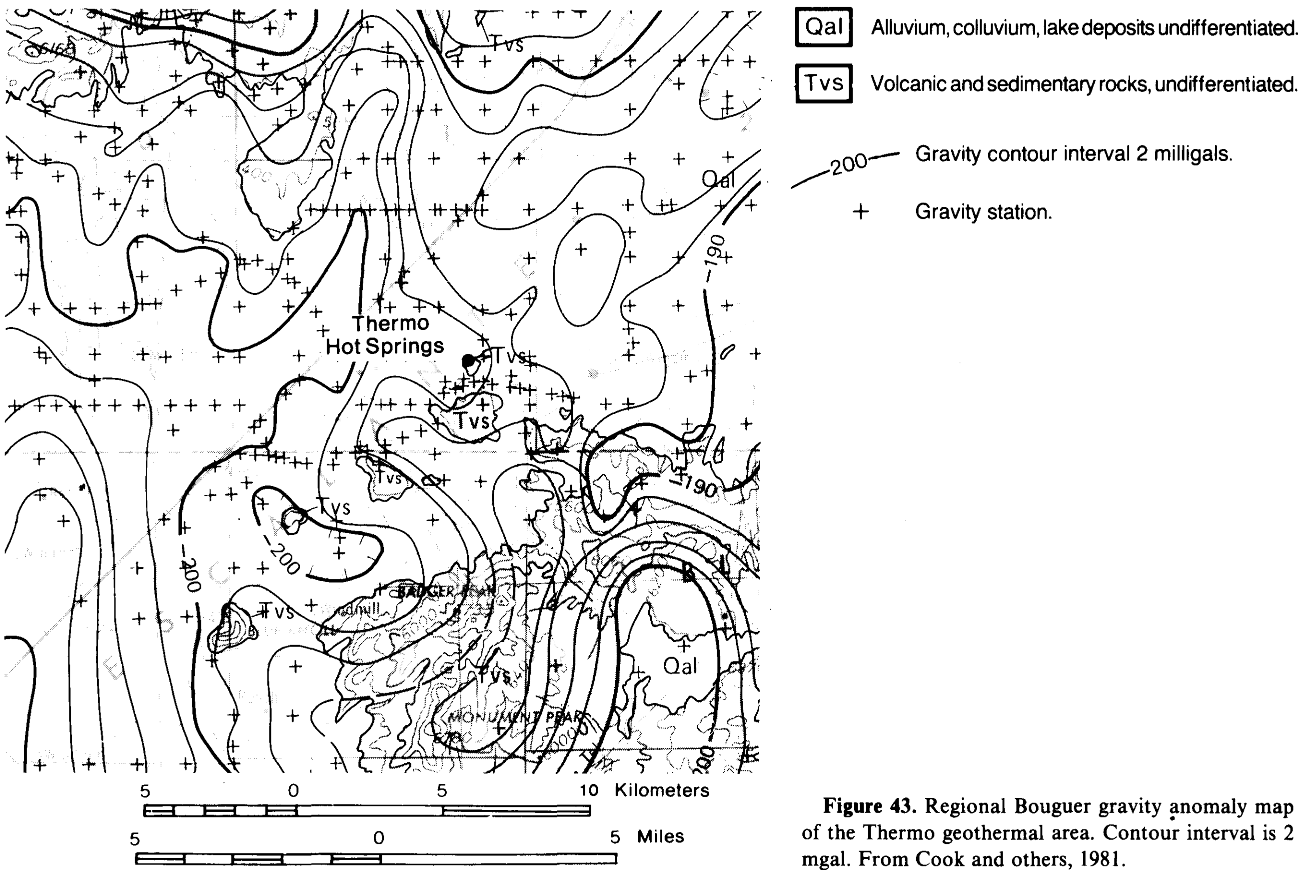


Figure 43. Regional Bouguer gravity anomaly map of the Thermo geothermal area. Contour interval is 2 mgal. From Cook and others, 1981.

dome is too old to be directly related to a heat source for the geothermal system, but does indicate that silicic igneous systems have persisted in this area much more recently than the main phases of activity in the Marysvale-Pioche zone which ended about 14 Ma (Steven and Morris, 1984). The possibility that a local heat source underlies the Thermo area should be considered. The movement of hot water toward the surface in the geothermal system is in a structurally complex area along faults in a north-northwest-trending fault zone on the west side of a structural high. The location of these faults within a valley and at a high angle to the axis of the valley suggests that they are not typical basin and range structures. Without more subsurface information the attitude of the faults cannot be determined, but there is no evidence that these are not relatively high-angle faults that may extend to considerable depth. Data currently available for the Thermo geothermal system are consistent with the heat being derived by deep circulation of ground water in a structurally complex zone. If a large reservoir of thermal water exists at depth it is likely to be in fractured Tertiary volcanic rocks or perhaps older rocks. The calculated reservoir temperatures could be obtained by deep circulation in a thermal gradient that is normal for the region but a local heat source may exist.

Additional exploration to evaluate the system seems justified. This should include a better definition of the deep resistivity structure, and at least one east-west seismic refraction profile across the system and of deep drilling.

Brook and others (1979) calculate that the Thermo geothermal system has a mean reservoir temperature of 142°C and a mean reservoir volume of 8.3 km^3. They indicate this is the third largest and third hottest system in Utah. There is no information available that suggests revising these calculations.

NEWCASTLE GEOTHERMAL SYSTEM

Thermal water has been encountered in wells drilled in several areas of the southwestern part of the Escalante Desert (figure 47). Near Newcastle, water at temperatures up to 95°C has been produced from a well (Christensen Brothers well) 152 m deep, and water from wells in this area is being used in commercial space heating. Brook and others (1979) report a calculated mean reservoir temperature of 130°C and Rush (1983) estimates a reservoir temperature of 140° to 170°C.

