Reconnaissance of the Hydrothermal Resources of Utah

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By F. EUGENE RUSH

GEOHYDROLOGY OF GEOTHERMAL SYSTEMS

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A brief description of the hydrothermal resources of Utah

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CONVERSION OF UNITS

For use of those readers who may prefer to use inch-pound units rather than metric units, the conversion factors for the terms used in this report are listed below:

Multiply metric unit	*By*	*To obtain inch-pound unit*
calories (cal)	3.974×10^{-3}	British thermal units (B.t.u.)
degrees Celsius (°C)	1.8°C + 32	degrees Fahrenheit (°F)
milligrams (mg)	1.543×10^{-2}	grains
kilograms (kg)	2.205	pounds (lb)
liters (L)	.2642	gallons (gal)
meters (m)	3.281	feet (ft)
millimeters (mm)	3.937×10^{-2}	inches (in)
centimeters (cm)	.3937	inches (in)
hectometers (hm)	3.281×10^{-2}	feet (ft)
kilometers (km)	.6214	miles (mi)
square centimeters (cm^2)	.1550	square inches (in^2)
square kilometers (km^2)	.3861	square miles (mi^2)
cubic meters (m^3)	35.31	cubic feet (ft^3)
liters per second (L/s)	15.85	gallons per minute (gal/m)

GEOHYDROLOGY OF GEOTHERMAL SYSTEMS

RECONNAISSANCE OF THE HYDROTHERMAL RESOURCES OF UTAH

By F. EUGENE RUSH

ABSTRACT

Geologic factors in the Basin and Range province in Utah are more favorable for the occurrence of geothermal resources than in other areas on the Colorado Plateaus or in the Middle Rocky Mountains. These geologic factors are principally crustal extension and crustal thinning during the last 17 million years. Basalts as young as 10,000 years have been mapped in the area. High-silica volcanic and intrusive rocks of Quaternary age can be used to locate hydrothermal convection systems. Drilling for hot, high-silica, buried rock bodies is most promising in the areas of recent volcanic activity. Southwestern Utah has more geothermal potential than other parts of the Basin and Range province in Utah. The Roosevelt Hot Springs area, the Cove Fort-Sulphurdale area, and the area to the north as far as 60 kilometers from them probably have the best potential for geothermal development for generation of electricity. Other areas with estimated reservoir temperatures greater than 150°C are Thermo, Monroe, Red Hill (in the Monroe-Joseph Known Geothermal Resource Area), Joseph Hot Springs, and the Newcastle area. The rates of heat and water discharge are high at Crater, Meadow, and Hatton Hot Springs, but estimated reservoir temperatures there are less than 150°C. Additional exploration is needed to define the potential in three additional areas in the Escalante Desert.

INTRODUCTION

PURPOSE AND SCOPE

The State of Utah has an abundance of thermal springs and probably is a promising area for geothermal exploration. This study, a 2-year reconnaissance of the geothermal resources on the public lands of Utah, was begun by the U.S. Geological Survey in the summer of 1975. The purpose of the reconnaissance was to describe the general geohydrologic framework for geothermal systems, and to provide more detailed descriptions and evaluations than were previously available for some of the more promising hydrothermal systems. Most of the data were gathered and evaluated during the summer of 1975 and in 1976. This report presents the results of the study. A data report (Rush, 1977) has already been released which contains subsurface-temperature data for 30 wells.

PREVIOUS WORK

The earliest known reference to geothermal systems of Utah is by Gilbert (1890, p. 332–335); he briefly described Fumarole Butte, gaseous discharges from the butte, and nearby Crater Hot Springs (fig. 1). Many years later Stearns, Stearns, and Waring (1937, p. 96, 108–109, 179–183) described about 60 thermal springs in Utah and summarized the literature about them. A similar summary was made by Waring (1965). At the East Tintic mining district, about 30 km northwest of Nephi (fig. 1), Lovering and Goode (1963) worked with geothermal gradient holes in their search for hydrothermal ore bodies. In another mining area, the Iron Springs district about 16 km west of Cedar City, (fig. 1), Sass and others (1971, p. 6399–6400) described temperature measurements in eight drill holes. They concluded that the heat flow in that area is about 1.9×10^6 cal/cm²/s.

Heylmun (1966) and Batty and others (1975, p. 233–241) presented brief, general discussions of geothermal resources in Utah. However, both papers presented few data. A comprehensive data report on the thermal springs of Utah (Mundorff, 1970) contains an abundance of information for about 60 springs. Additional data have been published by Milligan, Marselli, and Bagley (1966). Additional estimates of reservoir temperatures were made for 47 hydrothermal systems in Utah by Swanberg (1974), using the Na-K-Ca geothermometer developed by Fournier and Truesdell (1973).

Olmsted and others (1975, p. 27–76) provided a discussion of hydrothermal concepts and of exploration and evaluation techniques in a report describing hydrothermal systems in the western part of the Basin and Range province. This discussion was useful as a guide in the study and other workers probably will find it of similar value. The University of Utah, Department of Geology and Geophysics, is currently (1977) investigating Roosevelt Hot Springs KGRA (Known Geothermal Resource Area) and other areas, primarily evaluating various geophysical techniques for geothermal exploration. In one of the resulting reports, Parry, Berson, and Miller (1976) describe the geology and water chemistry of Roosevelt and Monroe Hot Springs.

WELL- AND SPRING-NUMBERING SYSTEM

The system of numbering wells and springs in Utah, used herein, is based on the cadastral land-survey system of the U.S. Government. The number describes the position on the land net of the well, spring, or site where geothermal observations were made. In the land-survey system, the State is divided into four quadrants by the Salt Lake base line and meridian, and these quadrants are designated by the uppercase letters A, B, C, and D, indicating the northeast, northwest, southwest, and southeast quadrants, respectively. Numbers designating the township and range (in that order) follow the quadrant letter, and all three are enclosed in parentheses. The number after the parentheses indicates the section, and the section commonly is followed by three letters indicating the quarter section, the quarter-quarter section, and the quarter-quarter-quarter section (generally 4 hm²)[1]; the letters a, b, c, and d indicate,

[1]The basic land unit, the section, is ideally 2.6 km²; however, many sections are irregular. Such sections are subdivided into 4 km² tracts, generally beginning at the southeast corner, and the surplus or shortage is taken up in the tracts along the north and west sides of the section.

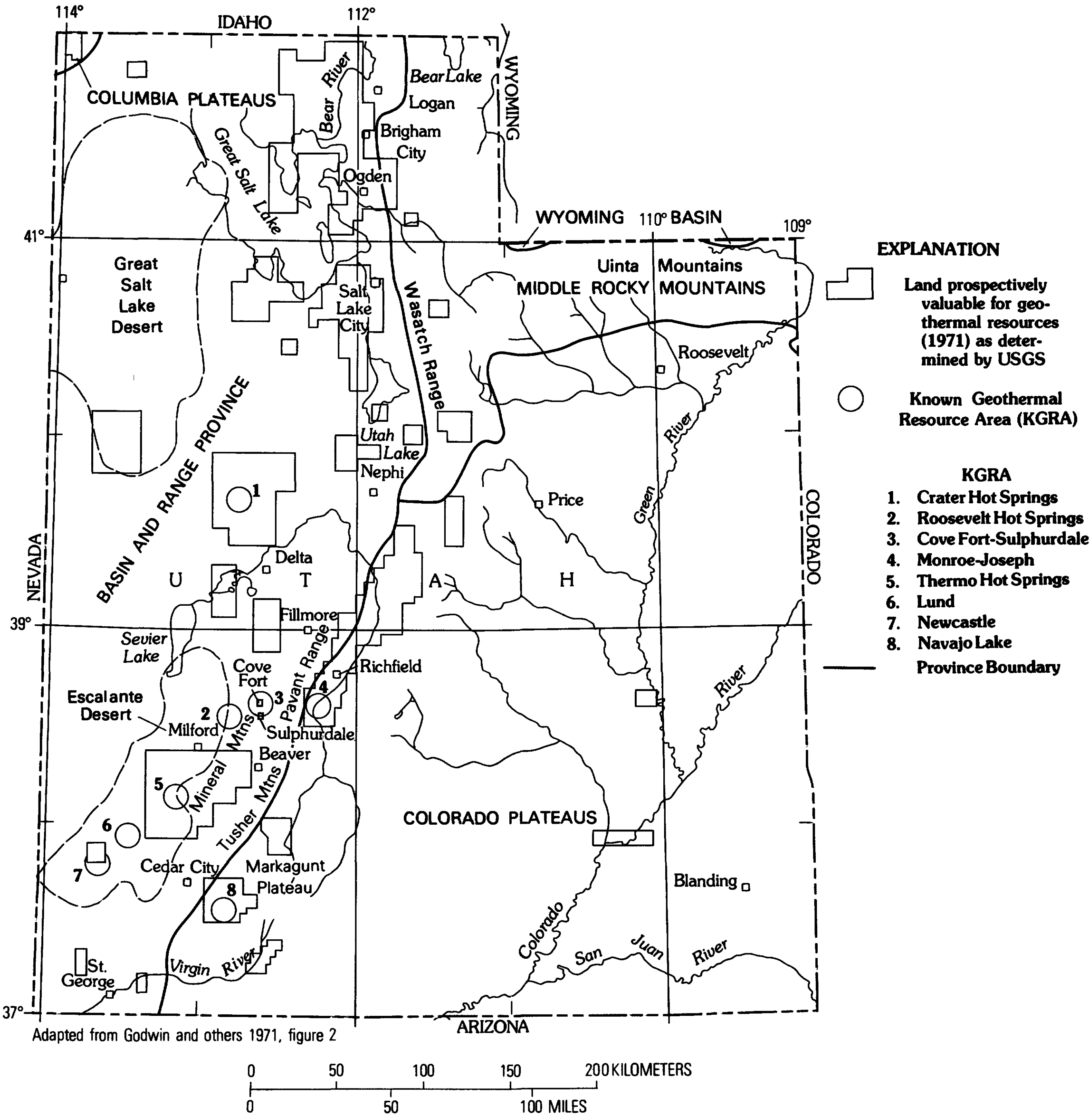


FIGURE 1.—Land prospectively valuable for geothermal resources in Utah, and an index map of Utah.

respectively, the northeast, northwest, southwest, and southeast quarters of each subdivision. The number after the letters is the serial number of the well or spring within the 4 hm^2 tract; the letter "S" preceding the serial number denotes a spring. If a well or spring cannot be located within a 4 hm^2 tract, less than three location letters are used and the serial number is omitted. Thus (C-29-8)9ba designates a well in the NW¼NW¼ sec. 9, T. 29 S., R. 8 W. The numbering system is illustrated in figure 2.

REGIONAL GEOLOGIC SETTING

Utah essentially includes parts of three physiographic provinces as defined by Fenneman (1931): the Middle Rocky Mountains, the Colorado Plateaus, and the Basin and Range province (fig. 1). Each area is described briefly below, but more emphasis is given to the Basin and Range province because of its greater potential for geothermal development.

MIDDLE ROCKY MOUNTAINS

In Utah the Middle Rocky Mountains province includes the Wasatch Range and the Uinta Mountains. The Wasatch Range rises to an altitude of 2,400 to 3,400 m above sea level, or between 1,200 and 2,000 m above the valley floors of the Basin and Range province. The range is an uplifted block of folded and faulted strata, bounded on the west by a major fault zone, the Wasatch Fault. The Uinta Mountains are generally higher than the Wasatch Range, reaching altitudes greater than 4,000 m above sea level. They are described by Fenneman (1931, p. 177) as a flat-topped anticline. Most of the consolidated rocks that crop out in both mountain ranges are pre-Cenozoic sedimentary or silicic plutonic rocks.

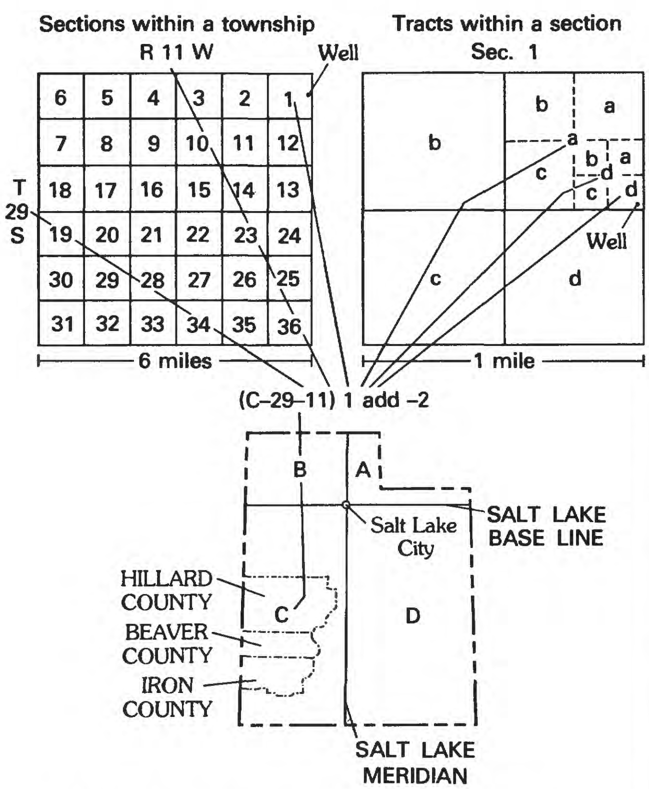


FIGURE 2.—Well- and spring-numbering system.

COLORADO PLATEAUS

The province, as implied by its name, is an area of broad uplift with strata nearly horizontal in most places. The outcrops are mostly Mesozoic and older sedimentary rocks. Notable exceptions are Tertiary and Quaternary volcanic rocks in the southwestern part of the province (south-central Utah) and a few scattered Tertiary intrusive bodies in the southeastern part of the state. Land-surface altitudes are commonly between 1,500 and 3,000 m above sea level. A continuation of the Wasatch Fault zone marks the western boundary of the province.

BASIN AND RANGE

The Basin and Range province is characterized by elongated, mostly north-trending mountain ranges and narrow flat-bottomed valleys. The province contains rocks widely ranging in composition and age. The older rocks consist of a wide variety of Mesozoic and Paleozoic sedimentary rocks and their metamorphosed equivalents. Overlying the sedimentary and metamorphic rocks are Cenozoic volcanic rocks and valley fill. Valley fill, mostly alluvium, may be as thick as 3,000 m in some basins. Lacustrine deposits are common.

According to Stewart (1971), most or perhaps all of the major valleys in the Great Basin of the Basin and Range province can be considered to be grabens, and most or all of the mountains can be considered to be horsts or tilted horsts. The geometry of block faulting related to these structures requires sizable east-west extension of the thin crust under the province; the extension was estimated by Stewart to be about 2.4 km for each major valley. Most of this extension took place in the last 17 million years, or perhaps even in the last 7–11 million years. In Utah, grabens which are not bounded by faults of equal displacement generally have the master fault on the east side.

In western Utah, igneous rocks and hydrothermal mineral zones are in well-defined east-west belts (fig. 3), each successively younger to the south (table 1), according to Stewart, Moore, and Zietz (1977). Figure 3 shows the distribution of igneous rocks less than 6 million years old; some are less than 10,000 years old. They are mostly basalt and crop out generally in southwestern Utah. These very young igneous rocks do not seem to be along an east-west belt but rather on an alinement parallel to Basin and Range structure. The implication is that Basin and Range structure controls the distribution of these rocks; whereas the older belts predate Basin and Range structure. Stewart, Moore, and Zietz (1977) see genetic and age similarity between these belts and the belt of upper Cenozoic volcanic rock extending along the Snake River Plain in southern Idaho eastward into the Yellowstone region of northwestern Wyoming.

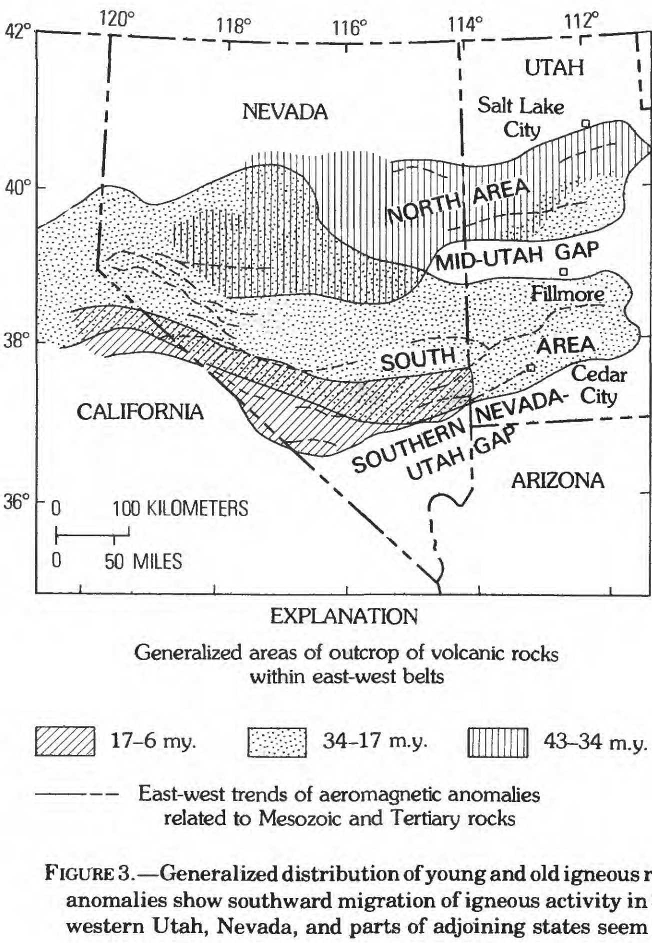
Very young volcanic rocks in Utah are reported by Rowley, Anderson, and Williams (1975, p. B18), and Smith and Shaw (1975, p. 82). Volcanic rocks probably less than 10,000 years old are found near Fillmore and 65 km southwest, 50 km south, and 30 km southeast of Cedar City (fig. 3). These young rocks are largely basaltic, but scattered rhyolitic cones are known (Liese, 1957).

Silicic intrusive rocks were emplaced at the same time as the silicic volcanic rocks (Whelan, 1970). The largest exposure of such an intrusive body in Utah is the Mineral Mountains, 80 km north of Cedar City.

REGIONAL HYDROLOGIC SETTING

Precipitation on the semiarid valley floors of the Basin and Range province in Utah, as well as much of the Colorado Plateaus, averages less than 200 mm per year (U.S. Weather Bureau, no date). The higher mountains of the Wasatch Range and the Uinta Mountains generally receive precipitation of 1,000 mm or more per year, most accumulating as snow in the winter. The mountains of the Basin and Range province average about 500 mm or less yearly.

The relatively large amounts of precipitation that fall in the mountains flow toward the ground-water reservoirs and major streams in two ways: (1) Flow from the mountains in small streams. Part of this water infiltrates the stream beds and percolates to the water table. (2) Flow percolates directly into the fractures and pore space of consolidated rocks of the mountains; this water then flows in the subsurface across the consolidated rock—valley-fill contact. The latter is considered to be the smaller volume of water in most areas; an exception is in areas of carbonate rocks that have developed interconnected solution channels. Most of the ground water



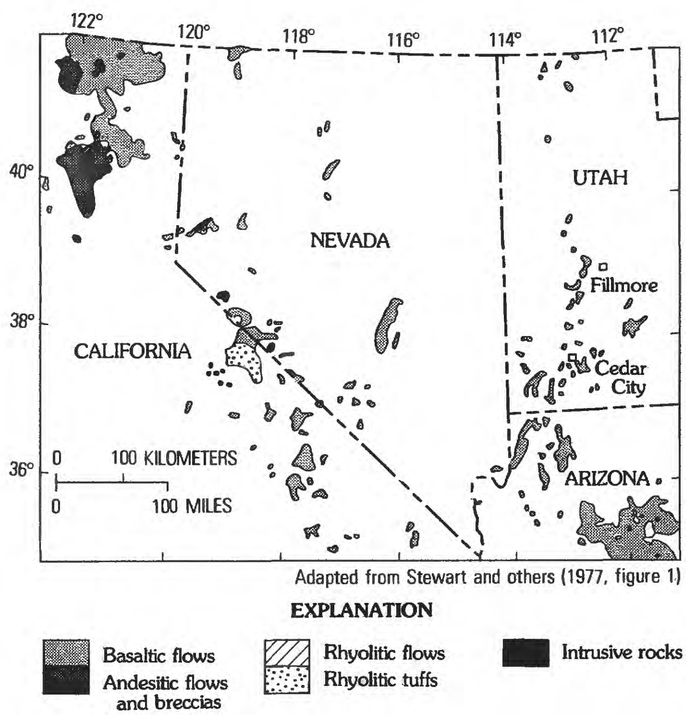


FIGURE 3.—Generalized distribution of young and old igneous rocks. *A*, East-west patterns of Cenozoic igneous rocks and positive aeromagnetic anomalies show southward migration of igneous activity in California, Nevada, and Utah. *B*, Igneous rocks less than 6 million years old in western Utah, Nevada, and parts of adjoining states seem to parallel Basin and Range structure.