The Newcastle geothermal system is near the south end of the Escalante Desert along the northwest edge of a range of hills extending south from Newcastle. The hills are mainly Tertiary lava flows and ash-flow tuffs. The youngest volcanic rocks are dacite and trachyte flow domes north of Newcastle Reservoir and east of the area of Figure 47. The Racer Canyon Tuff is probably from the Caliente caldera complex of southeastern Nevada and the source of the Harmony Hills Tuff is the Bull Valley district to the southwest (Blank, 1959; Noble and McKee, 1972). A Quaternary normal fault zone forms the northwest margin of the hills and the hot water is produced from an aquifer in Quaternary alluvium. There is evidence of

Quaternary movement along the extension of this fault to the northeast in the Silver Peak quadrangle (M. Shubat, oral commununication, 1985). A system of northwest-trending faults occurs in the hills and Rush (1983) infers that a northwest-trending fault in Pinto Canyon is a right lateral strike-slip fault

The temperature profile for the Christensen Brothers well shows a maximum temperature of 107.8°C from 85 to 95 m below the surface, which is in the central part of the principal hot-water aquifer. Below this aquifer the temperature declined to 103.7°C at the bottom of the hole. Rush prepared a map of the temperature at a depth of 100 m (figure 48). This map defines a thermal area centered around the well and elongated north and northeast parallel to the mountain front. The Christensen Brothers well on the heat flow map prepared by Rush (1983, figure 15) is mislocated. Rush (written communication, 1985) believes the heat flow contours are properly located. Rush also prepared a potentiometric contour map of the area which indicates ground water is flowing west and north in the thermal area. Chapman and others (1981) using Rush's (1983) data calculated the thermal power loss from an area of 9.4 km^2 as 13 MW and, assuming a water temperature of 110°C, a volume discharge of 32 l/s. They calculate an energy content of 5.9 x 10^{17}J for a reservoir extending from 75 m to 2 km below the surface with an area of 1.2 km^2.

Detailed gravity data are not available in the Newcastle area but the regional gravity map (figure 6) defines a large northeast-trending gravity low centered northwest of Newcastle. Pe and Cook (1980) interpreted a gravity profile through the north edge of the town of Newcastle as indicating that a normal fault parallel to the mountain front lies near the west edge of Newcastle buried under about 1200 m of alluvium and volcanic rock. The gravity profile can also be interpreted as indicating a fault extending to the surface with the surface trace near the mountain front, and this is the interpretation we prefer.

The Newcastle geothermal area lies along a magnetic gradient with magnetic intensity increasing to the west. This gradient is probably related to the fault along the front of the hills. However, the direction of magnetization of the volcanic rocks is not known and the magnetic anomaly has not been modeled.

The USGS made audiomagnetotelluric soundings at eight stations in the Newcastle area (Hoover and Pierce, 1987). The lowest apparent resistivity values were measured at the station east of the Christensen Brothers well and near the mountain front. Higher apparent resistivity values were measured to the west and southwest, but no soundings were made to the north.

A helium sniffer was field tested in the Newcastle geothermal area by the USGS (Denton, 1976; also in Rush, 1983). Helium concentrations in about 200 gas samples collected 0.6 m below the surface were determined (figure 49). In an area around the Christensen Brothers well and along the range front fault concentrations were greater than 17 ppb (parts per billion) above normal atmosphere, and sample site 300 m east of the well was 170 ppb above atmosphere. Denton concluded that the highest concentrations delineated a dome beneath which hot water has accumulated.

A geothermal model that is consistent with all of the data available for the Newcastle geothermal system consists of hot water at a temperature greater than 108°C rising in the fault zone along the hills southeast of Newcastle. The hot water discharges from the fault zone conduit into an alluviual aquifer and moves northwest and north cooling and mixing with non-thermal water. The primary area of discharge from a fault zone appears to be southeast or east of the Christensen Brothers well. Rush (1983) calculated that the 140° to 70°C reservoir temperatures he estimated could be obtained by circulating water to a depth of 3,000 to 4,000 m.

Water at a temperature higher than that measured in the Christensen

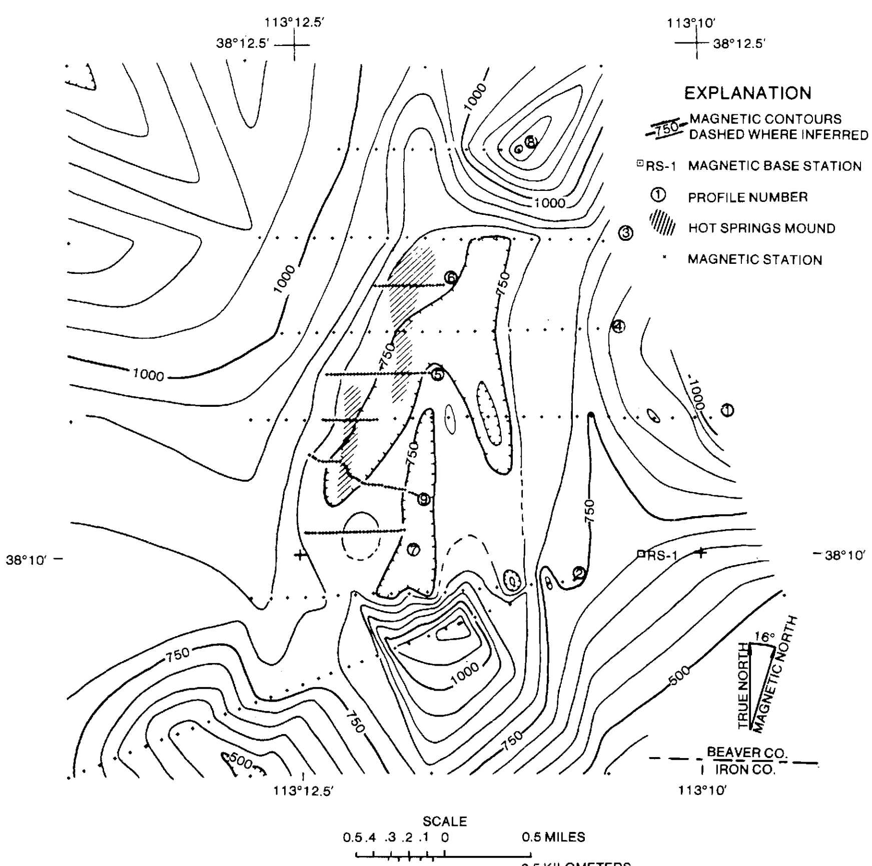


Figure 44. Total intensity magnetic anomaly map of the Thermo geothermal area. Contour interval is 50 gammas. From Sawyer and Cook, 1977.

Brothers well could probably be obtained by drilling into the fault zone; however, it is unlikely that large volumes of water could be produced from the zone. A more productive reservoir may exist at depth either in fractured volcanic rdocks or Cenozoic sediments, although no evidence of a deep reservoir has been reported.

Brook and others (1979) assumed that a deeper reservoir existed and calculated a volume of 6.1 km^3, a temperature of 130°C, and an energy content of 1.9 x 10^{18}J; that is considerably more than the 5.9 x 10^{17}J calculated by Chapman and others (1981) for a smaller reservoir. Chapman and others used a more conservative approach to the energy calculation and one that is better supported by data. However, no information exists that was not available to Brook and others (1979) and therefore, their values are consistent with those used for the other systems.

Less is known about the Newcastle geothermal system than any other high-temperature system in Utah. Until the results of additional exploration become available the possibility should be considered that a moderately high-temperature reservoir (greater than 150°C) with adequate volume to justify development may exist in the Newcastle area. A seismic refraction profile could be used to obtain information on the major faults that apparently control flow of thermal water into the shallow subsurface and provide information on the structure and seismic velocity of rocks underlying the valley. Using this information it should be possible to locate a deep drill hole to test the deeper parts of the geothermal system.

PAVANT VALLEY

In Pavant Valley, located in the southern part of the Sevier Desert, there is a north-trending alignment of Quaternary basalt vents and flows and one small flow of Quaternary rhyolite. Within and near this zone of Quaternary volcanic rock are Meadow and Hatton Hot Springs (figure 3) and several thermal wells. Meadow and Hatton Hot Springs are in what Rush (1983) describes as a "very low alluvial spring mound" about 7 km across. At Meadow Hot Springs discharge temperatures as high as 41°C have been measured (Mundorff, 1970), but the maximum temperature reported by Rush (1983) was 30°C. In September 1985, a temperature of 32°C was measured. Mundorff (1970) estimated the discharge as 4 l/s. Rush (1983) estimated the total discharge from Meadow and Hatton Hot Springs in the summer of 1976 as less than 1 l/s.

Hatton Hot Springs discharges from a spring-deposited travertine ridge about km long, up to 0.5 km wide, and up to 20 m high (figure 50). The flow is generally low and in some years it has been reported that no flow occurs. Mower (1965) reported that a temperature of 143°F (62°C) was measured for springs along Devil's Ridge west of the town of Hatton and the flow was 1 or 2 gpm (about 0.1 l/s). Mundorff (1970) reported that in 1967 Hatton Hot Springs had not flowed for several years, but that in May 1957 the discharge temperature was 100°F (38°C) and in August 1958 the discharge was 25 gpm (1.6 l/s). Rush (1983) reports a discharge temperature of 36°C for the springs and a

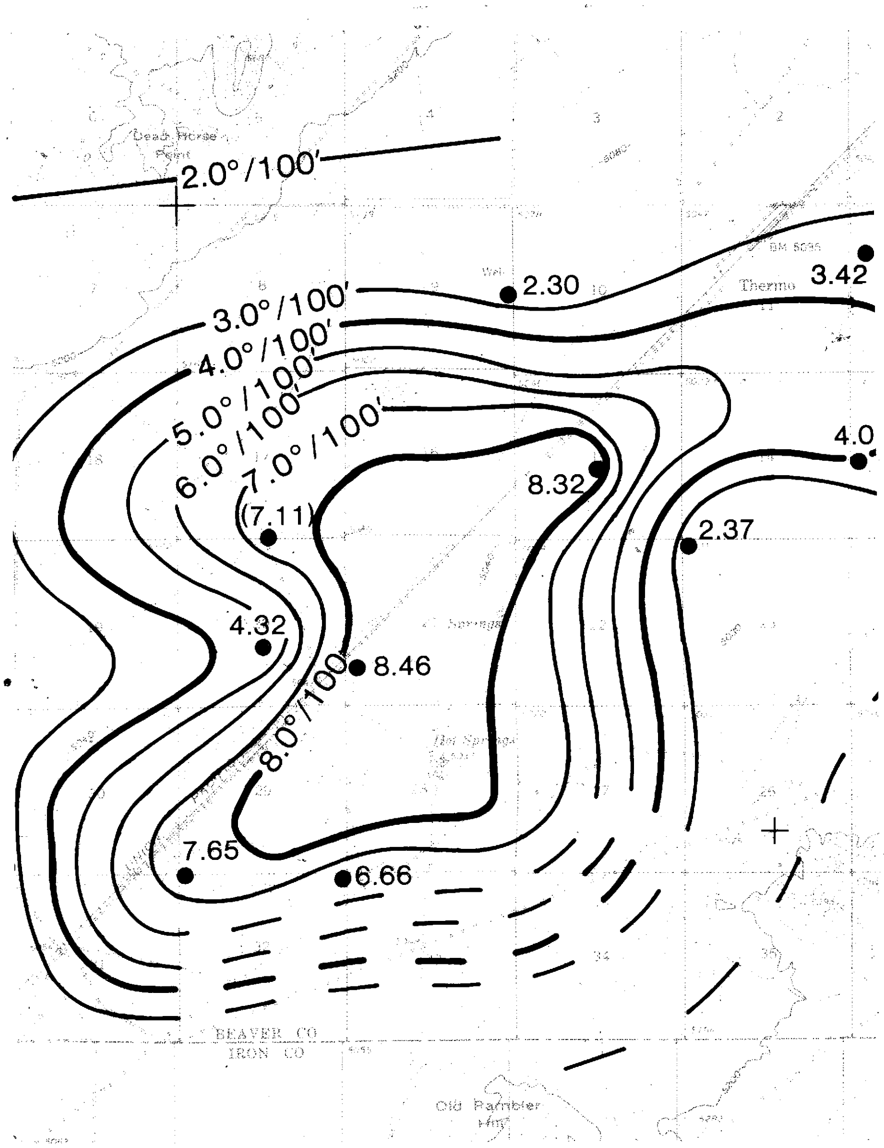


Figure 45. Temperature-gradient contour map of the Thermo geothermal area. Values are in °F/100 feet in the interval between 107 and 137 m (350 and 450 feet) below the surface. From Republic Geothermal, Inc., written communication, 1985.