TABLE 1.—*Relation of hydrothermal areas to mineral belts and to age of Cenozoic igneous rocks*

[The areas of western Utah are as shown in figure 4. The areas and mineral belts are listed in order from north to south. The ages of the rocks are from Stewart and others (1977, p. 71)]

Hydrothermal areas	Mineral belt	Approximate average age of rocks (m.y.)	General hydrothermal characteristics
North of Cenozoic igneous rock area.	---------------	>43	Small, low-temperature thermal springs.
Northern Cenozoic igneous rock area.	Oquirrh-Uinta	35	Small or low-temperature thermal springs common.
	Deep Creek-Tintic	33	Includes Crater Hot Springs and hot water at east Tintic mining district.
Between Cenozoic igneous rock areas.	Mid-Utah gap	--	Very few thermal springs or wells.
Southern Cenozoic igneous rock area.	Wah Wah-Tushar	26	Includes Cove Fort-Sulphurdale area, Roosevelt, Thermo, Joseph, and Monroe Hot Springs.
	Iron Springs	20	Includes Newcastle area.
South of Cenozoic igneous rock area.	Southern Nevada-Utah gap	--	---------------

circulates to depths of only several tens to several hundred meters, and generally has a temperature near or slightly higher than the ambient land-surface temperature for the lowlands (10°–16°C). Some ground water migrates through fault-created fractures to great depth where it absorbs heat from wall rock. This heated water returns to the land surface along with the shallow-circulating ground water, where it is ultimately discharged as springs, to streams, by evapotranspiration, or from wells. In agricultural areas, a secondary source of ground-water recharge to the valley-fill reservoir is infiltration from fields, canals, and reservoirs. Summers are usually hot with low humidity. As a result, on lowlands, potential lake evaporation greatly exceeds precipitation.

Some of the valleys in the Basin and Range province are hydrologically isolated; that is, water that falls as precipitation remains within the basin until it is discharged back to the atmosphere. However, in areas where interconnected solution channels have developed, ground water may follow complex flow paths beneath interbasin divides and flow for tens or hundreds of kilometers and for thousands of years before discharging to the land surface. Commonly, this interbasin flow involves moderately deep circulation beneath mountain ranges and, as a result, its discharge is significantly above ambient temperature. A probable example of discharge from an interbasin regional flow system is the Fish Springs group (Mundorff, 1970, p. 37), about 90 km northwest of Delta on the south edge of the Great Salt Lake Desert. The estimated discharge is 1.4 m^3/s. Water temperatures reportedly range from 18° to 76°C. The recharge areas for these springs are probably to the south and west and probably include parts of Nevada.

The principal sources of geothermal fluids are water stored in the hydrothermal reservoir and water entering the geothermal-circulation system as recharge from precipitation. Because of the semiarid climate of much of the area, most geothermal development for generation of electricity will remove fluids from storage at a higher rate than natural replenishment.

GEOTHERMAL RELATIONS

REGIONAL HEAT FLOW

The average conductive heat flow to the earth's surface is approximately 1.6 HFU (1 heat-flow unit [HFU] = 1×10^{-6} cal/cm²/s or 1 μcal/cm²/s, according to Schubert and Anderson, 1974). Considerable variation in flow exists in Utah. Based on data from Sass and others (1971) and Sass and Munroe (1974), the area of highest heat flow in Utah is the Basin and Range province, which has heat-flow values commonly in the range of 1.5 to 2.5 HFU. By comparison the "Battle Mountain High" (in Nevada) is an area of abnormally high heat flow where conductive heat-flow values are commonly in the range of 2.5 to 3.5 HFU. According to Lachenbruch and Sass (1977), the "Battle Mountain High" may be a part of a larger region of exceptionally high heat loss extending from western Nevada to Yellowstone Park, Wyo., and including the northwestern corner of Utah. No data were collected as part of this study to determine whether the northwestern corner of Utah is an area of very high heat flow. Data from this study, and the work of others, indicate that the Basin and Range province in Utah probably has an average heat flow of about 2 HFU.

The Colorado Plateaus and the Middle Rocky Mountains provinces in Utah have heat-flow values near the average for the earth's surface. Values published by Sass and others (1971) and by Sass and Munroe (1974) for these areas generally range from 1.3 to 2.0 HFU and average about 1.6 HFU.

The causes of variation in heat flow to the earth's surface are complex and poorly understood. Some of the factors that contribute to the diversity are (1) variations in crustal thickness, (2) convection of magma beneath and possibly within the lower parts of the crust, (3) movement of ground water in hydrothermal convection cells, (4) variations in the distribution of radioactive elements such as uranium, thorium, and potassium-40 in crustal rock, (5) intrusion into the upper crust of young magmas, and (6) general circulation of shallow ground water.

RELATION OF THERMAL WATERS TO HYDROGEOLOGIC FRAMEWORK

The present level of geothermal knowledge is in part presented by White and Williams (1975) and is briefly summarized as follows: (1) Some geothermal systems are supplied only by a "normal" geothermal gradient; some by magmatic heat. (2) Youngest igneous rocks have the best potential as heat sources. (3) Purely basic volcanic systems rarely form thermal anomalies of economic interest for generation of electricity, whereas silicic volcanic systems may do so if they are large enough. (4) Young basic volcanoes are produced by magma sources in the mantle and, under some conditions, are potential indicators of buried high-level silicic bodies with no obvious surface manifestations. (5) Silicic magmas are always erupted from high-level storage chambers, probably in the upper 10 km of the crust. (6) High-temperature convection systems can be sustained for many thousands of years with heat from high-silica magma bodies. Perhaps because of the very high viscosities of such magmas, these systems are associated with magma chambers at shallow levels in the crust. (7) Cooling by hydrothermal convection tends to offset continued heating, but the rate of supply of magma from deep crustal or mantle sources is the dominant heat supply for both high-level magmatic and hydrothermal systems. (8) Basic magmas rise through the crust to the surface through narrow pipes and fissures created by faulting; the individual magma pulses are volumetrically small, and such systems contribute little stored heat to the upper crust until magma chambers begin to form at high levels. (9) Fluid temperature is of critical importance in determining how a hydrothermal system may be utilized and is the most important single factor in evaluating a system. Hot-water convection systems can be divided into those of three temperature ranges: (a) Above 150°C; these systems may be considered for generation of electricity; (b) from 90°C to 150°C; these systems are attractive for space and process heating; and (c) below 90°C; these systems are likely to be utilized for heat only in locally favorable circumstances. (10) Natural geysers and active deposition of siliceous sinter (amorphous hydrous silica) are reliable indicators of subsurface temperatures at least as high as 180°C. On the other hand, travertine deposits (calcium carbonate) and opaline residues produced by sulfuric acid leaching have no reliable relation to reservoir temperature. (11) In the Basin and Range province, heat flows are sufficiently high that the existence of a thick blanket having low thermal conductivity (high-porosity clay beds, for example) could locally raise the temperatures to levels of economic interest. The above abbreviated summary can be used as a partial guide to the general relation of thermal waters to the hydrogeologic framework.

Most thermal springs and wells are in valleys near the margins of the mountains. Spring positions probably are controlled by Basin and Range faults. Some springs are in valley bottoms; others are on upland slopes. Only a few thermal springs are in a mountainous setting; the most prominent example is Midway Hot Springs, 45 km southeast of Salt Lake City (Mundorff, 1970, p. 46).

Recharge to the hydrothermal systems is by either meteoric water in the nearby shallow, ground-water reservoir or by percolation in the nearby mountains. The dominant driving force for deep circulation probably is the difference in density between cold recharge water and hot upflowing water, but head differences between recharge area and springs may contribute as shown in figure 4.

The hot water rising from the hydrothermal reservoir may be greatly diluted by shallower-circulating cold water, lowering the temperature but increasing the flow of thermal springs above the hydrothermal reservoir. As shown in figure 4, only part of the upflow of thermal water may directly reach the land surface, because part may enter near-surface aquifers and cool by conduction as it flows laterally from the spring area.

The model described above may be modified in several ways to approximate the variety of hydrothermal systems: (1) No convecting magma may be present in the upper crust, but rather the heat source may be deeper in the crust or in the mantle; (2) the downward flow of cold water to the hydrothermal reservoir may be through any deep permeable route; and (3) calcium carbonate and amorphous silica may deposit on the walls of the upflow zone, creating an isolated or semi-isolated conduit or self-sealing cap. As a result, movement of water between the isolated part of the hydrothermal convection cell and the surrounding rock, alluvium, or land surface would be reduced or eliminated.

Vertical flow of shallow nonthermal ground water may modify conductive heat flow to the land surface. In areas of recharge (downward percolation of water), heat flow to the land surface is decreased; whereas in areas of ground-water discharge to the atmosphere (upward flow) the normal heat flow to the land surface is increased. The magnitude of the distortion is related to the velocity and quantity of vertical flow of water.

Conductive heat flow is computed as follows:

$$\mathrm{HFU} = 10^{-2} KI$$

where

HFU is heat-flow unit, in microcalories per square centimeter per second,

K is thermal conductivity of the rock material, in millicalories per centimeter per second per degree Celsius, and

I is the geothermal gradient, in degrees Celsius per kilometer.

The estimated K values used in this report are summarized in table 2. From the formula it has been seen

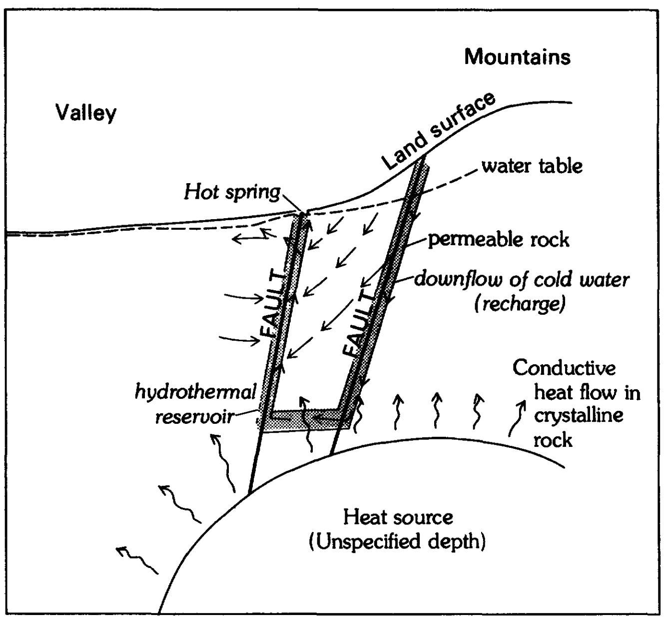


FIGURE 4.—Conceptual model of a hydrothermal convection cell.

TABLE 2.—*Thermal-conductivity values used in this report*

[Values based on work done in Nevada and Utah by Olmsted and others (1975, p. 64; Olmsted, oral commun., 1976) and Sass (written commun., 1975). See also Sass and others (1976)]

Lithology	Estimated conductivity values (mcal/cm/s°C)	
	Saturated	Unsaturated
Clay	1.5–2.5	1.5–2.0
Silt	2.5–3.0	2.0–2.5
Sand	3.0–4.0	2.0–3.5
Gravel	4.0–6.0	3.0–5.0
Basalt	4.5–6.0	3.0–5.5
Vesicular basalt	3.5–5.0	2.5–4.5
Sedimentary rocks	5.0–9.0	4.0–7.0
Igneous rocks of felsic to intermediate composition	6±	----

that, with no variation in regional heat flow, the geothermal gradient generally will vary as lithology varies. As a result, heat flow to the land surface is a more consistent index of geothermal-resource potential than shallow temperature-gradient data.

The values of thermal conductivity for the various lithologies, as listed in table 2, are based on laboratory determinations using cores and drill cuttings. The values listed in the table are the ranges into which most samples fall. The factors that control thermal conductivity are: (1) texture of rock or alluvium, (2) mineral content, (3) layering within the rock, (4) porosity and pore size, (5) degree of water saturation, and (6) the dissolved mineral content of the saturating water. There may be additional factors.

Chemical composition of thermal spring waters can be used to estimate hydrothermal-reservoir temperatures. The geothermometers used in this report are listed in table 3 and are from Fournier and Rowe (1966) and Fournier and Truesdell (1973). The basic assumptions in using these geothermometers (Fournier and others, 1974) are: (1) Temperature-dependent reactions occur at depth; (2) there is an adequate supply of chemical constituents; (3) water-rock chemical equilibrium occurs at reservoir temperature; (4) re-equilibration at lower temperatures as the water flows to the surface is negligible; and (5) hot water is not diluted by shallow cold water. If mixing occurs, the least adversely affected geothermometers listed in table 3 are the Na-K-Ca geothermometer and the graphic method. Therefore, where temperature calculations differ and mixing is suspected, the Na-K-Ca and graphic-method calculations should be favored.

DISTRIBUTION OF THERMAL WATERS

In this report a thermal well or spring is defined as having a water temperature above the average land-surface ambient temperature, which, as stated previously, commonly ranges from 10°C to about 16°C depending on altitude and geographic location. In the

TABLE 3.—*Formulas for geothermometers used in this report*

[From R. O. Fournier, written commun., 1975 and 1977. Concentrations: Na, K, and Ca in molality; SiO_2 in mg/kg, which is approximately the same as mg/L]

Formula identification used in table 6	Formula	Remarks
NKC	*Na-K-Ca Geothermometer* $t_{°C} = \dfrac{1647}{\log(\mathrm{Na/K}) + \beta \log(\mathrm{Ca/Na}) + 2.24} - 273$	If magnesium concentrations are in excess of 10 mg/kg or if travertine (tufa) is being deposited, computed temperature may be too high. $\beta = 1/3$ in all computations for table 7.
Q	*Quartz (conductive) Geothermometer* $t_{°C} = \dfrac{1309}{5.19 - \log SiO_2} - 273$	Computed temperature usable if spring discharge is lower than boiling. Computed temperatures between 120°C and 180°C are of questionable meaning, because either chalcedony or quartz may be controlling the silica concentration.
C	*Chalcedony Geothermometer* $t_{°C} = \dfrac{1032}{4.69 - \log SiO_2} - 273$	Used mostly in basalt areas. Computed temperature usable if <120°C, but may be usable up to 180°C.
S	*Amorphous Silica (silica gel) Geothermometer* $t_{°C} = \dfrac{731}{4.52 - \log SiO_2} - 273$	Computed temperature should be considered if opal deposits are present.
GM	Graphic methods for estimating temperature of a hot-water component in a mixed water (Truesdell and Fournier, 1977, and Mariner, R. H., written commun., 1978). Use solubility curve of quartz above 120°C and chalcedony solubility curve below this temperature where reservoir is volcanic rock.	A plot of dissolved silica and enthalpy are used. Valid for spring with temperature lower than about 80°C and flow rate greater than about 2 L/s.

low-altitude areas of southwestern Utah (near St. George), the highest ambient temperatures prevail. Elsewhere on valley floors, the range is commonly from 10°C to 14°C.

A map compiled by the U.S. Geological Survey for Utah (fig. 1) shows the eight KGRA's and lands prospectively valuable for geothermal resources. Not surprisingly, the general geographic distribution is the same as distribution of thermal springs described by Mundorff (1970, fig. 2). Table 4 summarizes information on the KGRA's.

The geographic distribution of thermal waters in Utah is shown in figure 5. This distribution is based on the thermal-spring report by Mundorff (1970) and on data collected during this study. The springs are grouped into northwest, southwest, and Wasatch Range Front areas in figure 5. In table 5, the spring groups have been categorized on the basis of several hydrothermal characteristics. The springs of the southwest area have the most favorable characteristics. This is also the area of geothermal leasing of Federal lands (fig. 6). Most of the following discussions will be concerned with hydrothermal prospects in the southwest area.

TABLE 4.—*Known geothermal resource areas in Utah*

[October 1976. Total area includes all land irrespective of ownership. Leased areas are 41 federal leases]

KGRA	County	Location		Area (km²)	
		Township (S.)	Range (W.)	Total	Leased
Cove Fort-Sulphurdale	Beaver, Millard	24–26	6–7	100	79
Crater Hot Springs	Juab	13	8	70	70
		14	8–9	--	--
Lund	Iron	32	14	16	14
Monroe-Joseph	Sevier	25–26	3–4	66	2.9
Navajo Lake	Kane	38	8	10	0
Newcastle	Iron	36	15	4.3	0
Roosevelt Hot Springs	Beaver	26	9	121	100
		27	9	--	--
Thermo Hot Springs	Beaver, Iron	29	13	105	54
		30	11–13	--	--
		31	12	--	--
Total (rounded)				490	320

Estimated reservoir temperatures for selected sites are given in table 6. The temperature estimates are based on chemical analyses of geothermometers (table 3). Six prospects may have reservoir temperatures above 150°C and, therefore, they may have potential for generation of electricity: Roosevelt, Thermo, and Joseph Hot Springs, Newcastle area, the Cove Fort-Sulphurdale area, and the Monroe-Red Hill Hot Springs complex. Six other hot springs listed in the table may have reservoir temperatures in the 90–150°C range, and therefore, they have value for space and process heating. The locations of the first group are shown in figure 1. The table contains location numbers for all the sites. In the table, graphic-method calculations are based on the quartz-solubility curve unless the chalcedony-solubility curve is indicated.

DISCUSSION OF PROSPECTS

In this section, geothermal systems that are considered to have the best potential for development are discussed. These systems include seven of the eight KGRA's in Utah and several other areas of interest. The Lund KGRA was not included because of lack of data. (See table 18 for temperature data for areas not discussed in this section.)

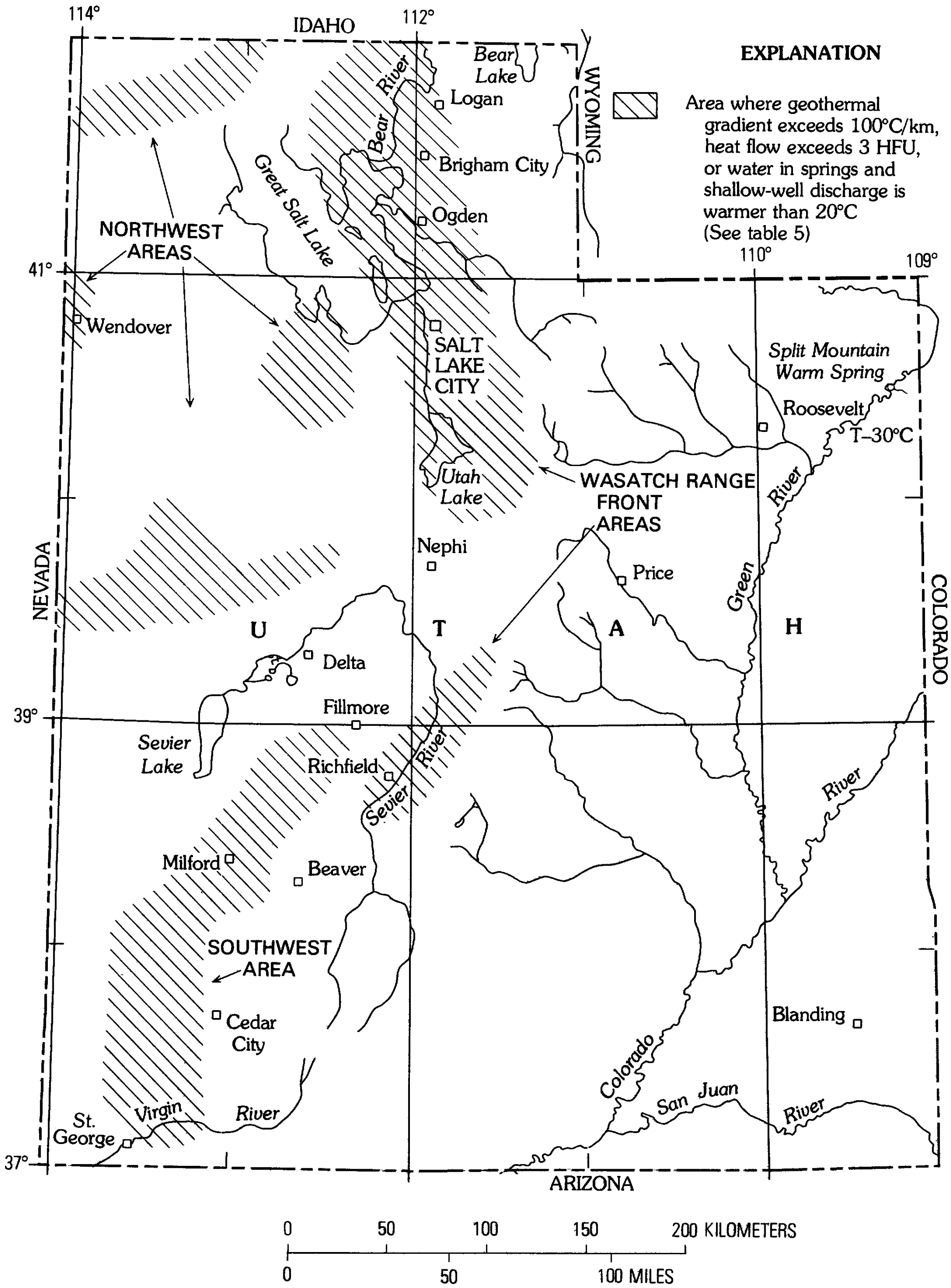


FIGURE 5.—Geothermal areas of Utah.

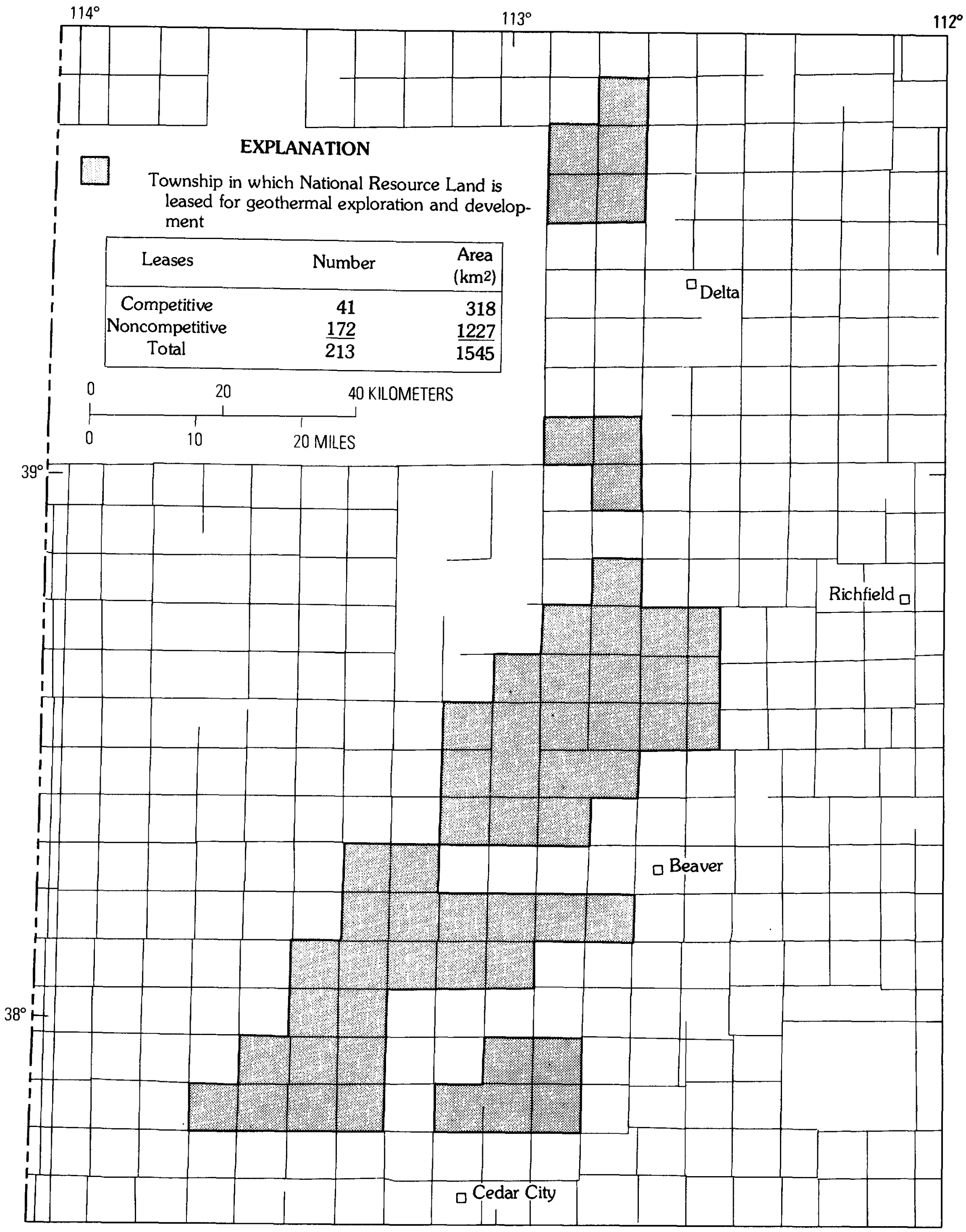


FIGURE 6.—Location of leased National Resource Land in Utah, September 1976.

TABLE 5.—*General characteristics of thermal-spring groups*

[Group data based mostly on data in Mundorff (1970) and Milligan and others (1966). For location of groups, see fig. 5]

Thermal spring groups	Location	Spring discharge rate	Dissolved-solids concentrations	Water temperature at land surface (°C)	Silica content of water (mg/L)	Spring deposits and gas discharges
Wasatch Range.	Along the major fault zone separating the Basin and Range province from the Colorado Plateaus and the Middle Rocky Mountains provinces.	Wide ranges.	Wide ranges.	20–77; hottest springs are at the south end of the area.	15–85; highest at the south end of the area.	Travertine common.
Northwest area.	In the Basin and Range province, an area of above normal conductive heat flow.	--do------	--do------	Generally low. 18–42, except for Crater Hot Springs (87°C) near south end of area.	5–33; Crater Hot Springs (57–59) near south end of area.	Travertine common; some hydrogen sulfide.
Southwest area.	--do------	--do------	--do------	Generally high, 32–85, commonly above 70.	28–400; highest at Roosevelt Hot Springs.	Siliceous sinter, travertine, and hydrogen sulfide common; sulfur deposits at Sulphurdale.

ROOSEVELT HOT SPRINGS AND THE COVE FORT-SULPHURDALE AREAS

The locations of these two adjacent KGRA's are shown in figure 1. Roosevelt Hot Springs (C-26-9)34dcS, is on the west flank of the Mineral Mountains in Beaver County, about 20 km northeast of the town of Milford. Cove Fort, (C-25-7)30, and Sulphurdale, (C-26-7)7, are about 25 km northeast of Roosevelt Hot Springs near the northeastern corner of Beaver County and on the west flank of the Tushar Mountains and the Pavant Range. The surface geology of the two areas is quite different. At Roosevelt Hot Springs, an alluvial valley lies to the west and a large Tertiary pluton of granite to the east. At Cove Fort and Sulphurdale, the dominant lithology is young basaltic lava flows, commonly of Quaternary age.

Lee (1908, p. 21) describes a silica deposit on the southwestern flank of the Mineral Mountains and about 25 km south of Roosevelt Hot Springs. He reports that cold water issues from a mound of silica 400 m in diameter. This geologic feature was not visited as part of this study, but if this is a hot-spring deposit, it may be an area of geothermal-resource potential. From the location description, the mound probably is at (C-29-10)24c.

Phillips Petroleum Co., Thermal Power Corp., and the University of Utah are actively exploring the hydrothermal system in the Roosevelt Hot Springs area. Phillips and Thermal Power, who have made many geophysical surveys, have drilled a total of nine deep exploratory wells (as of April 1977) to depths commonly less than 1,500 m. Results of well tests reportedly were very favorable, and additional exploration and development wells are planned. The University has made extensive geophysical surveys, funded by the National Science Foundation, the U.S. Department of Energy (formerly the Energy Research and Development Administration), and the U.S. Geological Survey, and the results were published in a series of reports. Because of these very extensive exploration activities, which have been continuing since the early 1970's, no additional field data were collected in this area as part of this study.

The geology of the Roosevelt Hot Springs area has been mapped by Petersen (1975), Parry, Berson, and Miller (1976, p. 22), and Liese (1957). Petersen mapped 10 lithologic units, including 4 units of hot-spring deposits. According to Patrick Muffler (written commun., 1976), the heat source is related to Pleistocene rhyolites that crop out on the west flank of the Mineral Mountains (fig. 1). Rhyolite as young as 490,000 years has been identified. Hydrothermally altered ground is common in the area. Brown (1977, p. 5) estimates the age of hydrothermally deposited opal near Roosevelt Hot Springs at roughly 350,000 years.

Basin and Range faulting controls the location of Roosevelt Hot Springs. One of the Basin and Range faults, the Dome fault, is marked by an abundance of siliceous sinter (opal) in mounds for a distance of nearly 5 km. The model shown in figure 6, with some

TABLE 6.—*Estimated reservoir temperatures derived by geothermometer formulas for springs (and one well) with temperatures greater than 50°C and silica concentrations greater than 50 mg/L*

[For formulas, see table 3. Chemical data mostly from Mundorff, 1970, table 1. Silica: First number is concentration for thermal-water sample; remaining numbers are concentrations for nearby cold water sources. Discharge temperature: First number is for thermal-water sample; remaining numbers are for cold-water sources and correspond to silica concentrations in adjoining column]

Hydrothermal source	Location	Concentrations in molality ($\times 10^{-3}$)			Concentrations in mg/L		Temperature (°C)		Formula used to compute reservoir temperature	Estimated reservoir temperature, in °C (rounded)	Remarks
		Calcium (Ca)	Sodium (Na)	Potassium (K)	Magnesium (Mg)	(SiO_2)	Discharge	Computed reservoir			
Christensen well, Newcastle, UT	(C–36–15)20bb	1.45	11.74	0.54	0.4	99 34 62	95 13 12	166 138 110 150–160	NKC Q C GM	140–170	No springs present. Difference in computed temperatures may indicate mixing of thermal and nonthermal waters.
Cove Fort-Sulphurdale	T. 25 and 26 S. R. 6 and 7 W.	----	----	----	----	----	----	----	----	200±	Estimate from Renner, White, and Williams (1975, p. 21).
Crater Hot Springs	(C–14–8)10S	8.61	35.50	1.23	[1]68	59 19 22	87 12 14	110 80 140	Q C GM	110–140	Reservoir rock may not be basalt. Differences in computed temperatures may indicate mixing of thermal and nonthermal waters.
Crystal Hot Springs	(C–4–1)11 and 12S	2.54	----	----	[1]24	73	58	120 92	Q C	90–120	Mixed water. Dissolved-solids are 1,665 mg/L.
Hatton Hot Springs	(C–22–6)35ddS	11.6	----	----	89	44 37	36 14	66 100–110	C GM	70–110	Subsurface temperature of 67°C measured in nearby shallow well. Mixed water.
Joseph Hot Springs	(C–25–4)23S	7.04	62.64	1.73	[1]36	85 50	65 12	101 170	C GM	100–170	----
Meadow Hot Springs	(C–22–6)26ccS	10.8	44.37	3.58	[1]114	47 37	41 14	69 120	C GM	70–120	Mixed water.
Monroe Hot Springs	(C–25–3)10ddS	7.01	24.06	1.25	[1]49	51 36 33	65.5 13 14	73 115	C GM	70–120	Reservoir temperature probably the same as for Red Hill Hot Springs or 100°–160°C.
Ogden Hot Springs	(B–6–1)23ccS	8.41	119.19	10.40	8	53 30	58 10	323 75 90	NKC C GM[2]	75–90	Reservoir temperature computed with formula NKC is probably too high. Dissolved solids are 8,820 mg/L.
Red Hill Hot Spring	(C–25–3)11caS	5.99	26.88	1.35	[1]34	83 36 33	75 13 14	99 160	C GM	100–160	Difference in computed temperatures may indicate mixing of thermal and nonthermal waters.
Roosevelt Hot Springs	(C–26–9)34dcS	.475	90.48	12.07	3.3	405	85	293 234 109	NKC Q S	260–290	Subsurface temperature as high as 262°C reported by Phillips Petroleum Co. Opal deposits.
Stinking Hot Springs	(B–10–3)30bbS	22.40	487.2	16.82	[1]335	53 30	51 10	75 95	C GM[2]	75–95	Dissolved solids are about 36,000 mg/L.
Thermo Hot Springs	(C–30–12)21S	2.07	15.57	1.25	9.7	108 23 49	82.5 14 14	199 115 141 170–200	NKC C Q GM	140–200	Travertine deposits indicate that temperature computed with formula NKC may be too high. Difference in computed temperatures may indicate mixing of thermal and non-thermal waters.

[1]Formula NKC not used where magnesium concentrations exceeded 10 mg/L.
[2]Based on chalcedony solubility curve.

modifications, is perhaps representative of this hydrothermal system. Recharge to the system probably occurs only in a permeable zone in the Mineral Mountains, the same general area where much of the recharge for Milford Valley originates. Nearly all the hot, saline water rising as part of the convection cell enters relatively shallow, fresh-water aquifers and then mixes with the fresh water as it flows westward toward the axis of Milford Valley (Mower and Cordova, 1974, pl. 4).

Roosevelt Hot Springs, the only thermal spring in the area, has had a very small flow during historic time, on the order of 1 L/s or less (Mundorff, 1970, p. 42). Since about the mid-1960's, the only flow to the surface has been a small seep supporting a very small area of tules. The measured temperature of the spring was 85°C in 1950. The spring was sampled at that time, and according to Mundorff (1970, p. 16), silica had a concentration of 405 mg/L and the dissolved solids were 7,040 mg/L. The dominant ions were sodium and chloride.

The estimated reservoir temperature (table 6), on the basis of geothermometer calculations and reported well temperatures, is 260°–290°C. If the reservoir temperature is dependent entirely on regional heat flow, the maximum depth of circulation would be about 6–7 km. This calculation is based on estimated reservoir temperature, a conductive heat flow of 2 HFU, mean thermal conductivity of 5×10^{-3} cal/cm/s°C, and an ambient land-surface temperature of about 10°C. If a shallow, magmatic-heat source is present, the local heat flow could be much higher and the depth to the hydrothermal reservoir would be less. The amount of heat and water discharged by the hydrothermal system under native conditions was not estimated.

No data have been collected as part of this study at the Cove Fort-Sulphurdale area for two reasons: (1) Phillips Petroleum Co., Union Oil Co. of California, and other companies are actively exploring the area, and (2) the only surface manifestations of hydrothermal activity are sulfur deposits, hydrothermally altered ground, and gaseous emissions at Sulphurdale. Mundorff (1970, p. 50) lists mine drainage, sampled by Lee (1908, p. 19–20) as having 10,810 ppm (parts per million) of dissolved solids, sulfate concentrations of 7,600 ppm, and iron concentration of 1,360 ppm.

Most of the Federal land in the KGRA was leased to Union Oil Co. of California, but no deep holes have been completed to date (1977). In 1976, Union Oil Co. failed to penetrate more than about 300 m in a hole scheduled to be drilled much deeper. Additional deep-well drilling attempts are planned.

The reservoir temperature has been estimated by Renner, White, and Williams (1975, p. 21) to be approximately 200°C. On the basis of the estimated reservoir temperature, a conductive heat flow of 2 HFU, estimated mean thermal conductivity of 5×10^{-3} cal/cm/s°C for the rock, and an ambient land-surface temperature of 10°C, the maximum depth of circulation to the hydrothermal reservoir is computed to be about 5 km. If a shallow, magmatic-heat source were present, the local geothermal gradient may be much higher and the depth of circulation could be much less.

Recharge for the area and for the hydrothermal system is believed to come from the Tushar and Pavant Ranges to the east. Ground water in the area probably flows generally westward or northwestward, and it probably is the only significant means of heat and water discharge from the hydrothermal system.

At Neels, a railroad siding about 60 km north of Roosevelt Hot Springs at (C-20-8)28b, Lee (1908, p. 32) reports that a well was drilled to a depth of 609 m, and hot water was encountered at a depth of 549 m. An entry in Lee's log of the well (1908, p. 33) states that gas under pressure was sufficient to raise 2,800 kg of drilling tools 122 m up the well bore. According to Kenneth Bull (oral commun., 1977), gas-discharging vents have been found in the young lava-flow area northwest of Cove Fort and north of the Mineral Mountains. The triangular area including Roosevelt Hot Springs and the Cove Fort-Sulphurdale area and extending northward to Neels probably has the best potential for geothermal development in Utah.

THERMO HOT SPRINGS

Thermo Hot Springs is about 50 km southwest of Roosevelt Hot Springs along the axial drainage of the northern part of the Escalante Desert (Milford Valley) at (C-30-12)21S and 28S. Schmoker (1972) interprets aeromagnetic and gravity data to indicate the general area to be underlain by a large Tertiary, intrusive pluton of tabular form having a thickness of about 8 km. The mountains to the northwest, the Shauntie Hills, are a horst dipping to the southeast under the Tertiary volcanic rocks and extending at least as far south as the Thermo Hot Springs area. Schmoker interprets Milford Valley as a graben with an alluvial thickness ranging from 760 to 1,070 m.

The Black Mountains, southeast of the springs, are mostly volcanic rocks associated with a possible caldera (Crosby, 1973) that range in age from 19 to 26 million years (Rowley, 1978). Rowley mapped a rhyolite about 3 km east of Thermo Hot Springs having an age of 10.3 million years (fig. 7). Rhyolites and other quartz-bearing volcanic rocks of Pliocene age have been mapped by Erickson (1973). Their occurrence is widespread in both the Shauntie Hills and the Black Mountains.

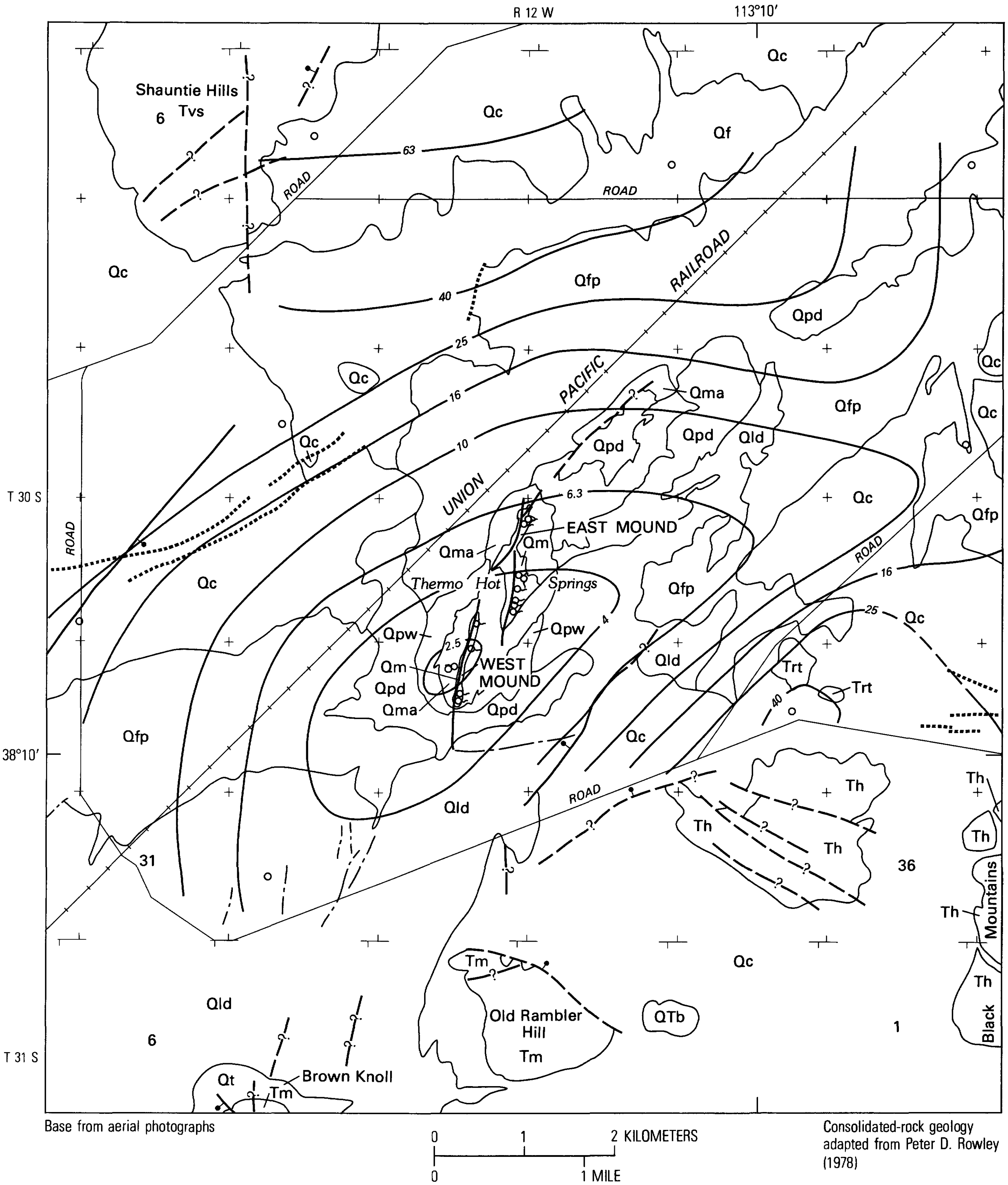
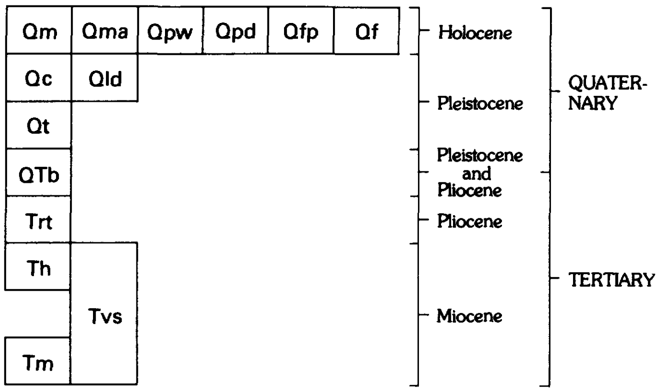


FIGURE 7.—Reconnaissance geologic map of the Thermo Hot Springs area, and audio-magnetotelluric configuration for the Thermo Hot Springs area.