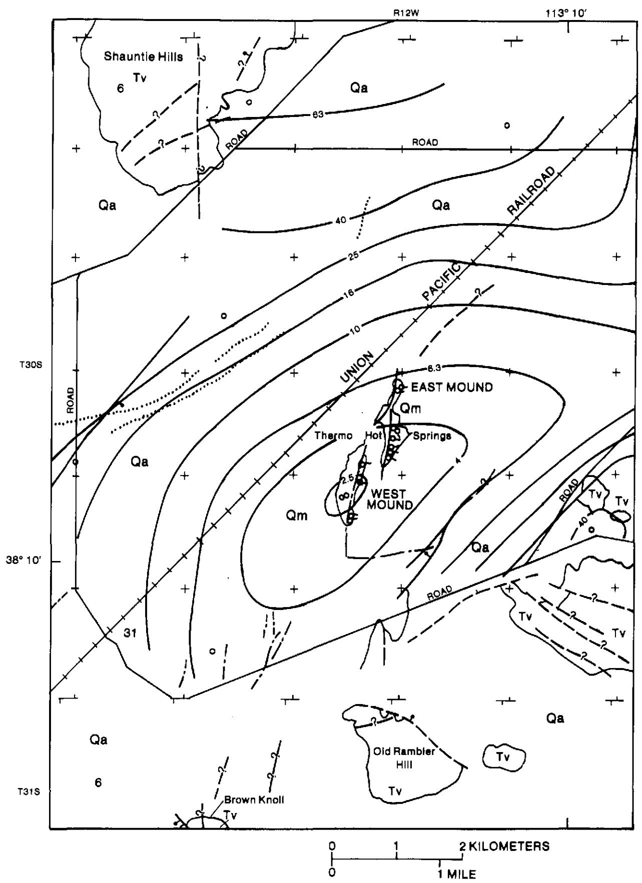


Figure 46. Audiomagnetotelluric contour and geologic map of the Thermo geothermal area. Contour values are apparent resistivity in ohm-m at 7.5 Hz with east-west telluric line. From Gardner, Williams, and Long, 1976 and geology modified from Rush, 1983. Geologic units Qm—spring-mound deposits, Qa—alluvial deposits, and Tv—volcanic rocks.

temperature of 67°C at 5 m below the surface in a well a few tens of meters to the north. In October 1985, Hatton Hot Springs was discharging water at 63°C. Flow was not measured.

Because of the large range of water temperature reported for Hatton Hot Springs the water was resampled for this report and the analysis is given in table 4 (for sampling techniques and analytical procedures see Klauk and Gourley, 1983, p. 17). Also listed is the analysis reported by Mundorff (1970) which is very similar. Geochemical reservoir temperatures based on these data using the applicable chalcedony geothermometer also correlate well. The USGS sampled Hatton Hot Springs in July 1985 and measured a temperature of 66°C, a flow of 0.24 l/s, and obtained a chemical analysis similar to those given in table 4 (Walter Holmes, oral communication, 1985). The explanation for the relatively low temperature of Hatton Hot Springs reported by Mundorff (1970) and Rush (1983) as contrasted with Mower (1965) and the current temperature is not known.

There is evidence of high heat flow over a substantial area adjacent to Hatton Hot Springs. The nearest heat flow determination reported by Rush (1983) was 420 mW/m^2 about 2 km southwest of the spring. Rush also prepared a rapid snowmelt anomaly map for an area around Hatton Hot Springs (figure 51). The map defines three areas totaling about 2.3 hectares where freshly fallen snow melted rapidly. Rush concluded that subsurface temperatures as high as 70°C can be expected at depths as shallow as 10 m in these areas.

Thermal water has been reported in several wells in Pavant Valley north of Meadow and Hatton Hot Springs. The water temperature in most of these wells is 20° to 30°C; however, one well on the west side of the desert is described as "hot" by Goode (1978).

The southern Sevier Desert has both the youngest basalt and the youngest rhyolite flows dated in Utah (figure 3). Quaternary volcanic rocks in the Pavant Valley are mostly basalt flows ranging in age from about 700 years to about one million years. The small rhyolite flow at White Mountain is 0.4 million years old (figure 52). Twin Peaks, near the south end of the desert, are Tertiary rhyolite domes 2.4 to 2.7 million years old.