CORRELATION OF MAP UNITS



DESCRIPTION OF MAP UNITS

HYDROTHERMALLY—RELATED DEPOSITS

- **Qm** Spring-mound deposits—Windblown quartz sand collecting in areas of spring discharge; mostly fine-to medium-grained, light tan. Mound surface commonly white where salt has accumulated from evaporated water. Minor amount of travertine and siliceous sinter
- **Qma** Spring-mound apron deposits—Mostly windblown quartz sand collecting in damp areas; some dark-gray basalt fragments. Sand fine- to medium-grained, tan

VALLEY—FLOOR DEPOSITS

- **Qpw** Wet-playa deposits—Light-tan, sandy silt and clay. Land surface soft due to saturation of water to the land surface. Surface white due to deposition from evaporated water
- **Qpd** Dry-playa deposits—Mostly tan sandy silt and clay. Land surface is hard most of the time
- **Qfp** Flood-plain deposits—Mostly light-tan to light-brown silt and clay; locally includes some sand. Deposits underlie hard surface marked by abundant and distinctive stream-channel meander scars
- **Qld** Dissected lake deposits—Mostly light-tan to light-brown, fine- to medium-grained quartz sand and silt; minor amounts of dark-gray volcanic rock fragments up to 2 cm in diameter locally present. Includes sand dunes that form low hills southwest of the spring mounds

ALLUVIAL—APRON DEPOSITS

- **Qf** Alluvial fan of Shauntie Hills drainage—Light-tan to light-brown silty fine- to medium-grained sand. Land surface is commonly soft
- **Qc** Colluvium—Tan to brown fine- to medium-grained sand and silt, commonly with a poorly developed gravel pavement on the land surface. Includes some gravel bars and sand dunes. Major drainage channels have dark-gray volcanic rock boulders up to 30 cm in diameter
- **Qt** Talus of Mount Dutton Formation—Accumulations of angular blocks of rock

CONSOLIDATED ROCKS

- **QTb** Basalt lava flows
- **Th** Horse Valley Formation—Rhyodacitic lava flows and volcanic mudflow breccia
- **Trt** Rhyolite of Thermo Hot Springs
- **Tvs** Volcanic rocks of Shauntie Hills—Volcanic mudflow breccia, lava flows, and ash-flow tuff
- **Tm** Mount Dutton Formation—Volcanic mudflow breccia and minor lava flows

Contact

Fault, bar and ball on downthrown side, dashed where inferred, queried where doubtful

Possible fault

Lineament

Shore line of Lake Bonneville

Hot spring

63 LINE OF EQUAL APPARENT RESISTIVITY—Line values in ohmmeters

Data point

FIGURE 7.—Continued.

Fault-controlled Thermo Hot Springs flow from two north-trending spring-controlled mounds each of which is about 1 km long and ranges from about 50 to about 200 m wide (fig. 8). The mounds, which rise about 4–8 m above the surrounding valley floor, are probably composed mostly of windblown sand and lesser amounts of siliceous sinter and travertine debris. On the spring mounds, a total of 69 spring orifices were inventoried (table 16 in table section), 54 of which were on the west mound. The total observed spring flow was estimated to be about 2 L/s and the maximum observed water temperature was 82.5°C.

Mundorff, (1970, p. 18) lists four chemical analyses of water samples from Thermo Hot Springs. The maximum silica concentration was 108 mg/L and the dissolved solids were 1,500 mg/L. The dominant ions were sodium and sulfate.

The estimated reservoir temperature (table 6), based on chemical analyses of water samples and geothermometer (table 3) calculations, is 140°–200°C. The maximum depth of circulation of water to the hydrothermal reservoir probably is between 3 and 4 km, on the basis of a regional conductive heat flow of 2 HFU, mean thermal conductivity of the rock and alluvium of about 4.5×10^{-3} cal/cm/s°C, and an ambient land-surface temperature of 12°C. The calculation assumes the absence of any shallow magmatic-heat source.

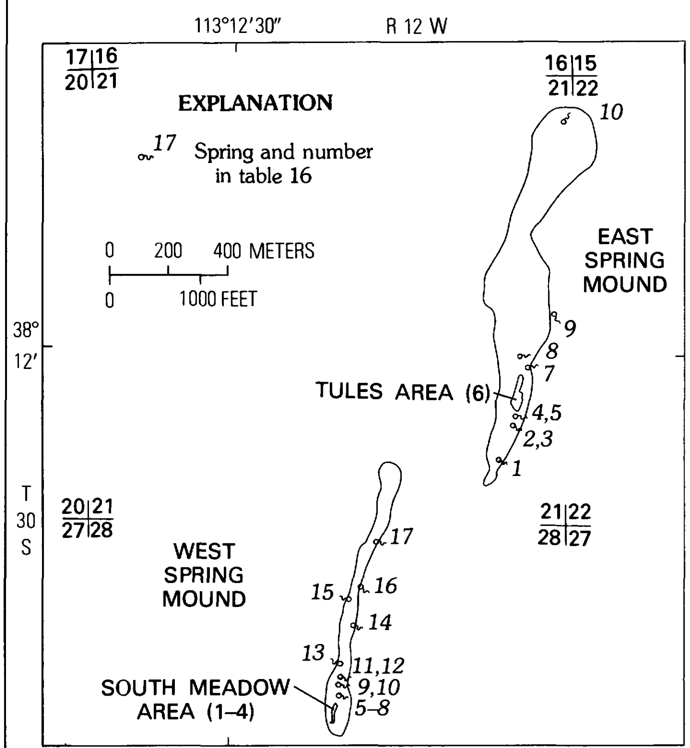


FIGURE 8.—Distribution of orifices at Thermo Hot Springs.

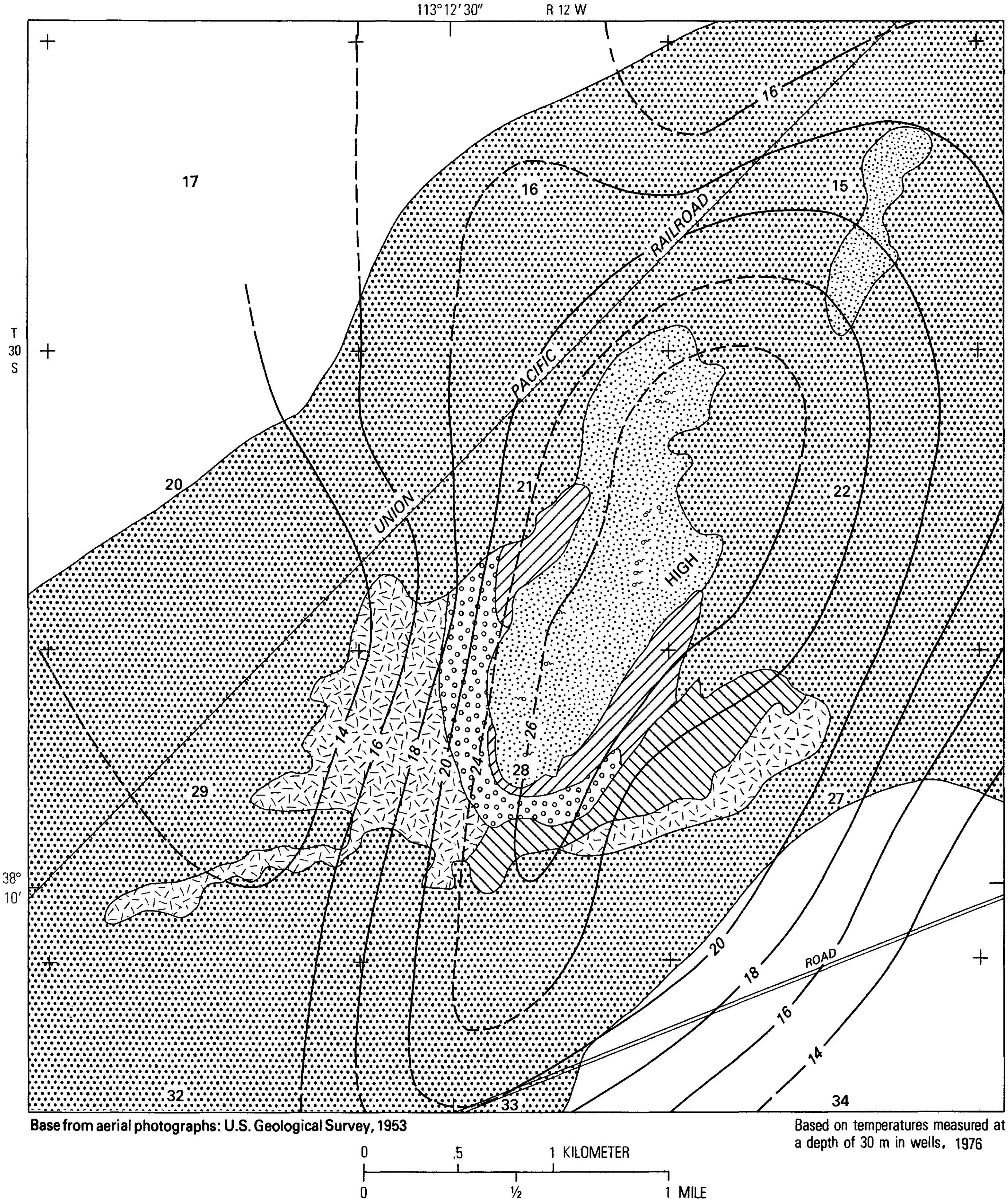


FIGURE 9.—Temperature at a depth of 30 m in the Thermo Hot Springs area; and distribution of vegetation in Thermo Hot Springs area, 1976.

Recharge for the hydrothermal system probably occurs in the nearby mountains or seeps downward through a permeable zone from the saturated alluvium of the valley. To estimate the amount of mixing of the upflowing thermal water with nonthermal water, a graphic method developed by Truesdell and Fournier (1977) was used. The amount of mixing was estimated using temperature and silica-concentration data for thermal and nonthermal ground water in the area. For nonthermal ground water, the range of silica concentrations used in the calculation was 23–49 mg/L and a water temperature of 14°C (table 6).

The results suggest that as the thermal water rises, it mixes with nonthermal water at an approximate ratio of 40 percent thermal water to 60 percent nonthermal water and enters a shallow aquifer within the alluvial valley fill. Some of this mixed water is discharged by Thermo Hot Springs, but most is discharged from a shallow water table by evapotranspiration of phreatophytes.

The distribution of the mixed thermal-nonthermal water in the alluvium is shown on three maps. Figure 7 includes an audio-magnetotelluric map which shows

DESCRIPTION OF MAP UNITS

HYDROTHERMALLY(?)—RELATED PHREATOPHYTES
Mostly saltgrass and pickleweed; locally some greasewood, rabbitbrush, and saltbush. Land surface commonly covered with salt, and soil is damp to land surface. Depth to water table generally less than 1 m. Includes tules around spring orifices and areas of ponded water

Bare playa soil except for locally scattered mounds of greasewood and saltbush. Land surface white with abundant salt east of spring mounds. Smaller salt accumulation west of mounds. Soils damp and very soft. Depth to water table generally less than 0.5 m

Mostly bare playa; locally some saltgrass and pickleweed. Land surface less salty and soils less damp than bare playa (preceding unit). Depth to water table generally less than 2 m

Greasewood and rabbitbrush mixed with lesser amounts of pickleweed and saltgrass. Depth to water table probably ranges between 1 and 3 m

Greasewood growing on a playa; land surface locally salty. Depth to water table probably ranges between 2 and 5 m

OTHER PHREATOPHYTES
Greasewood, rabbitbrush, and saltbush. Mixed with big sage and shadscale on higher parts of the area. Proportion of big sage and shadscale increases with increase in altitude. Depth to water table ranges between 3 and 15 m

XEROPHYTES
Big Sage, shadscale, and associated xerophytes growing on upland areas where depth to water table exceeds about 15 m; plants obtain moisture only from soils above water table

Hot spring

LINE OF EQUAL TEMPERATURE IN °C—Dashed where approximately located

FIGURE 9.—Continued.

the low-resistivity, elongated body of thermal water underlying the general spring-mound area; this body has its long axis parallel to the northeastward direction of ground-water flow as defined by Mower and Cordova (1974, pl. 4). Figure 9 includes temperatures measured at a depth of 30 m below land surface. In addition to the thermal anomaly near the spring mounds, two additional thermal areas are shown to the southeast and east. These two areas are probably associated with permeable fault zones separate from the permeable zones underlying the spring mounds. Contours of conductive heat flow on figure 10 show similar patterns.

Figure 9 shows the distribution of vegetation in the Thermo Hot Springs area. Three types of vegetation are shown: (1) The xerophytes in the northwestern and southeastern parts of the map that do not root to or use ground water, (2) greasewood, rabbitbrush, and saltbush that obtain much of their water from the shallow ground water in the valley fill reservoir, and (3) phreatophytes near the spring mounds that use rising thermal water at rates greater than the surrounding phreatophytes. Net mixed-water discharge by phreatophytes from the hydrothermal system is computed in table 7 as the difference between the gross discharge of the area and the estimated discharge if no hydrothermal system were present. The estimated discharge of mixed water by phreatophytes is 1.4×10^{-6} m^3/yr or equals a continuous flow of 44 L/s. The portion of this flow that is deep-circulating thermal water is about 40 percent or 18 L/s.

Olmsted and others (1975, p. 66, p. 220–224) have developed two methods of estimating conductive heat discharge from hydrothermal systems. Their method (A) is based on a subsurface-temperature map and the thermal gradient from the map's datum plane to the land surface. Method (B) is based on a heat-flow map for a hydrothermal system. Because both methods use the same data base, only method (B) is presented in table 8.

Method (B) yields a conductive heat discharge from the system of 16×10^{13} cal/yr; method (A), 15×10^{13} cal/yr. A third method was also used, based on an estimated reservoir temperature of 200°C (table 6), convective water flow through the deep hydrothermal reservoir of $0.6 \times 10^6 m^3$/yr, and an ambient land-surface temperature of 12°C. This last method yielded the lowest of the three estimates, 11×10^{13} cal/yr, and was applied only to the area of evapotranspiration near the spring mounds. In that computation of convective flow, because no increment for warm ground-water flow to the northeast from the thermal areas was included, the estimate of total heat discharge is probably too small. The value 15×10^{13} cal/yr was selected to represent the probable heat discharge from the entire hydrothermal system. As depth increases, the three heat anomalies shown in figures 9 and 10 probably expand and merge into a single hydrothermal system.

SOUTHWESTERN ESCALANTE DESERT

The southwestern part of the Escalante Desert is northwest of Cedar City (fig. 1). Three communities are referred to in this section: Newcastle ([C-36-15]17), Beryl ([C-33-16]32), and Lund ([C-32-14]21) (fig. 11). Newcastle is in the southeastern part of the desert, about 40 km west of Cedar City. Beryl is northwest of the valley axis and 30 km northwest of Newcastle. Lund is 25 km northeast of Beryl along the Union Pacific Railroad tracks.

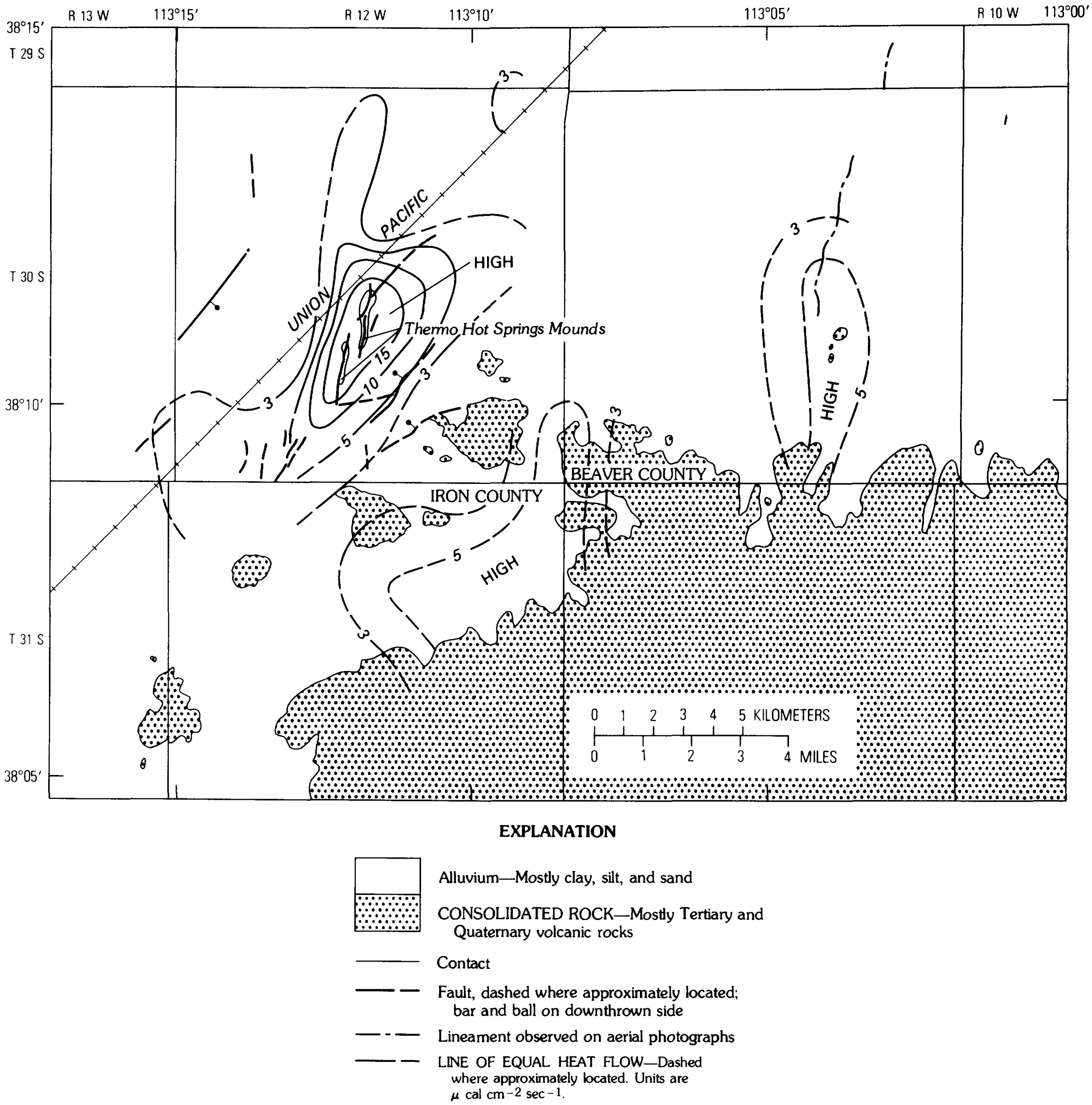


FIGURE 10.—Shallow conductive heat flow in the Thermo Hot Springs area.

TABLE 7.—*Evapotranspiration of mixed water from Thermo Hot Springs hydrothermal system*

[1976 conditions. The combined and nonhydrothermal system discharge rates are based on research by Lee (1912), White (1932), Young and Blaney (1942), Houston (1950), Robinson (1965), and Harr and Price (1972) in other areas]

Phreatophyte area	Depth to water table (m)	Average annual evapotranspiration rates (approximate): Combined discharge rate (m)	Nonhydrothermal system discharge rate (m)	Net hydrothermal discharge (m)	Area ($\times 10^3 m^2$)	Estimated average annual net discharge (rounded $\times 10^3 m^3$)
Mostly saltgrass pickleweed	<1	0.5	0.06	0.44	1,700	750
Bare playa soil	< .5	.6	.06	.54	430	230
Bare playa, saltgrass, and pickleweed	<2	.4	.06	.34	520	180
Greasewood, rabbitbrush, pickleweed, and saltgrass	1–3	.2	.06	.14	1,400	200
Greasewood on playa	2–5	.1	.06	.04	420	20
Greasewood, rabbitbrush, and saltbush	3–15	.06	.06	.0	--	0
Total (rounded)	--	--	--	[1].3	4,500	[2]1,400 ($1.4 \times 10^6 m^3/yr$)

[1]Average for area.
[2]Mixed thermal and nonthermal water.

Newcastle is on the floor of the Escalante Desert near the northwestern flank of the Pine Valley Mountains, a range composed mostly of Tertiary volcanic rocks (Hintze, 1963). The floor of the Escalante Desert is described by Crosby (1973, p. 28) as the central region of a probable large caldera about 50 km in diameter. The rim of the caldera includes the surrounding mountains and Table Butte as shown in figure 11. Crosby gives no descriptions of the age or structure of the caldera in his report, but it is probably Tertiary.

The detailed distribution of young igneous rocks in this area is poorly known. Some basic igneous rocks less than 10,000 years old lie about 40 miles south of Newcastle, near the town of Veyo (Smith and Shaw, 1975, p. 82, listed under Utah as Santa Clara). The reconnaissance geology of the area is shown in figure 12.

During December 1975, a newly drilled irrigation well (C-36-15)20bbd was test pumped at rates as high as 108 L/s. This well, owned by the Christensen Brothers of Newcastle, is 152 m deep, has a 40-cm-diameter casing, and a static water level of about 43 m. The water discharged was boiling, about 95°C at land surface (altitude = 1,605± m). A water sample was collected after 6 hours of pumping; the chemical analyses are given in table 9. The dissolved-solids concentration was only 1,120 mg/L; the silica concentration was 99 mg/L; and the dominant ions were sodium and sulfate. On January 20, 1976, after a period of several weeks during which the pump was idle, subsurface temperatures were measured in the well, resulting in the temperature profile in figure 13. The profile shows an alluvial aquifer containing hot water at a depth below land surface between 70 and 110 m. Subsurface temperatures were lower above and below this aquifer. The maximum temperature recorded in the well was 107.8°C; the bottom-hole temperature was 4.1°C less. The well was pumped during the 1977 irrigation season; the water was cooled in two ponds and applied to cropland by sprinklers.

TABLE 8.—*Estimated conductive heat discharge from Thermo Hot Springs hydrothermal system—alluvial area only*

[Method B of Olmsted and others (1975, p. 67). The approximate area was determined from figure 10]

Range in heat flow units (HFU) ($\times 10^{-6}$ cal/cm²/s	Approximate mean heat flow (HFU)	Approximate area (km²)	Heat discharge ($\times 10^5$ cal/s)
>15	20	3.3	6.6
10–15	12.5	3.3	4.1
5–10	7.5	24	18
3–5	4	50±	20±
Total (rounded)	6	80	50 (16×10^{13} cal/yr)

The estimated reservoir temperature for the hydrothermal system, based on chemical analysis and geothermometers, is 140°–170°C in table 6. The difference in the calculations in table 6 probably results from mixing of thermal and nonthermal waters. The maximum depth of circulation in the hydrothermal reservoir required to produce these temperatures is estimated to be 3–4 km, on the basis of regional heat flow of 2 HFU, thermal conductivity of the volcanic rocks that underlie the area of 5×10^{-3} cal/cm/s°C, and an ambient land-surface temperature of about 14°C. The calculation assumes the absence of a shallow magmatic-heat source. Using the graphic method of Truesdell and Fournier (1977) to estimate the mixing of thermal with nonthermal water, the hot-water component of the mixed water is estimated to be about 60 percent on the basis of data in table 6.

Shape and size of the hot-water body in the valley-fill (alluvial) aquifer are indicated in figures 12, 14, and 15, as well as the temperature profile shown in figure 13.

The recharge area for the hydrothermal system is probably in the Pine Valley Mountains southeast of the town of Newcastle. The upward flow of thermal water is through a permeable range-front fault zone 1 km southeast of the thermal well (fig. 12). As hot water flows northward from the fault zone, its temperature declines owing to conductive-heat loss, mixing with cold, shallow-aquifer water, or both. Westward flow is relatively small compared to northward flow; this probably results from higher transmissivity toward the north than toward the west. Faulting and lithologic changes

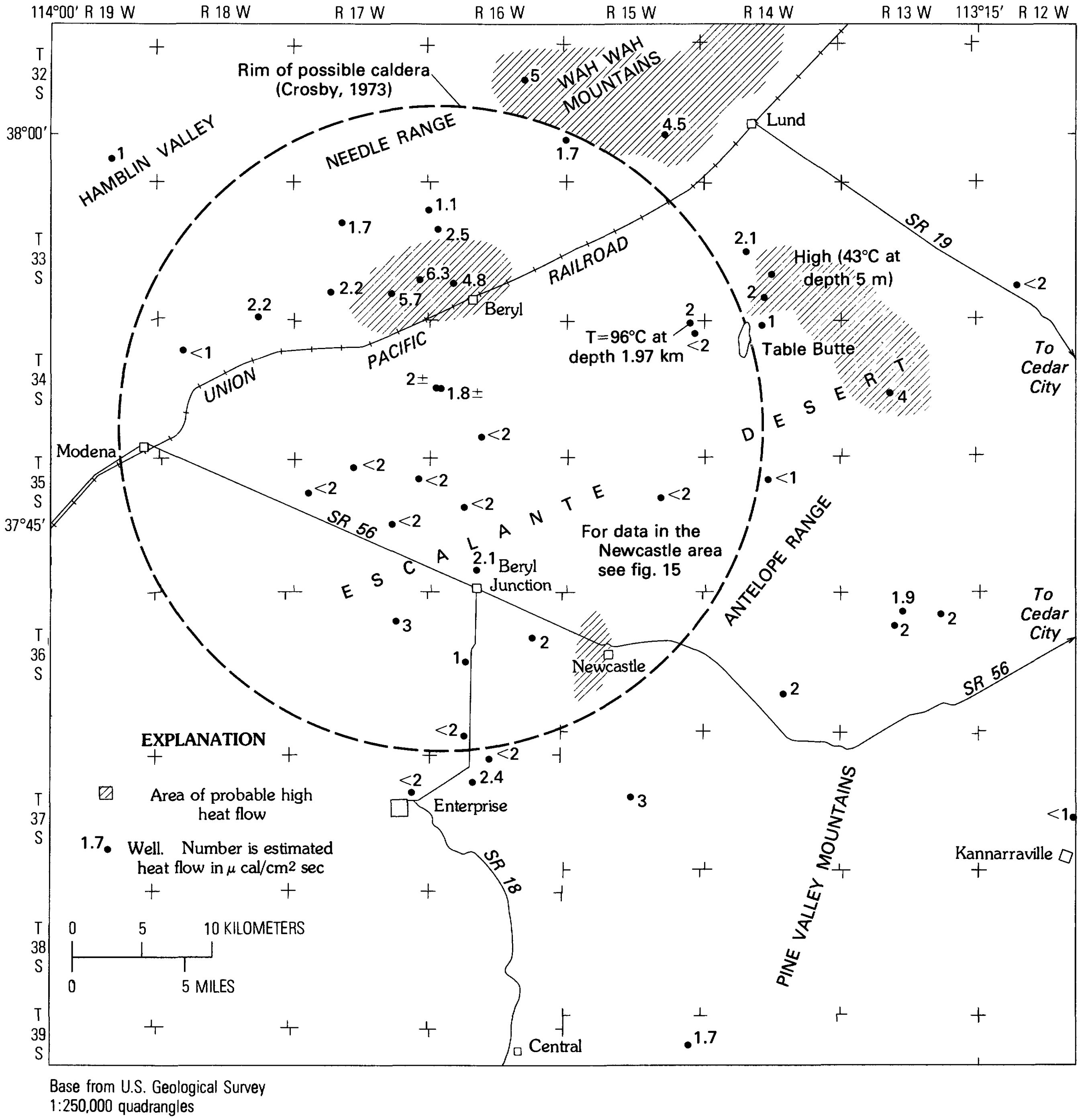


FIGURE 11.—Estimated heat flow in southwestern Utah.

TABLE 9.—*Chemical analyses of water from Christensen brothers thermal well near Newcastle, Utah*

[Location (C–36–15)20bbd. mg/L, miligrams per liter; μg/L, micrograms per liter; L/s, liters per second μmho, micromhos; °C, degrees Celsius]

Constituent	Unit	Value
Alkalinity, total (as $CaCO_3$)	mg/L	53
Aluminum, dissolved	μg/L	40
Arsenic, dissolved	μg/L	100
Barium, dissolved	μg/L	300
Bicarbonate	mg/L	64
Boron, dissolved	μg/L	710
Calcium, dissolved	mg/L	58
Carbonate	mg/L	0
Chloride, dissolved	mg/L	52
Cobalt, dissolved	μg/L	0
Fluoride, dissolved	mg/L	7.3
Hardness, noncarbonate	mg/L	95
Hardness, total	mg/L	150
Iron, dissolved	μg/L	10
Lead, dissolved	μg/L	1
Lithium, dissolved	μg/L	460
Magnesium, dissolved	mg/L	.4
Manganese, dissolved	μg/L	70
Mercury, dissolved	μg/L	.1
Molybdenum, dissolved	μg/L	13
NO_2+NO_3 as Nitrogen, dissolved	mg/L	.22
pH, field		7.6
Phosphate, Ortho, dissolved as phosphorus	mg/L	.04
Phosphate, dissolved, Ortho	mg/L	.12
Potassium, dissolved	mg/L	21
Residue, dissolved calculated sum	mg/L	1,120
Sodium absorption ratio		9.7
Selenium, dissolved	μg/L	0
Silica, dissolved	mg/L	99
Sodium, dissolved	mg/L	270
Sodium	Percent	77
Specific conductance, field	μmho	1,550
Specific conductance, laboratory	μmho	1,600
Strontium, dissolved	μg/L	1,100
Sulfate, dissolved	mg/L	580
Water temperature	°C	97.0
Yield-well	L/s	95
Zinc, dissolved	μg/L	20

TABLE 10.—*Estimated conductive heat discharge from the Newcastle hydrothermal system—alluvial area only*

[Method B of Olmsted and others (1975, p. 69). The approximate area was determined from figure 15]

Range in heat flow units ($\times 10^{-6}$ cal/cm²/s)	Geometric mean heat flow (HFU)	Approximate area (km²)	Heat discharge ($\times 10^5$ cal/s)
>30	50	1.2	6.0
20–30	24	1.0	2.4
10–20	14	3.3	4.6
5–10	7.1	7.8	5.5
3–5	3.9	9.1	3.5
Total (rounded)	[1]10	22	22 (7.0×10^{13} cal/yr)

[1]Average for area.

to the west may be factors reducing westward flow. A helium-concentration survey (fig. 16) made by Denton (1976) produced a pattern generally similar to those in figures 14 and 15 near the fault zone that produces the hot water. He concludes that the helium is released from solution in the water as a result of either temperature or pressure decline while the water flows laterally in the shallow aquifer (E. H. Denton, oral commun., 1977).

Estimates of conductive-heat discharge to the land surface were made using method (B) of Olmsted and others (1975) (table 10) yielding identical results to method (A) of 7×10^{13} cal/yr. Calculations of method (A) are not presented. Essentially all the discharge from the hydrothermal system is by lateral ground-water flow from the fault zone. The flow rate was computed on the basis of an estimated conductive-heat discharge and a reservoir temperature of 170°C. The estimated flow from the hydrothermal reservoir is about $0.4 \times 10^6 m^3$/yr. The mixed water flows through the alluvium at an estimated rate of $0.7 \times 10^6 m^3$/yr.

Figure 11 shows estimated heat flow based on measured temperature gradients for many sites in and near the southeastern part of the Escalante Desert. In addition to the Newcastle area, high heat flow is shown west of Beryl and northwest of Lund. The Lund KGRA (fig. 1) is a few kilometers southeast of Lund, but no data were available that indicated high heat flow in the KGRA. Another area east of Table Butte may have high heat flow, but data are sparse. In the irrigated area between Enterprise and Beryl, both upward and downward flow of shallow ground water may be distorting conductive heat flow to the land surface.

MONROE AND JOSEPH KGRA

The Monroe and Joseph KGRA lies along the Sevier River in Sevier County, south-central Utah (fig. 1). The Pavant Range is to the northwest, the Sevier Plateau to the southeast. The towns of Joseph ([C-25-4]14) and Monroe ([C-25-3]8 and 17) are on the flood plain of the river. Joseph Hot Springs is about 2 km southeast of Joseph (fig.17). Monroe Hot Springs is on the east edge of Monroe; Red Hill Hot Spring is about 1 km farther northeast.

Tertiary volcanic rocks are dominant in the area. Monroe and Joseph are near the north edge of the Marysville volcanic area, where volcanism was extensive and prolonged from middle to late Tertiary. The volcanic rocks range in composition from basalt to rhyolite. Smith and Shaw (1975, p. 72) cite the age of the latest eruption, a rhyolite, as 20 million years. In the following discussion of the KGRA, Joseph Hot Springs will be described separately from Monroe and Red Hill Hot Springs because superficially it is a separate hydrothermal system. Monroe and Red Hill Hot Springs are parts of a single system.

Monroe and Red Hill Hot Springs flow from travertine mounds forming a bench at the western foot of the

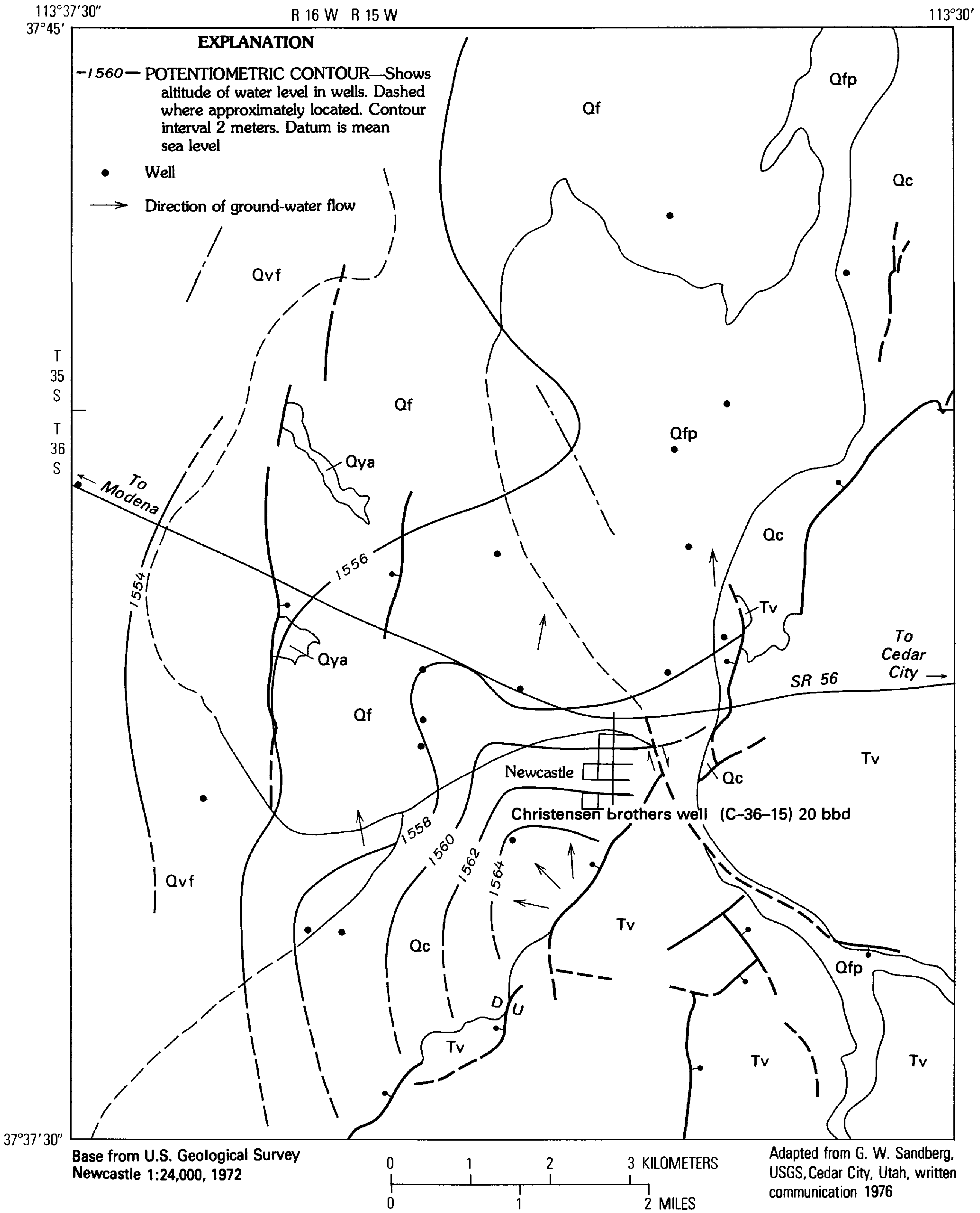


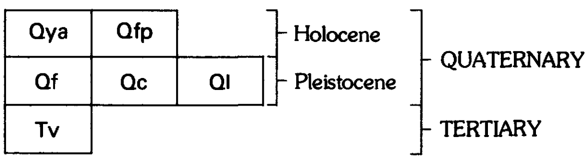
FIGURE 12.—Reconnaissance geologic map of the Newcastle area; and ground-water levels and direction of flow in the alluvium of the Newcastle area, spring 1976.

Sevier Plateau. The springs are near the Sevier Fault, as shown on figure 18. Parry, Berson, and Miller (1976, p. 67) have mapped the consolidated-rock units near the spring. The principal units are Pliocene andesite and Miocene volcanic flows and tuffs. A flat valley floor west of the spring bench is underlain by alluvium of unknown thickness (fig. 18)

As part of this study, periodic measurements were made of all spring flow and water temperatures at Monroe and Red Hill Hot Springs (table 11). From March 1976 to March 1977 total flow varied from 11.0 to 21.1 L/s and temperature ranged from 72°–75°C. The cause of the variation is not understood because data were collected for only 1 year, probably a period too short to establish trends and causes of the trends. The annual spring flow was about 0.5 × $10^6 m^3$. A small amount of additional water, about 0.05 × $10^6 m^3$/yr, is estimated to be discharged by evapotranspiration by phreatophytes on the travertine bench. The flow of Red Hill Hot Spring, the largest spring in the complex, was at its maximum during October through March of the period of record. Generally, this was also the period of lower measured temperatures of the spring. This fact suggests that flow from the spring probably is a mixed water and that during the period of high flow the proportion of nonthermal water in the mixture is larger than during periods of low flow when the spring has higher temperatures. On the basis of the graphic method of estimating hot-water component in a mixed water (Truesdell and Fournier, 1977) and data in table 6, it is concluded that thermal water probably is mixing with an equal amount of nonthermal water between the point where it leaves the hydrothermal reservoir and the springs. The thermal-water component in the mixed water that is discharged by the springs was about 0.2 × $10^6 m^3$/yr.

Samples of water from Red Hill and Monroe Hot Springs (Mundorff, 1970, p. 16) had sodium, sulfate, and chloride as the dominant ions as well as similar dissolved-solids concentrations of 2,630 mg/L and 2,700 mg/L, respectively. However, Red Hill had a silica concentration of 83 mg/L compared to only 51 mg/L for Monroe Hot Springs. Farther south on the Sevier fault, Johnson Warm Springs (fig. 17), whose water chemistry is much different than the other two springs, has dissolved solids of only 428 mg/L.

CORRELATION OF MAP UNITS



DESCRIPTION OF MAP UNITS

Qya YOUNGER ALLUVIUM—Light-tan sandy silt and clay deposited in nearly horizontal, playalike areas. Land surface is hard and commonly has desiccation cracks. Deposit is thin and unsaturated.

Qfp FLOOD DEPOSITS OF PINTO CREEK—Light-tan to light-brown sandy silt deposited by Pinto Creek on its alluvial fan during frequent flood flows. Sand is mostly fine-grained quartz and volcanic-rock fragments. Material is mostly reworked Qf, described below. Forms moderately hard land surface with some desiccation cracks. Deposit is thin and generally unsaturated by ground water. Includes sand and gravel deposits along Pinto Creek in the mountains.

Qf ALLUVIAL-FAN DEPOSITS OF PINTO CREEK—Mostly light-brown fine- to medium-grained silty sand and sandy silt deposited by Pinto Creek. Sand is mostly quartz and volcanic-rock fragments. Material is derived mostly from volcanic rocks of the Pinto Creek drainage basin in the Pine Valley Mountains. Forms moderately soft land surface. Depth to ground-water saturation ranges from 15 m to 30 m beneath the land surface.

Qc COLLUVIUM—Mostly medium- to dark-gray poorly sorted silt, sand, gravel, and boulders derived from the adjacent volcanic rocks of the mountains. Underlie a generally steep-sloping apron. Depth to ground-water saturation is generally greater than 30 m.

Qvf LAKE BONNEVILLE SEDIMENTS—Mostly light-tan sandy silt. Largely undissected. Land surface generally hard. Depth to ground-water saturation is generally less than 30 m.

Tv VOLCANIC ROCKS—Generally dark-gray basalt; commonly vesicular.

Contact; dashed where approximately located

Fault; dashed where inferred. Bar and ball on downthrown side

Lineament observed on aerial photographs

FIGURE 12.—Continued.

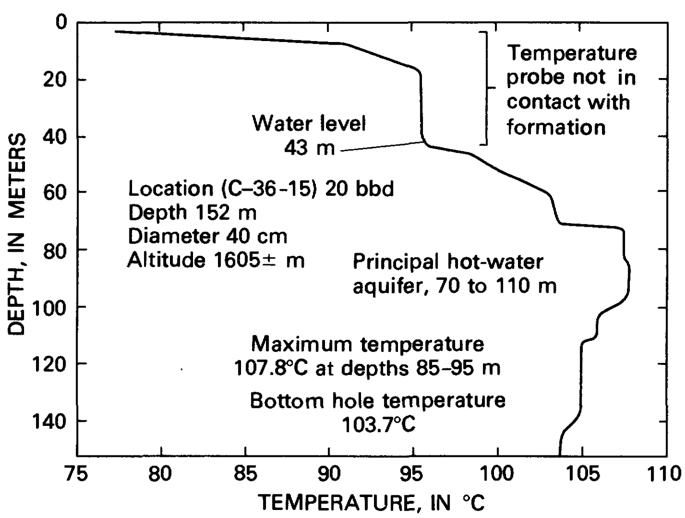


FIGURE 13.—Temperature profile of the Christensen Brothers thermal irrigation well near Newcastle, Utah.