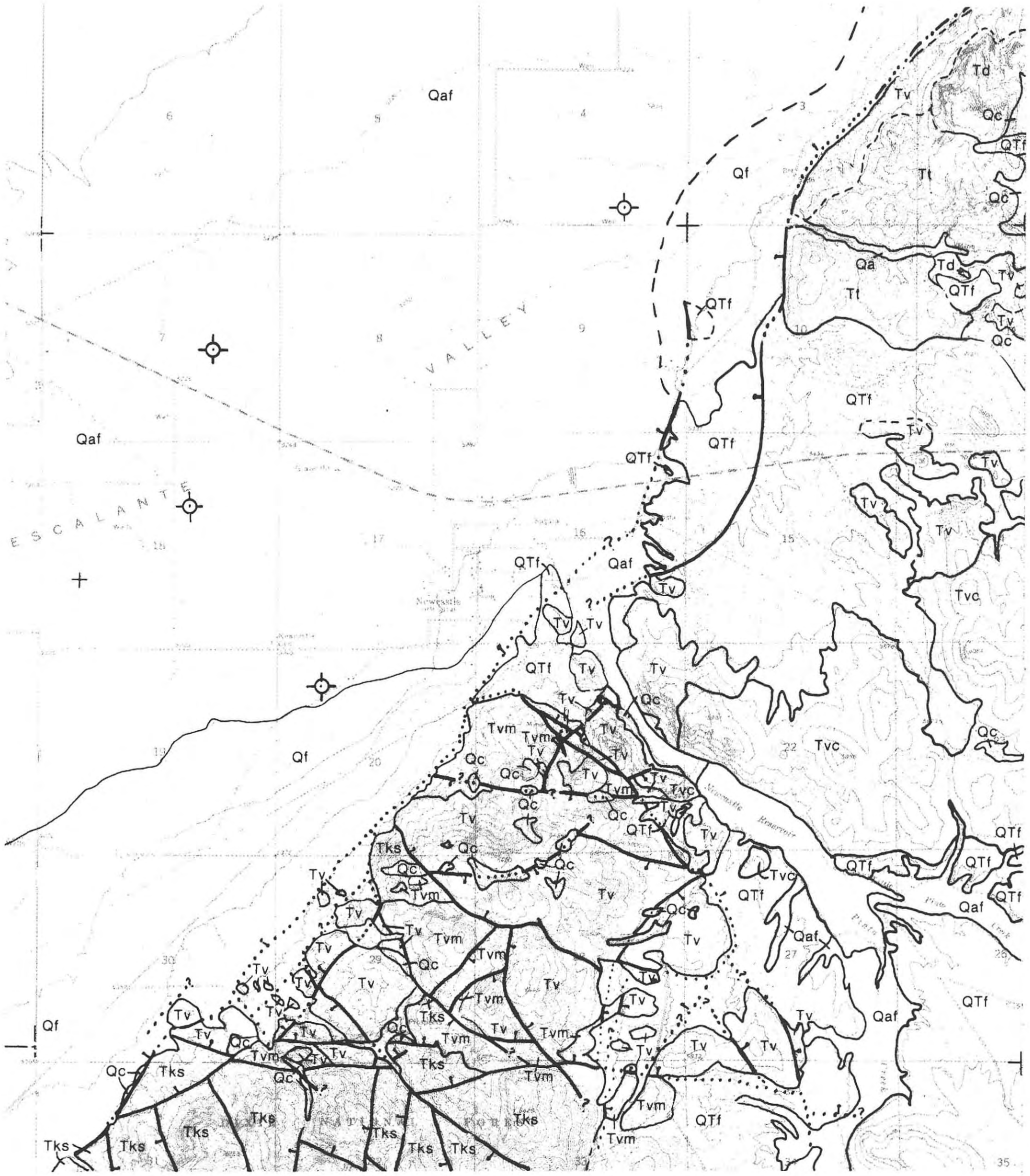
Relatively few earthquakes have been recorded in Pavant Valley but a north-trending zone of late Quaternary faults extends through the area. One fault in the zone has been dated as having moved less than 10,000 years ago. Despite the lack of historic seismicity the area should be considered a tectonically active zone.

The regional magnetic anomaly map (figure 7) defines a broad, low-amplitude high extending northwest across the northeastern part of the desert. Extending south from this deep-seated high are two zones of high magnetic intensity that,

TABLE 4. GEOCHEMICAL ANALYSES AND RESERVOIR TEMPERATURES-HATTON HOT SPRINGS

	UGMS (10/25/85)	Mundorff (6/19/57)
Na	1041.00	1090 (Na + K)
K	137.00	--
Ca	438.00	465
Mg	86.00	89
Fe	0.30	--
S_iO_2	48.00	44
B	3.50	--
Li	3.05	--
Sr	6.06	--
HCO_3	425.00	427
SO_4	1018.00	985
CL	1790	1780
F	3.80	--
NO_3	--	2.4
TDS Meas.	4848.00	--
TDS Calc.	4783.00	4670
pH Meas.	7.0	6.7
ph Calc.	7.1	--
Geothermometer		
Chalcedony	70°	66°
Quartz Conductive	100°	96°
Na-K-Ca	201°	not calc.-data lacking
Na-K-Ca Mg corrected	85° (R=21.8)	not calc.-data lacking
Field measurement	63°	30°

Note: Concentrations in ppm, temperatures in Centigrade.



- **Qa** ALLUVIUM (Holocene and Pleistocene)
- **Qaf** VALLEY-FILL ALLUVIAL-FAN DEPOSITS (Holocene and Pleistocene)
- **Qf** ALLUVIAL-FAN DEPOSITS (Holocene and Pleistocene)
- **Qc** COLLUVIUM (Holocene and Pleistocene)
- **QTf** OLDER ALLUVIAL-FAN DEPOSITS (Pleistocene, Pliocene ?, and Miocene ?)—principally dissected alluvial-fan and pediment gravel deposits
- **Td** DACITE FLOW-DOME ROCKS (Miocene?)
- **Tt** TRACHYTE FLOW-DOME ROCKS (Miocene?)
- **Tvc** VOLCANICLASTIC ROCKS (Miocene)—breccia and debris flow of fluvial and lava origins with clasts of Tv
- **Tv** TERTIARY VOLCANICS (Miocene and Oligocene)—includes calc-alkalic Racer Canyon, Harmony Hills, Condor Canyon, Leach Canyon, and Isom ash-flow tuffs (19-25 Ma)
- **Tvm** TERTIARY VOLCANICS, MAFIC (Oligocene)—mafic lava flows
- **Tks** TERTIARY AND CRETACEOUS SEDIMENTARY ROCKS—includes fluvial and lacustrine deposits of the Claron and Iron Springs Formations

Contact, dashed where inferred

Fault - Dotted where concealed, queried where uncertain. Bar and ball on downthrown side.

well

Figure 47. Geologic map of the Newcastle geothermal area. Generalized from Galyardt, unpublished map, and Schubat and Siders, mapping in progress.

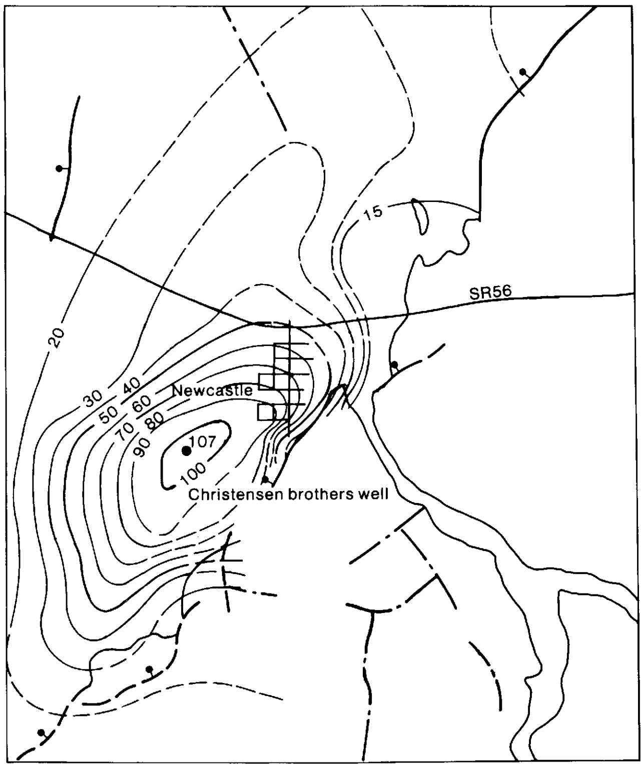


Figure 48. Contour map of temperature at 100 m depth in the Newcastle geothermal area. Contour interval is 10°C. From Rush, 1983.

at least in part, are caused by near-surface rocks. One zone passes through Fumarole Butte and continues about 15 km to the south. The other extends over the basalt flows west of Fillmore. A separate magnetic high is located over the rhyolite and basalt in the Twin Peaks area. Although exposed volcanic rocks produce a significant part of the magnetic relief over Pavant Valley, they do not cause the entire anomaly.