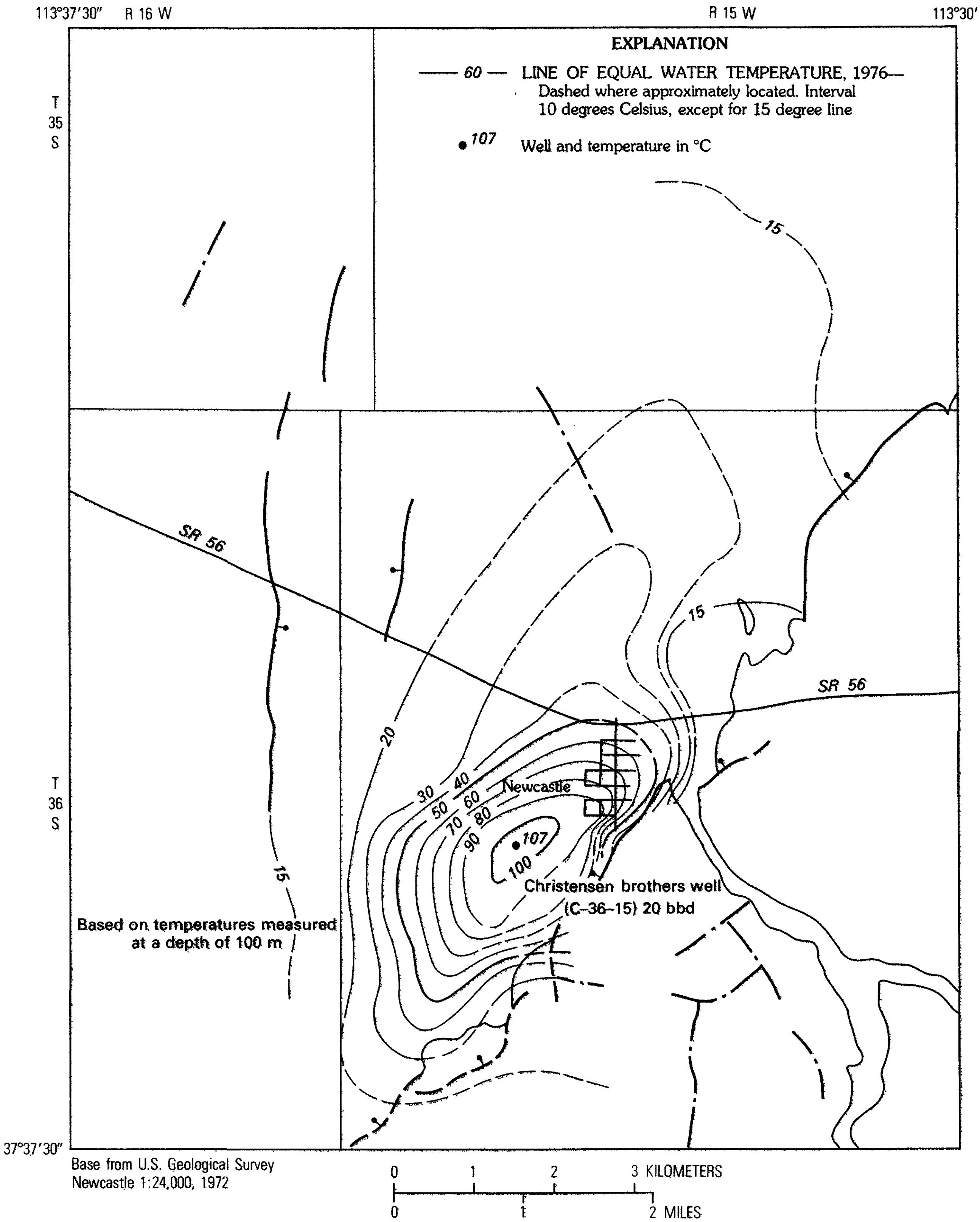


FIGURE 14.—Temperature at a depth of 100 m in the Newcastle area.

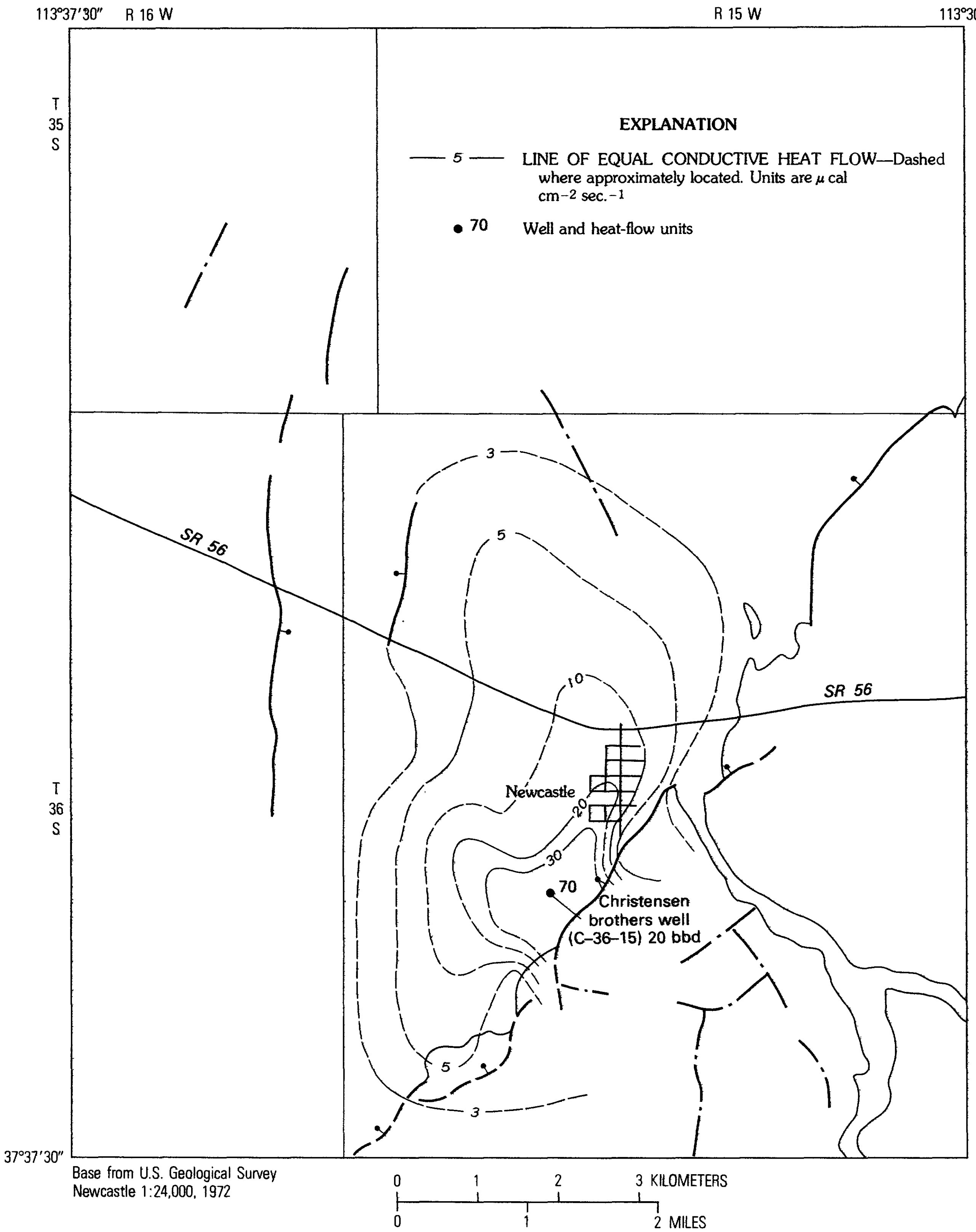


FIGURE 15.—Distribution of heat flow from the principal hot-water aquifer in the Newcastle area.

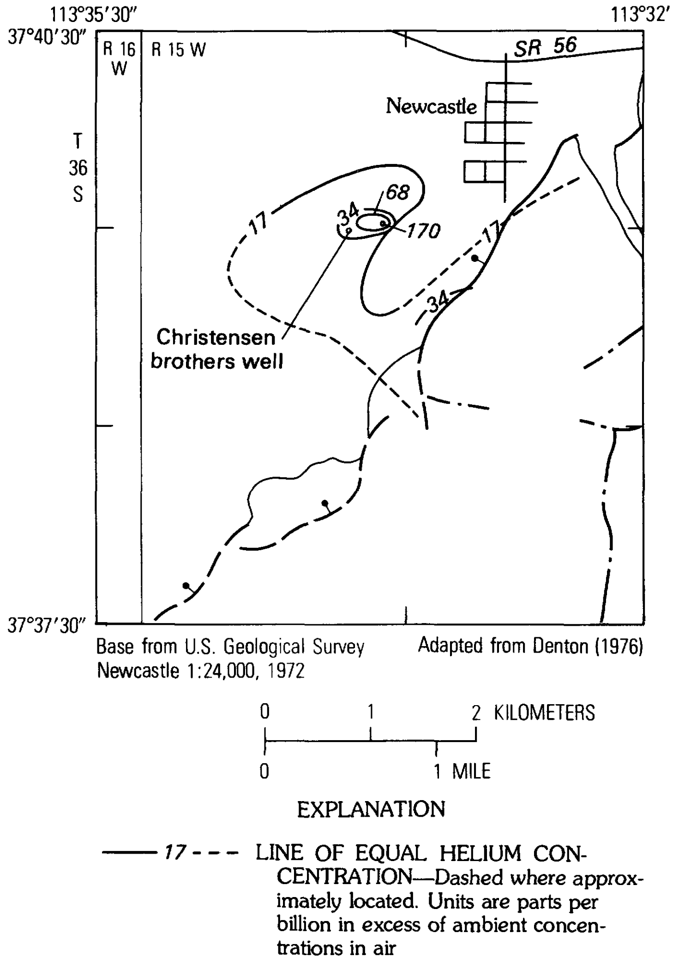


FIGURE 16.—Helium concentrations in the Newcastle area at a depth of 0.6 m below land surface.

The estimated reservoir temperature for the Monroe-Red Hill Hot Springs hydrothermal system is 100°–160°C, as determined from data in table 6. The maximum depth of circulation to the hydrothermal reservoir, in order to produce these temperatures, is estimated to be about 2–4 km on the basis of a regional conductive heat flow of 2 HFU, thermal conductivity of the rock of about 5×10^{-3} cal/cm/s°C, and an ambient land-surface temperature of 12°C. The calculation assumes the absence of a shallow magmatic heat source.

The hydrothermal system probably is recharged on the Sevier Plateau, and thermal water circulates upward through a permeable fracture zone associated with the Sevier fault. In addition to the spring flow and evapotranspiration, discharge of mixed water may include an undetermined amount of lateral subsurface flow from the fault zone. Figure 17 shows a few data points for heat flow and the area of probable above-normal regional heat flow confined to the vicinity of the fault.

The estimated heat discharge by spring flow and evapotranspiration is 3×10^{13} cal/yr. This estimate is based on a waterflow from the hydrothermal reservoir of a minimum of $0.2 \times 10^6 m^3$/yr (50 percent of spring and evapotranspiration discharge) and an estimated maximum reservoir temperature of 160°C. Lateral subsurface flow from the fault zone would discharge additional heat from the system.

Joseph Hot Springs is 8 km southwest and across a low unnamed mountain ridge from Monroe Hot Springs (fig. 17). The general geologic and topographic settings are similar to those at Monroe Hot Springs. The springs flow from a travertine bench at the western foot of the

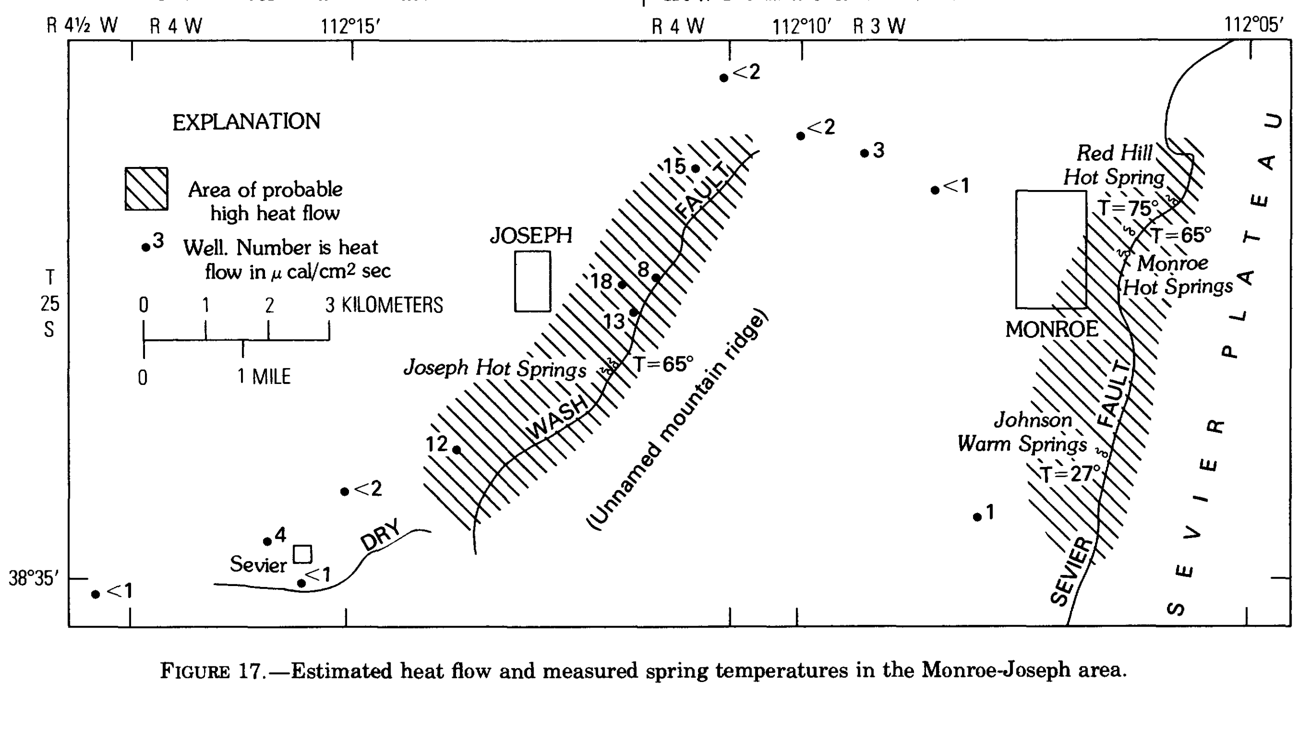


FIGURE 17.—Estimated heat flow and measured spring temperatures in the Monroe-Joseph area.

TABLE 11.—*Measured flow and temperature of Monroe and Red Hill Hot Springs*

[Locations of sites shown on figure 18]

Date	Sites																		Total flow (L/s)
	Red Hill		Tunnel		1		2		3		4		5		6		7		
	L/s	°C	L/s	°C	L/s	°C	L/s	°C	L/s	°C	L/s	°C	L/s	°C	L/s	°C	L/s	°C	
3–16–76	5.7	----	0.6	----	0.3	----	0.8	----	0.4	----	0.1	----	0.1	----	4.1	----	0.2	----	12.3
4–13–76	5.7	75	1.0	45	.2	40	.4	56	.4	50	.1	40	.1	----	4.4	47	.0	----	12.3
5–11–76	5.6	75	.2	43	.1	38	.4	66	.3	48	.2	38	.1	----	4.1	48	.0	----	11.0
6– 7–76	6.2	75	.5	45	.1	40	.4	65	.4	50	.2	40	.0	----	4.7	48	----	----	12.5
7–21–76	6.6	75	<0.1	44.5	.1	39.5	.4	65.5	.4	50	.2	40.5	.0	----	4.7	48	.0	----	12.4
8–24–76	6.6	75	< .1	45	.1	40	.4	65	.4	50	.2	40	.1	----	4.6	48	.0	----	12.4
10– 5–76	13.6	75	< .1	45	.1	40	.4	65.5	.4	48	.2	40	.0	----	4.7	48.5	.0	----	19.4
12–27–76	15.2	74	< .1	36	.5	40	.5	60	.5	45	.1	32	.0	----	4.0	45	.0	----	20.8
2– 1–77	15.2	72	< .1	37	.4	39	.4	60	.4	46	.1	30	----	----	4.6	35	.0	----	21.1
3–26–77	11.2	73	< .1	36	.4	39	.3	60	.4	41	.1	30	----	----	4.2	44	.0	----	16.6

ridge. The permeable zone through which hot water flows upward is the Dry Wash Fault zone. The narrow flood plain of the Sevier River is west of the fault. Table 12 lists the periodic measurements of all flow from Joseph Hot Springs. The flow ranges from 1.3 to 2.7 L/s or an annual flow of about $0.07 \times 10^6 m^3/yr$. During the year that measurements were made, flow was generally declining for unknown reasons, but no pattern was observed in any relationship between flow and temperature as was observed at the Monroe-Red Hill Hot Springs complex. The maximum observed temperature of the springs was 65°C. In addition to the spring flow, a small amount of water, about $0.01 \times 10^6 m^3/yr$, is estimated to be discharged by evapotranspiration of phreatophytes on the travertine bench.

Chemical analyses of samples from Joseph Hot Springs (Mundorff, 1970, p. 16) show that the highest silica concentration was 85 mg/L. That sample had a dissolved-solids concentration of 5,150 mg/L. The dominant ions were sodium and chloride. The estimated reservoir temperature (table 6) and depth of circulation of thermal waters are about the same as for the Monroe-Red Hill Hot Springs system.

Recharge for the hydrothermal system may originate as precipitation in the nearby highlands or come from saturated alluvium underlying the Sevier River flood plain. During upward circulation from the hydrothermal reservoir through the permeable fault zone, the thermal water is estimated to mix with nonthermal water in the proportion 35–65 percent, respectively.

TABLE 12.—*Measured flow and temperature of Joseph Hot Springs*

[Sites 1–12 are progressively farther north, with 1 the southernmost and 12 the northernmost]

Date	Sites											
	1		2		3		4		5		6	
	L/s	°C	L/s	°C	L/s	°C	L/s	°C	L/s	°C	L/s	°C
3–16–76	0.8	----	0.1	----	<0.1	----	0.1	----	0.3	----	0.5	----
4–14–76	.4	58	< .1	47	.1	50	.1	65	.3	65	.5	61
5–11–76	.2	47	.1	47	.1	49	.2	65	.3	64	.6	60
6– 7–76	.2	56	.1	48	.1	49	.2	64	.2	64	.7	60
7–21–76	.2	56	.1	47	.1	50	.2	64	.3	64.5	.6	60
8–26–76	.2	56	.1	47.5	.1	50	.2	64.5	.3	64.5	.7	60
10–26–76	.2	56	.1	48	.1	49.5	.2	64	.3	64.5	.7	60
12–27–76	.2	52	< .1	48	< .1	48	.1	64	.3	55	.4	62
2– 1–77	.2	45	< .1	46	< .1	48	.1	62	.4	57	.4	59
3–26–77	.2	44	< .1	45	< .1	49	.1	61.5	.2	61	.2	60

Date	Sites													Total flow (L/s)
	7		8			9		10		11		12		
	L/s	°C	L/s	°C	Fence line	L/s	°C	L/s	°C	L/s	°C	L/s	°C	
3–16–76	0.1	----	<0.1	----		0.1	----	0.4	----	0.2	----	0.2	----	2.7
4–14–76	.2	52	.1	61		.2	59	.4	42	.2	46	.2	37	2.7
5–11–76	.2	47	.1	59.5		.2	58	.3	43	.2	59.5	.1	38.5	2.6
6– 7–76	.2	48	< .1	61		.2	59	.2	43	.2	59	.1	38.5	2.4
7–21–76	.2	47.5	< .1	61		.2	59	.2	42.5	.2	59.5	.1	39	2.4
8–26–76	.2	48	< .1	60.5		.2	60	.2	42.5	.2	60	.1	38.5	2.4
10–26–76	.2	47.5	< .1	61		.2	59	.2	43	.2	59.5	.1	39	2.5
12–27–76	< .1	47	< .1	60		< .1	58	.5	43	.2	60	.1	38	1.8
2– 1–77	< .1	45	< .1	59		.1	58	.3	41	.2	59	.1	38	1.8
3–26–77	< .1	48	< .1	58.5		< .1	50	.3	44	.2	45	.1	33	1.3

Therefore, only about one-third of the earlier reported rates of spring and evapotranspiration discharge would have circulated through the hydrothermal system. An additional unknown amount of thermal water is being discharged into the alluvium from the fault zone.

The heat-flow values on figure 17, ranging from 8 to 18 HFU, probably are the result of high temperature gradients in the shallow ground-water system. Beneath the alluvial aquifers containing this shallow lateral flow, conductive heat flow toward the land surface probably is much less, such as is illustrated by the temperature profile in the Christensen Brothers well near Newcastle (fig. 13).

The estimated heat discharge by spring flow and evapotranspiration is 0.3×10^{13} cal/yr on the basis of a cooling of thermal water from a temperature as high as 170°C to an ambient land-surface temperature of 12°C. Lateral subsurface flow of mixed water from the fault zone to alluvium would discharge additional heat from the system.

CRATER HOT SPRINGS

Crater Hot Springs, also known as Baker Hot Springs and Abraham Hot Springs, is about 30 km northwest of Delta, Utah (fig. 1), at (C-14-8)10S. The springs and Crater Bench, a nearby, young basalt flow, are on the northwestern part of the broad, nearly flat alluvial floor of the Sevier Desert. The springs flow from a low travertine and alluvial mound a few hundred meters east of the lava flow. The altitude of the land surface at the spring mound is about 1,410 m above sea level.