Serpa and Cook (1980) have modeled the magnetic data obtained in a survey flown 305 m above the surface, assuming all the rocks were magnetized in the direction of the present earth's field. They concluded most of the anomaly was produced by normally magnetized basalt flows generally less than 200 m thick. The east-west gravity and magnetic profile they modeled nearest to the hot springs has a complex distribution of magnetic units in the subsurface. One unit is a thin layer of high magnetic susceptibility and very low density that extends more than 20 km west of the surface basalt. Although the validity of this layer and the magnetic model in general is questionable, there is evidence that the flows are more extensive in the subsurface than at the surface and that magnetic units occur at considerable depth.

A large regional gravity high extends down the east side of Pavant Valley (figure 6). The anomaly is approximately parallel to the Pavant Range and Canyon Mountains to the east but the crest of the anomaly lies several kilometers west. West of the high is an elongate gravity low with areas of low closure on the north and south. The southern area of closure is the location of Twin Buttes and has been interpreted by Carrier and Chapman (1980) as indicating a buried pluton. The area of basalt flows is aligned along the gravity gradient between the regional high and low, but Meadow and Hatton Hot Springs are near the axis of the high.

Serpa and Cook (1980), in modeling two profiles across Pavant Valley, assume that the entire gravity anomaly is produced by mass variations within 5 km of the surface. We prefer an interpretation that attributes a major part of the regional gravity high to greater depths. The Serpa and Cook model assumes the vents for the lava flows are controlled by normal faults which flatten at depth and do not extend to depths greater than 4,000 m. Hal Morris (written communication, 1984) has prepared a regional cross section that crosses Pavant Valley. His cross section is based on surface geology supplemented by information from two deep oil tests in the area, and assumes a structural style suggested by the seismic reflection line to the north (Allmendinger and others, 1983). His section is also consistent with the gravity data. A generalized version of his section is shown in Figure 53. An important difference between the Morris cross section and those by Serpa and Cook (1980) is that Morris assumes the normal faults that provide the conduits for the basaltic magma dip steeply and extend to depths greater than 4 km. We believe that the Morris cross section is a good representation of the subsurface and can be incorporated into a model of a geother-mal system for the area of Meadow and Hatton Hot Springs.

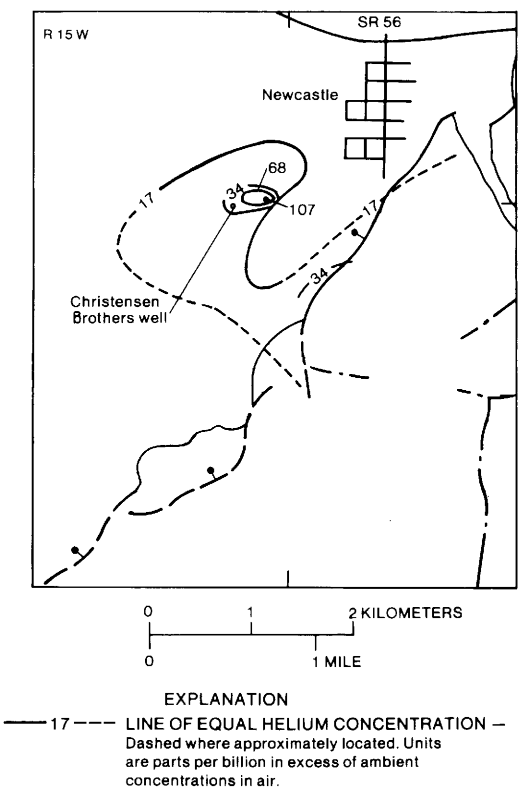


Figure 49. Contour map of helium concentration in soil gas in the Newcastle geothermal area. From Rush, 1983, as adapted from Denton, 1976.

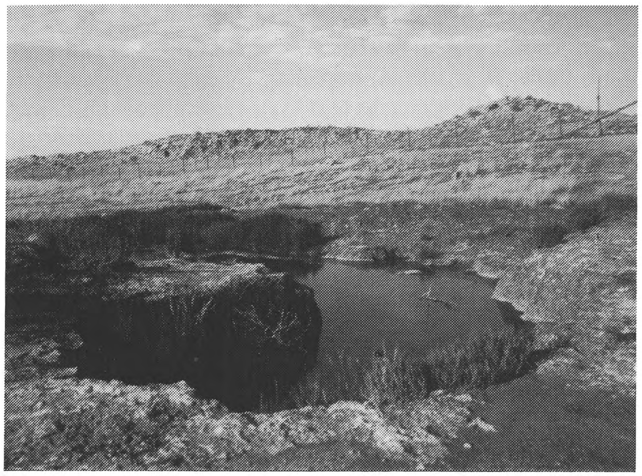


Figure 50. Photograph of Hatton Hot Springs looking southwest with travertine ridge in background.