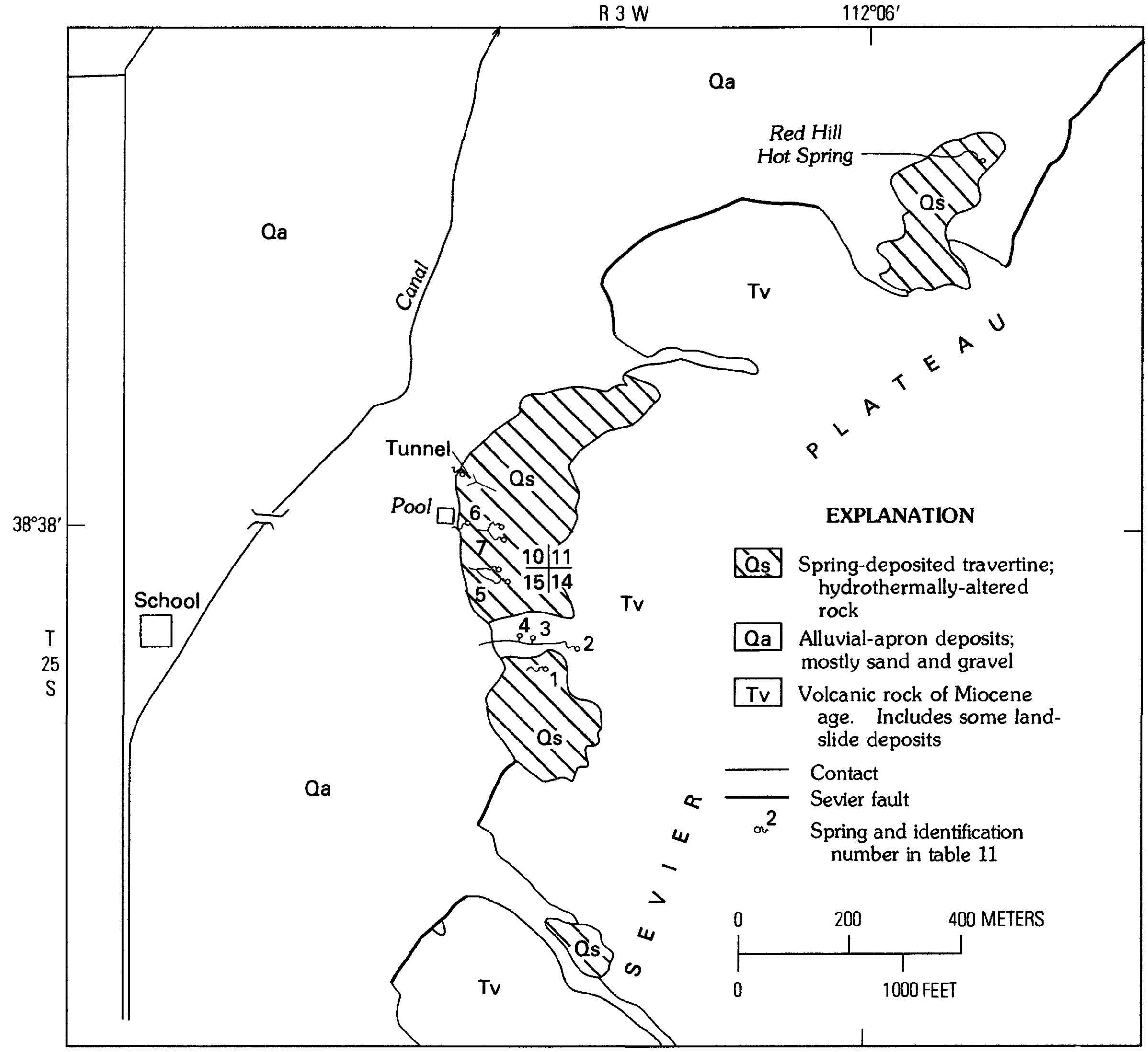


FIGURE 18.—Surficial geology at Monroe and Red Hill Hot Springs.

Crater Bench is generally oval in shape, 16 km by 10 km. The high point on the bench is at the top of Fumarole Butte (fig. 19), the neck of the volcanic conduit, having an altitude of 1,609 m. The butte takes its name from gas vents that were reported to have been active during historic time. The altitude of the surrounding area of the bench generally ranges from 1,460 to 1,520 m above sea level; therefore, the upper surface of the lava flow generally ranges from about 50 to 100 m higher than the springs. At the northeast end of the bench, rhyolite, possibly Pliocene in age, has been mapped by Galyardt and Rush (1979). The remainder of the bench is probably either Pleistocene (<2–3 million years old) or late Pliocene in age. The bench is cut by a series of northeast-trending normal faults.

Crater Bench lies along the south edge of an east-west line of three calderas (fig. 19) described by Shawe (1972). The last eruption of the calderas is reported to consist of late Tertiary and Quaternary(?) basalt and rhyolite. Silicic-rock ages as young as 3.4±0.2 million years were reported by Shawe. Although the relation of a controlling fault of Crater Hot Springs to arcuate ring faults associated with magma-chamber roof collapse is unknown, the controlling faults and ring faults are proximal in location and age.

As part of the present study, the surface lithology of the Crater Hot Springs area was mapped (fig. 20). 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. A gravity map of the area (Smith, 1974) (fig. 20) shows a high-gravity anomaly beneath the spring area that extends northwestward. The contour pattern is interpreted as possibly being caused by a shallow body of volcanic rock or hot-water deposits of relatively high density. The location of Smith's gravity anomaly suggests that any controlling fault for Crater Hot Springs may extend beneath Crater Bench and the spring area, as shown in figure 20.

Figure 21 is a generalized cross section of Crater Bench, based on Schlumberger resistivity soundings provided by A. A. R. Zohdy (written commun., 1975). The basalt flow is pictured as resting on top of alluvial valley fill; the depth to basement rock is about 1.1 km.

An inventory was made of about 40 spring orifices (fig. 22). (See table 17.) The estimated spring flow was 90 L/s during February 1976. Some additional water seeps to and ponds on the land surface. No accurate measurement of this seepage was possible because of lack of observable flow in channels. In the table, this seepage is estimated as half the observed flow or a maximum of 45 L/s, for a total of about 140 L/s. This total flow rate probably is slightly too high because some water was "counted" twice at orifice pools R1, R2, and R3 and perhaps elsewhere (fig. 22). Each of these three pools has two orifices—one yielding water, one receiving water.

Four sites of which flows were relatively large and accurate measurements could be made were selected for periodic discharge measurements. Nearly all flow from the springs is included in the periodic measurements (table 13). For example, on February 26, 1976, the flow in the four channels was 87 L/s, compared to a total flow of 90 L/s as noted earlier during the same month. Flow was generally declining during this period, but no relation between flow rate and water temperature was observed. During the period of measurement, the flow rate averaged 69 L/s.

The largest single flow was from the main-drain orifice on the northeastern part of the mound (fig. 22). Flow was about 80 percent of the total channelized flow. The maximum observed temperature of Crater Hot Springs was 87°C (table 6).

According to Mundorff (1970, p. 14), the dominant ions are calcium, sulfate, and chloride. A water sample collected in 1958 had a silica concentration of 28 mg/L and a dissolved-solids concentration of only 1,440 mg/L. However, these concentrations represent flow from an orifice having a temperature of only 43°C, much less than the maximum. As a result, it is not known whether the sample analysis represents the hottest water.

The estimated temperature of the hydrothermal reservoir in table 6 is 110°–140°C. The computed depth of water circulation to the hydrothermal reservoir is 1.3–1.7 km, or only about 200–600 m into the bedrock underlying the alluvial valley fill. This estimate is based on a regional heat flow of 2 HFU, a mean thermal conductivity of 2.7×10^{-3}cal/cm/s°C, an ambient land-surface temperaure of 14°C, and the estimated reservoir temperatures.

The hydrothermal system may be recharged by percolation of ground water from saturated alluvium or by infiltration of precipitation in the mountains, perhaps the mountains of the calderas to the north. Upflow from the reservoir presumably is along a fault. During the upflow, the thermal water mixes with nonthermal water in the proportion: 50 percent thermal water, 50 percent nonthermal water, as computed by a graphic method of Truesdell and Fournier (1977). The mixed water flows upward to Crater Hot Springs where it flows onto the surface and supports vegetation or evaporates. Additional mixed water seeps into the shallow alluvium in the general spring area and supports phreatophytes (fig. 23); their water consumption is summarized in table 14 using the same general procedure described for Thermo Hot Springs. The total evapotranspiration of

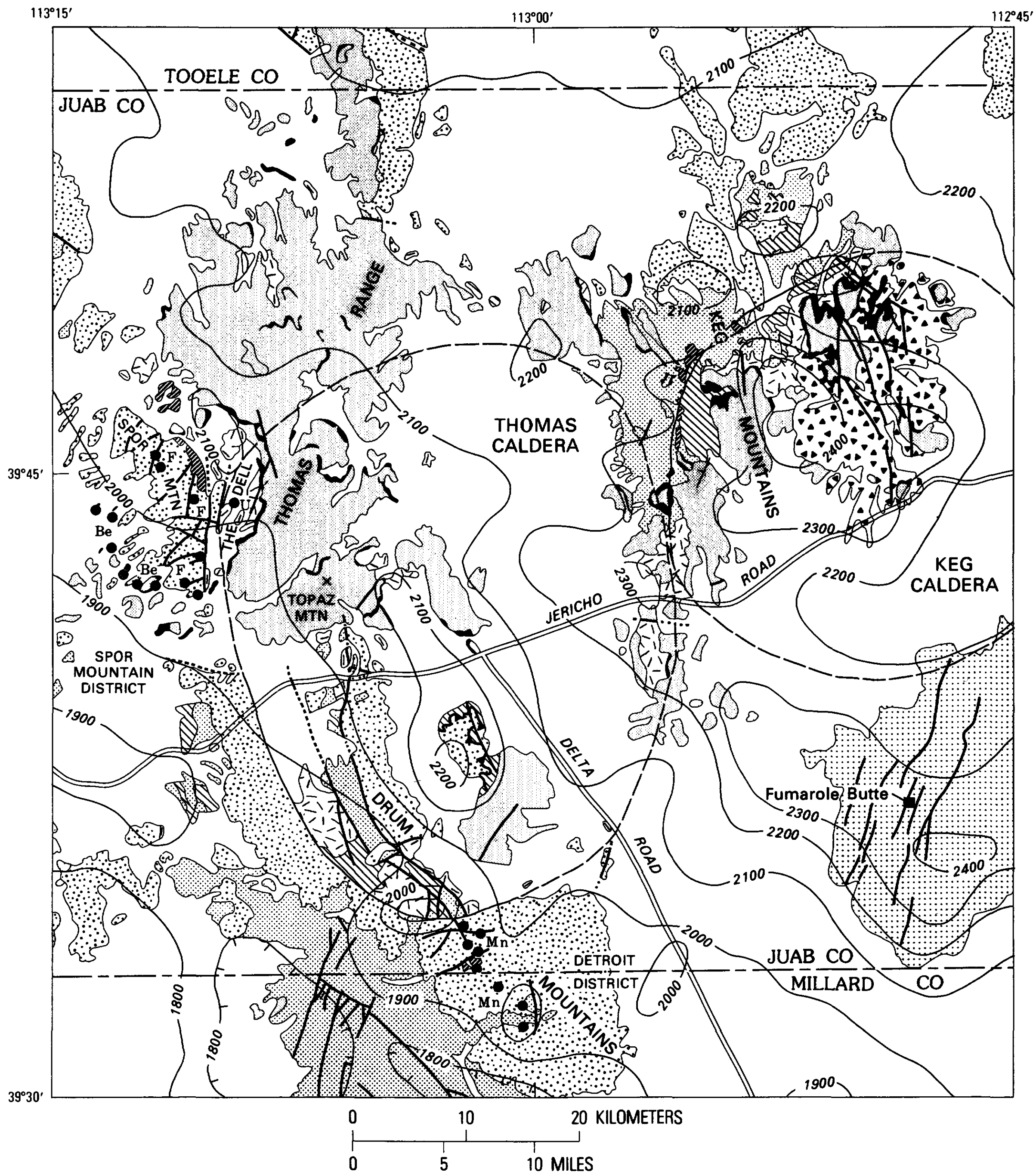


FIGURE 19.—Thomas, Keg, and Desert calderas, near Crater Hot Springs.

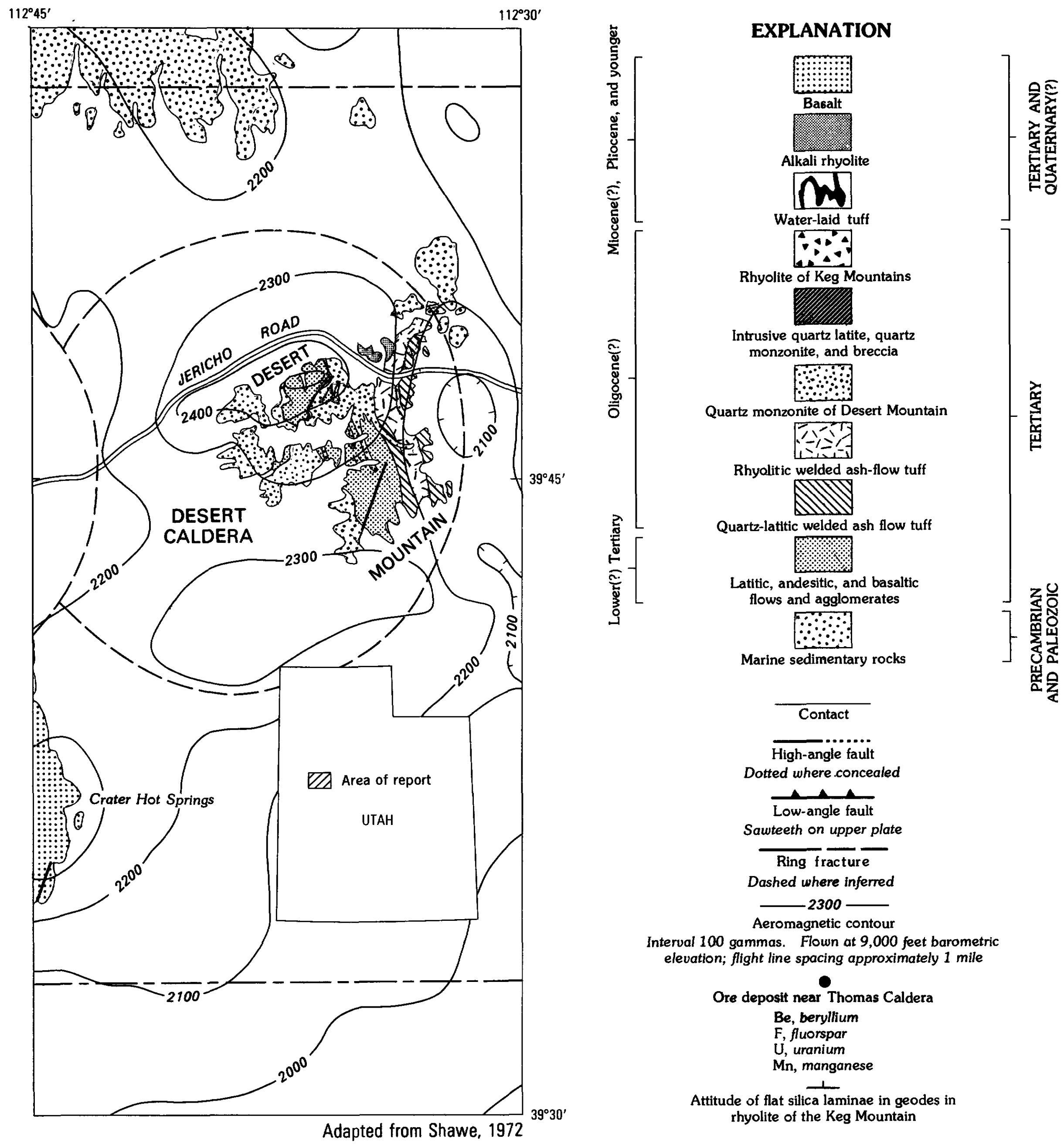


FIGURE 19.—Continued.

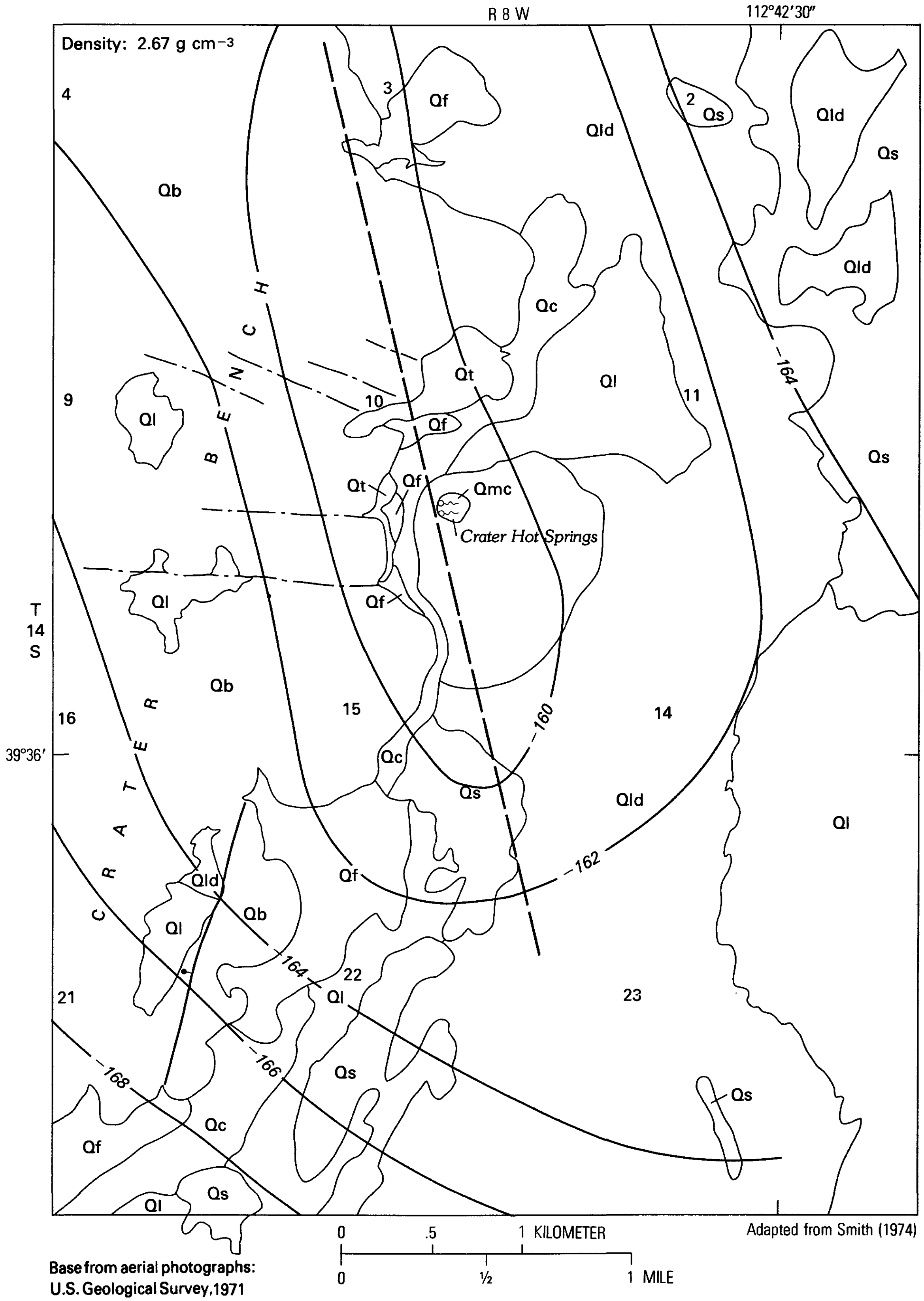
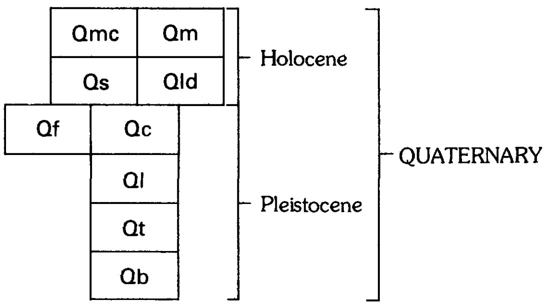


FIGURE 20.—Reconnaissance geologic map of the Crater Hot Springs area and a simple Bouguer gravity map of the Crater Hot Springs area.