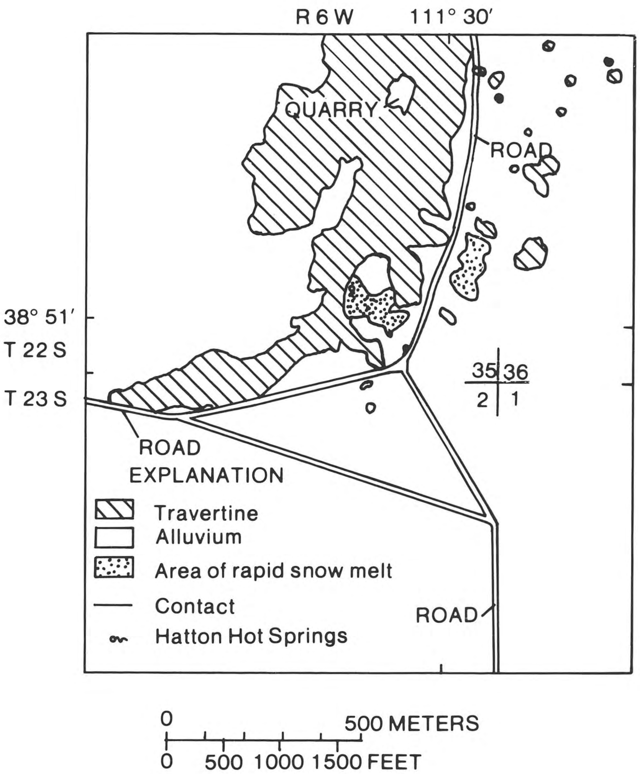


Figure 51. Map of areas of rapid snowmelt in the Hatton Hot Springs area. From Rush, 1983.

Mundorff (1970, p. 40), in discussing Meadow and Hatton Hot Springs, concluded that "Some heat undoubtedly is furnished by the nearby volcanic flows of late Tertiary and Quaternary age." Rush (1983) assumed that no shallow magmatic heat source was present and calculated that the reservoir temperatures he estimated could be obtained by circulating water to depths of 2 to 3 km in the normal thermal gradient for the region. Rush also estimated that the water discharged in the springs was 40 percent thermal and 60 percent non-thermal water. Serpa and Cook (1980) also conclude the water is heated by deep circulation.

Although it is unlikely that the basalt flows in Pavant Valley are contributing heat to a geothermal system as indicated by Mundorff (1970) or that the near-surface feeder dikes for the flows are significant sources of heat, the possibility that a local heat source exists in the area should not be discounted. The small 400,000 year old rhyolite flow at White Mountain, 7 km north of Hatton Hot Springs, is evidence of a large thermal event in the upper crust that occurred in Quaternary time. A significant reservoir of heat related to this event may still exist in Pavant Valley and be the source of heat for a high-temperature geothermal system.

With existing data, only a very preliminary appraisal of the geothermal resource in Pavant Valley is possible. Moderate-temperature waters in the Meadow and Hatton Hot Springs and in wells distributed over much of the desert appear to be a mixture of thermal and non-thermal water. The waters in the hot springs have computed reservoir temperatures of 69°C (Meadow) and 66°C (Hatton) (Rush, 1983). Using a mixing model for these waters Rush concluded that the maximum reservoir temperature was 120°C. The computed reservoir temperature for the water collected at Hatton Hot Springs in 1985 is 70°C using the same chalcedony geothermometer. No direct evidence has been reported for a reservoir of high-temperature water.

Evidence of an underlying thermal anomaly is abundant. The extensive basalt probably originated at considerable depth and was erupted at the surface through narrow conduits, thus most heat transferred from the deep heat reservoir was dissipated into the atmosphere as the basalt flow cooled. The rhyolite, being more viscous than basalt, probably originated from a magma body at much shallower depths than the basalt. The parent body must have been at least partly molten 400,000 years ago when the rhyolite erupted. The parent body, if large enough, may still be a local source of heat. Several lines of evidence suggest that the crust in the Pavant Valley region is thinner than in surrounding areas (Thompson and Zoback, 1979), and it is the axis of the Sevier oval of Steven and others (written communication, 1986). Although it is not defined by existing heat flow determinations, a heat flow high might be associated with a local thinning of the crust thus making this a likely area for the development of deep-circulation geothermal systems. The deep-penetrating normal faults suggested by the vents and basalt flows along the fault zone could provide conduits for deep circulating waters.

Despite the absence of direct evidence for a high-temperature geothermal resource in Pavant Valley, young volcanic rocks and evidence of a thin crust suggest that this is a favorable area for the development of geothermal systems. Reservoirs might occur in either fractured Tertiary or fractured or cavernous pre-Tertiary rocks. The logical first step in the investigation of the geothermal resource in this area is a regional heat flow survey to determine if an important thermal anomaly is present. The results of this study could be used to determine what, if any, additional exploration is justified.

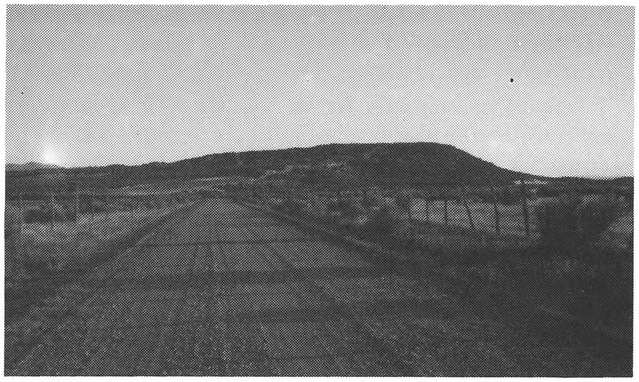


Figure 52. Photograph, looking south, of 0.4 my old rhyolite flow at White Mountain.

UNDISCOVERED GEOTHERMAL SYSTEMS

Quaternary sediments cover one-half of the Sevier thermal area with Cenozoic volcanic rocks covering more than half of the remaining area. Pre-Cenozoic rocks are generally exposed in thrust plates that have slid eastward tens of kilometers during the Sevier orogeny. Thus, the deep subsurface geology cannot be reliably inferred from the surface geology. Five of the seven known high-temperature geothermal systems in the area were first identified because of hot springs. A sixth system (Cove Fort) was identified because of sulphur deposits, altered ground, and emissions of gas. There is no surface evidence of the seventh system (Newcastle) although hot water has been found at relatively shallow depths. Undiscovered high-temperature systems may extend into the shallow subsurface, as with the Newcastle system, and there may be other systems that are confined to greater depths. Brook and others (1979) estimated the undiscovered resource in the eastern part of the Basin and Range Province to be about five times the discovered systems, excluding the Roosevelt system. They estimated that about half of this resource was in the extension of known systems and about half in unidentified systems. They apparently thought it unlikely that another system of the size and quality of the Roosevelt system would be discovered in Utah.

The Roosevelt geothermal system is unusual. Only one comparable system has been proven in the Basin and Range Province: Desert Peak in western Nevada. A system with a temperature approaching that at Roosevelt probably requires a local heat source. Such heat is most likely to occur in areas of active volcanism and be associated with very young silicic volcanic rocks. The youngest volcanic rocks in Utah are in a zone, here called the Sevier volcanic belt, trending a few degrees east of north from the central Mineral Mountains to near Pavant Butte. Bimodal volcanism in the belt began about 3 million years ago and has continued episodically. Near the two ends of this belt are rhyolite flows and domes about 0.5 million years old; at the north end are basalt flows about 700 years old. Local heat sources may exist at several locations in the Sevier volcanic belt.

In addition to the local heat source for the Roosevelt geothermal system another key element is the structure. The Mineral Mountains have been elevated more than any other range in the region. This large, late Cenozoic uplift appears to have been the major factor in forming the structures that resulted in the development of the geothermal system. No similar uplifts exist in the Sevier volcanic belt and, therefore, it is unlikely that a system with structure similar to that of the Roosevelt system exists.

The intersection of the Sevier volcanic belt and the east-west structures through the Cove Fort area might be an area where both a local heat source and favorable structure occur. The Cove Fort geothermal area is on the east edge of the Sevier volcanic belt in the area of this intersection. Much of the central and western parts of the belt in this area are covered by Quaternary basalt flows which would likely obscure any surface expression of an underlying geothermal system. This is a promising area for the occurrence of an undiscovered resource.

Regional heat flow determinations are needed to guide additional exploration in the Sevier volcanic belt. The existing heat flow values are poorly distributed. Some are clearly affected by local hydrologic conditions and do not accurately indicate deep, conductive heat flow. A few good heat flow determinations in key areas could make a substantial contribution to the understanding of regional heat flow. A survey designed to define all of the thermal anomalies of interest would involve a

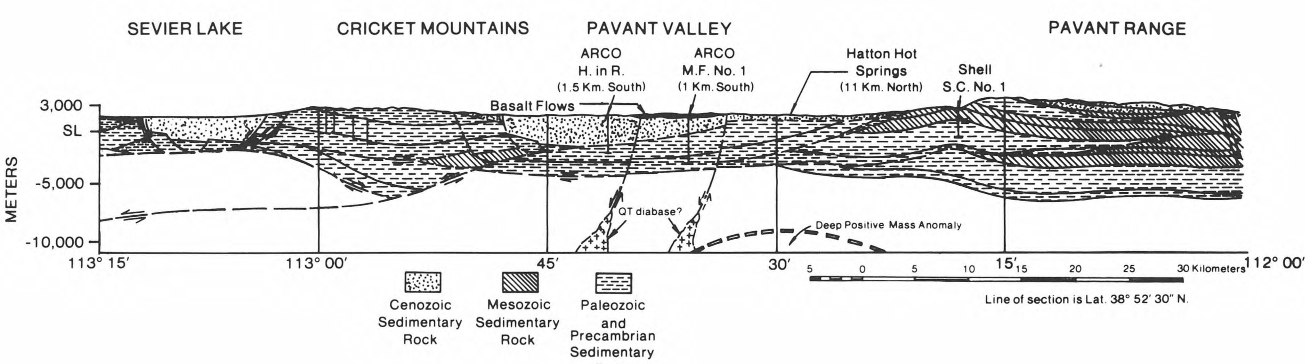


Figure 53. Geologic cross section across Pavant Valley. Generalized from H.T. Morris, written communication, 1984.

great many heat flow determinations and is not economically feasible.

Although the Sevier volcanic belt appears to be the most promising area for locating large, high-temperature geothermal systems, other areas within the Sevier thermal area should not be ignored. Klauk and Gourley (1983) used existing drill holes to identify two thermal areas in the Escalante Desert in addition to Thermo and Newcastle. Water wells with anomalously high temperatures are widespread throughout the Sevier thermal area and are particularly common in the Delta area. No evidence exists that suggests high-temperature geothermal systems are related to any of these low-temperature thermal anomalies but the possibility should be considered. Several, as yet undetected, thermal anomalies likely occur in the Sevier thermal area, and any one of them might be related to a high-temperature system. Temperatures should be measured in all wells drilled in the region in an effort to detect thermal anomalies. The USGS is conducting a ground-water study in Pavant Valley and this may further define the shallow thermal anomalies in this area.

The exploration of the seven known high-temperature systems in Utah provides an evaluation of most exploration techniques. This experience indicates that most standard techniques, when properly applied and interpreted, provide meaningful information on geothermal systems. However, the wide variation in the geology of the different systems clearly indicates no single exploration plan is valid throughout the area and the exploration program must be tailored for individual areas. The case histories of the Roosevelt system (Ward and others, 1978; Ross, Nielson, and Moore, 1982) and Cove Fort (Ross and Moore, 1985) provide good examples of the application of most exploration techniques.

CONCLUSIONS

The geothermal resources of Utah are in an early stage of commercial development. Two of the seven known high-temperature geothermal systems, Roosevelt and Cove Fort, are currently generating significant amounts of electrical energy. These two geothermal systems appear capable of supporting larger development. The approximate size and quality of one of the systems (Cove Fort) has not yet been determined. The Thermo, Fumarole Butte, and Newcastle systems have not been completely evaluated and additional exploration might define a significant resource in these systems particularly if water at a temperature of 150°C is of interest. The geological and geophysical setting of the southern Pavant Valley appears favorable for the occurrence of high-temperature geothermal systems and, therefore, additional exploration is encouraged. Undiscovered geothermal systems are likely to exist elsewhere in the Sevier thermal area and all new geological, geophysical and geochemical data obtained should be examined for evidence of local thermal anomalies.

Although no direct evidence of high-temperature geothermal systems in Utah outside of the Sevier thermal area has been reported, some are likely to exist. New data and new concepts on the control of geothermal systems will likely lead to the discovery of new geothermal systems in several areas of the state.

The pace of discovery and development of Utah's high-temperature geothermal resource cannot be accurately predicted. However, geothermal systems are already an important energy source in Utah and the importance of geothermal energy—both high-and low-temperature—to the state is likely to continue to increase in the foreseeable future.

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