CORRELATION OF MAP UNITS



DESCRIPTION OF MAP UNITS

HYDROTHERMALLY—RELATED DEPOSITS

Qmc SPRING-MOUND CREST DEPOSITS—Mostly red silt and sand deposits rich in manganese oxide. Where deposits are dry, soil is fluffy. Travertine and travertine debris locally present

Qm SPRING-MOUND DEPOSITS—Mostly light-gray silt and clay, commonly transported by wind and trapped by moist ground. Minor amounts of travertine and travertine debris on east slope of mound which has an average slope of about 20 m/km

VALLEY—FLOOR DEPOSITS

Qs SAND DUNES—Mostly small dunes of windblown, light-yellowish-brown sand partly stabilized by greasewood. Direction of sand movement is northeast

Qld DISSECTED DEPOSITS OF LAKE BONNEVILLE—Light-yellowish-gray silt and clay. Dissected by local runoff and axial drainage southward on the valley floor

Ql LAKE BONNEVILLE SEDIMENTS—Undissected, light-yellowish-gray silt and clay underlying a valley floor having an average slope southwestward of less than 1 m/km. Also present on Crater Bench in relatively low areas

ALLUVIAL—APRON DEPOSITS

Qf ALLUVIAL FAN DEPOSITS—Mostly basalt boulders and gravel in a matrix of light-yellowish-gray silt and sand below the mouth of small canyons cut into basalt flows of Crater Bench

Qc COLLUVIUM—Mixture of mostly light-yellowish-gray silt and clay with lesser amounts of slope-wash debris derived from local upslope materials. Underlies alluvial apron of intermediate slope

Qt BASALT TALUS—Angular basalt blocks eroded from Crater Bench and underlie irregular, steep slopes on the flank of the Bench

CONSOLIDATED ROCKS

Qb FUMAROLE BUTTE LAVA FLOWS—Black vesicular to massive basalt forming Crater Bench. Mostly large, angular blocks. Bench extends 30 m to 60 m higher than adjacent valley floor

Contact

Fault, bar and ball on downthrown side

Lineament of unknown origin. May be fault or fracture

Hot spring

—162— Gravity contours in milligals

FIGURE 20.—Continued.

mixed water is estimated to be $5 \times 10^6 m^3/yr$; this sum is equal to a continuous flow of 160 L/s. Additional water of unknown volume flows laterally from the area to be discharged elsewhere. On the basis of the evapotranspiration estimate above, the estimate that 50 percent of the mixed water is thermal water, a reservoir temperature of 140°C and an ambient land-surface temperature of 14°C, the convective heat discharge is 36×10^{13} cal/yr.

Heat-flow estimates made for four water wells near Crater Hot Springs are shown on figure 24. The highest heat flow was 2.8 HFU, only slightly higher than the regional heat flow.

NAVAJO LAKE KGRA

Navajo Lake KGRA is on the Markagunt Plateau, about 30 km southeast of Cedar City in southwestern Utah (fig. 1). The Markagunt Plateau, part of the Colorado Plateaus, is a horst bounded by two major fault systems—the Hurricane fault zone on the northwest and the Sevier fault on the southeast. The width of the plateau near the KGRA is about 60 km. The Navajo Lake KGRA has an area of only 10 km² (table 2). It was designated solely because of overlapping lease application on Federal land, as is required by law.

Most of the rocks exposed at the surface are Tertiary limestones (Wasatch Formation) and volcanic rocks including highly permeable Quaternary basaltic flows. Beneath these units is a thick sequence of mostly Mesozoic and older sedimentary rock, of which sandstone is the most common (Hintze, 1963). Silicic lavas were erupted, but they are generally older than the basaltic lavas. Smith and Shaw (1975, p. 82) list the largely unvegetated Markagunt basalt field as less than 10,000 years old.

No hot springs have been found in the Navajo Lake area. A well drilled at (C-36-7)33, about 10 km northeast of the KGRA had a reported temperature of drilling mud returning in the bore to land surface of 43°C with a drilling depth of 1,862 m. This mud temperature could be produced with a heat flow of less than 2 HFU. As a result, the only possible indicator of geothermal potential is the very young basaltic flows.

MEADOW AND HATTON HOT SPRINGS

Meadow and Hatton Hot Springs are 18 km southwest of Fillmore (fig. 1) and a few kilometers north of the Cove Fort-Sulphurdale KGRA (fig. 1) in southwestern Utah. The springs are at (C-22-6)27ddS and (C-22-6)35ddS, respectively, on a low alluvial spring mount in Pavant Valley, a few kilometers west of the Pavant

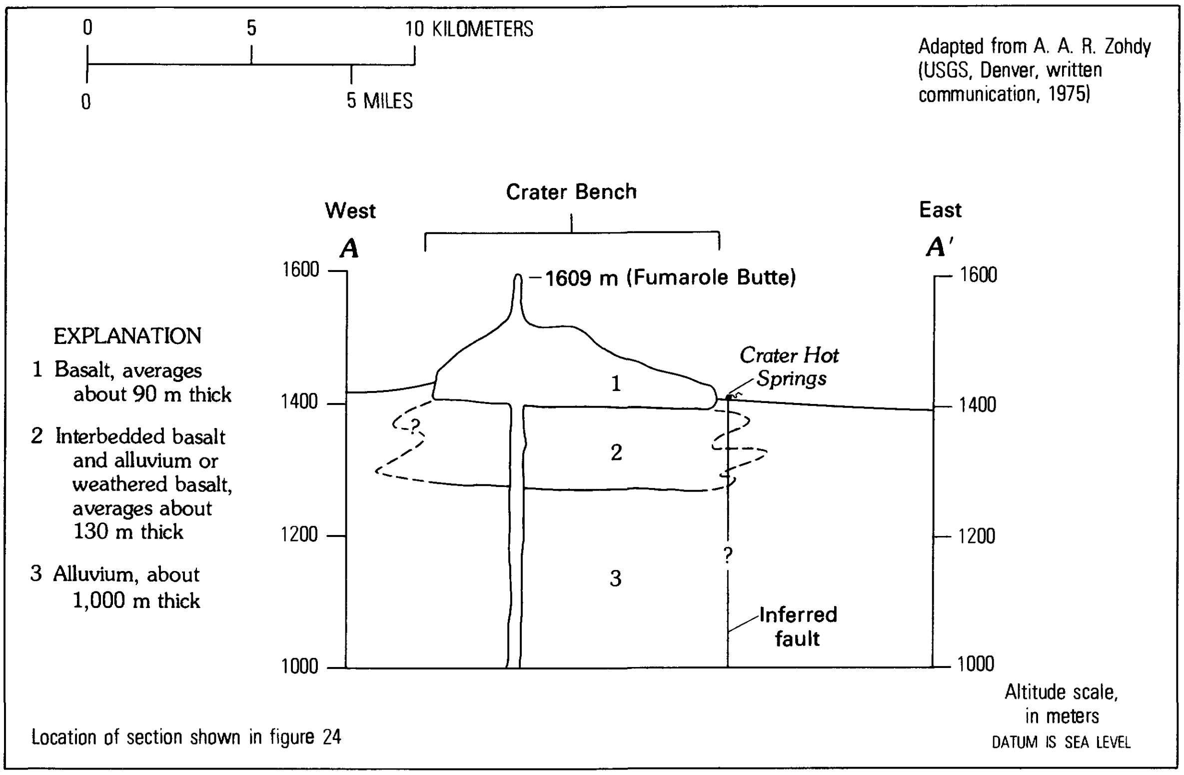


FIGURE 21.—Generalized cross section of Crater Bench.

Range. Underlying the alluvium of the spring area, volcanic rocks may be as shallow as 30 m from the land surface.

The Pavant Range has Paleozoic and Mesozoic quartzite and other sedimentary rocks thrust westward over Navajo Sandstone of Triassic (?) and Jurassic age (Hintze, 1963). North, west, and south of the springs, upper Tertiary and Quaternary basalt and basaltic andesite flows form low hills. A basaltic flow about 13 kilometers north of Hatton Hot Spring has been dated as less than 10,000 years old by Smith and Shaw (1975, p. 82, Ice Springs field).

Lithologic units are shown in figure 25. The most prominent geologic and topographic feature in the area is the spring-deposited travertine ridge at Hatton Hot Spring. The ridge is about 2 km long and rises perhaps 20 m above the general land surface. The shape and orientation of the ridge suggests fault control for the deposit-producing springs. Extending outward from the ridge is the very low alluvial spring mound (fig. 25) that is marked by many north-trending lineaments that may be faults. Travertine deposits at Meadow Hot Spring are small in volume, encircling the spring pool at the general land surface.

TABLE 13.—*Measured flow and temperature of Crater Hot Springs*

[Figure 22 shows locations of Spring E2, Spring R2, and the main-drain orifice. The Southwest ditch was measured south of Springs D1–D10 (table 17) and about 3 m northeast of concrete pools]

Date	Spring E2		Spring R2		Main-drain orifice		Southwest ditch	
	Flow (L/s)	Temperature (°C)	Flow (L/s)	Temperature (°C)	Flow (L/s)	Temperature (°C)	Flow (L/s)	Temperature (°C)
2-26-76	4.8	65	3.1	70	72.5	52	7.4	--
4- 8-76	3.7	--	4.5	--	73.6	--	8.4	--
5-19-76	4.5	66	.3	74	52.1	59	6.5	74
6- 7-76	3.6	66	.6	76	52.9	60	6.4	78
7-23-76	--	--	.0	--	60.8	--	6.5	--
9- 2-76	4.8	66	.0	--	52.3	61	5.9	70
10-13-76	3.9	65	1.1	69.5	51.8	65	6.0	70
11-15-76	7.1	64.5	1.4	69	51.5	55	6.0	68
12-30-76	4.3	68	3.9	67	50.5	55	6.0	68
2- 3-77	5.1	64	1.9	70	46.2	54	5.2	72
3-19-77	3.1	65	1.4	69.5	58.1	54	6.0	70

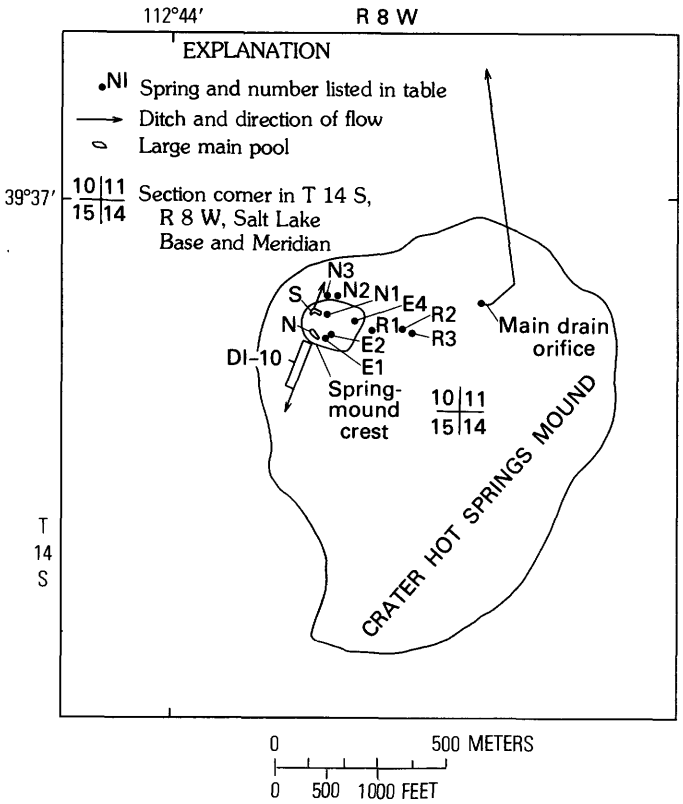


FIGURE 22.—Spring orifices and pools on the mound of Crater Hot Springs.

The observed discharge from both Meadow and Hatton Hot Springs was less than 1 L/s in the summer of 1976. Mundorff (1970, p. 40) reports a flow of about 4 L/s from Meadow Hot Spring and an absence of flow at Hatton Hot Spring during several recent years. The temperatures at these two springs, and those at other nearby sites, are shown in figure 26. The highest temperature, 67°C, was measured in a 27-m well a few tens of meters north of Hatton Hot Spring. The 67°C temperature was measured from a depth of 5 m to total depth. Hatton Hot Spring had a temperature of 36°C at that time; Meadow Hot Spring pool was 30°C. Agricultural wells, 3 km and more to the east of the spring mound, not shown on figure 26, generally have discharge temperatures of 13°C.

Mundorff (1970, p. 16) has published several chemical analyses for both Meadow and Hatton Hot Springs. He reports water temperatures for Meadow Hot Spring ranging from 29°C to 41°C. A sample collected in 1967 had a silica concentration of 47 mg/L and dissolved solids of 4,900 mg/L. The principal ions were sodium and chloride. The Hatton Hot Spring sample was collected at a temperature of 38°C, had a silica concentration of 44 mg/L, had dissolved solids of 4,670 mg/L, and had ion concentrations very similar to those of Meadow Hot Spring.

The temperature of the hydrothermal reservoir, estimated with geothermometers, is possibly in the range of 70°C to 120°C (table 6). The maximum depth of circulation required to produce the estimated reservoir temperatures with normal regional heat flow is about 2–3 km; this estimate is based on a mean thermal conductivity of the underlying rock and thin alluvium of 5×10^{-3} cal/cm/s°C and an ambient land-surface temperature of 12°C. The calculations assume the absence of a shallow magma heat source.

Recharge to the hydrothermal system probably results from percolation downward from saturated alluvium of the area or from infiltration of precipitation in the Pavant Range. During upward flow, the thermal water may be mixing with nonthermal water in the proportion of about 40 percent thermal water and 60 percent nonthermal water, as estimated by the graphic method of Truesdale and Fournier (1977). The mixed water is discharged by the springs, by evapotranspiration of pheatophytes on the spring mound, and by subsurface flow from the mound area principally to the north and west. The observed spring flow in 1976 was small, but the evapotranspiration was large. Much of

TABLE 14.—*Evapotranspiration of ground water from Crater Hot Springs hydrothermal system*

[1975 conditions. The combined and nonhydrothermal system discharge rates are based on research by Lee (1912), White (1932), Young and Blaney (1942), Houston (1950), Robinson (1965), and Harr and Price (1972) in other areas]

Phreatophyte area (See fig. 23)	Depth to water table (m)	Average annual evapotranspiration rates (approximate): Combined discharge rate (m)	Nonhydrothermal system discharge rate (m)	Net hydrothermal discharge (m)	Area ($\times 10^3 m^2$)	Estimated average annual net discharge (rounded, $\times 10^3 m^3$)
Mostly greasewood	3–15	0.07	0.06	0.01	2,200	20
Mostly bare soil, saltgrass, and picklewood	0–3	.3	.2	.1	4,000	400
Mostly bare soil of spring mound	0.4–1.5	.1	.06	.04	200	10
Meadow	.5–2	.4	.06	.34	470	160
Wet meadow	<1	.8	.06	.74	890	660
Mostly tules and very wet meadow	<.5	1.2	.1	1.1	3,400	3,700
Total (rounded)	----	----	----	----	11,000	5,000 (160 L/s)

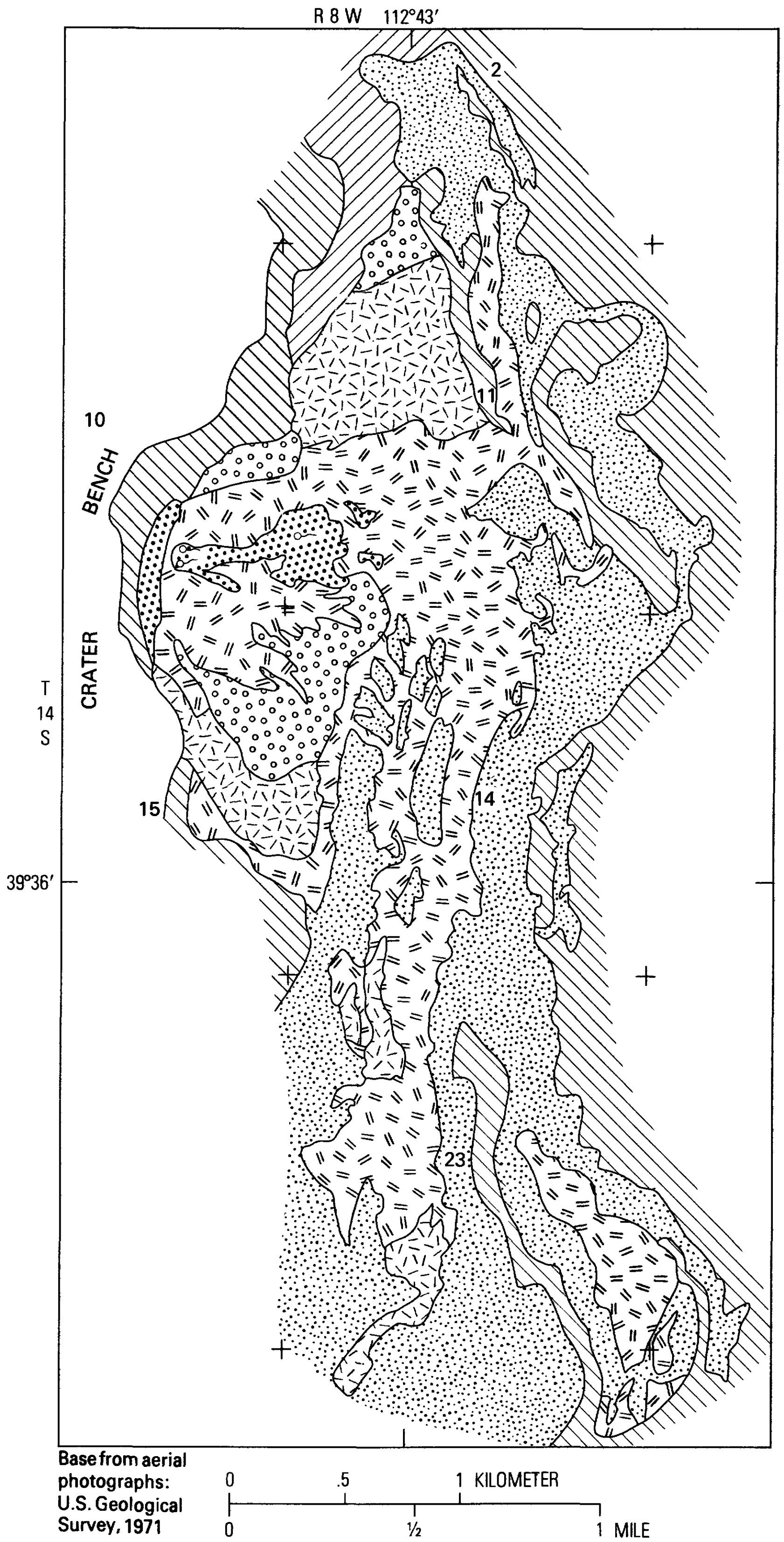


FIGURE 23.—Phreatophyte distribution in the Crater Hot Springs area.

DESCRIPTION OF MAP UNITS

Mostly bare soil of lowlands. Minor amounts of saltgrass and pickleweed locally. Discharging water probably not from Crater Hot Springs

Mostly greasewood, locally some rabbitbrush, saltbush, pickleweed, and big sage

Mostly bare soil, saltgrass, and pickleweed on lowlands. Soils are commonly moist and soft in the winter and spring and firm during the summer and fall

Mostly bare soil of spring mound; saltgrass and rabbitbrush grow locally

Meadow, saltgrass, and pickleweed on slopes of spring mound and areas north of the mound. Some cottonwood and willow of north edge of spring mound. Soils firm most of the time

Wet meadow with pickleweed and saltgrass. Soils are wet and soft during winter and spring and dryer during summer and fall

Mostly tuffs and very wet meadow on slopes of spring mound and nearby lowlands; locally some willow growth. Soils are usually soft at all times. Some lowlands may be covered with standing water during most winters and springs

Crater Bench; no phreatophytes or evapotranspiration of ground water

Hot spring

FIGURE 23.—Continued.

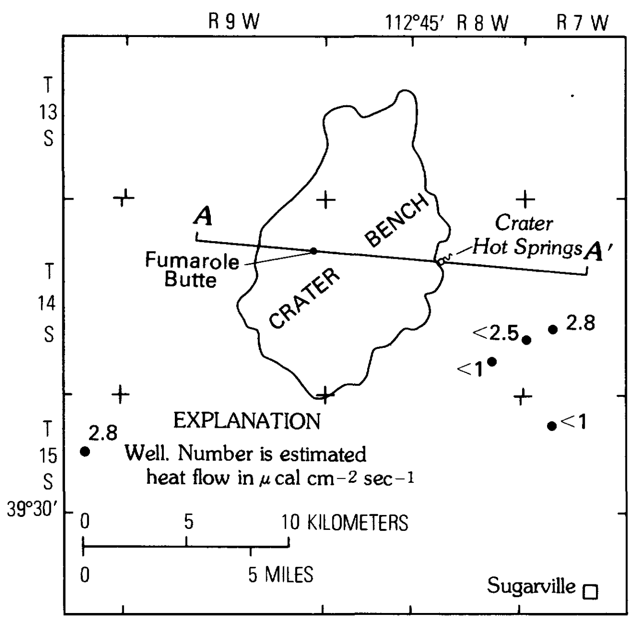


FIGURE 24.—Heat flow in the Crater Hot Springs area.

the mound area commonly is saturated to land surface during the winter and spring; at other times, the water table is at a depth commonly no greater than 2–3 m. The spring mound has an area of approximately 32 km². The net evapotranspiration rate over the mound is estimated to average on the order of 0.3 m; therefore, the estimated evapotranspiration of mixed water is about $10 \times 10^6 m^3/yr$. The rate of underflow to the west and north is unknown, but is estimated to be much less than the rate of evapotranspiration. Only 40 percent of the total discharge of mixed water is convective flow from the hydrothermal reservoir or at least $4 \times 10^6 m^3/yr$. The heat convectively discharged by this flow of water is at least 43×10^{13} cal/yr; this discharge figure is based on the cooling of the convective flow of thermal water from a possible reservoir temperature of 120°C to an ambient land-surface temperature of 12°C.

Figure 27 shows a melting pattern of freshly fallen snow near Hatton Hot Spring. According to White (1969), very high heat flow rates are required to produce such melting. Subsurface temperatures as high as 70°C can be expected at depths of as little as 10 m under these snowmelt areas.

VICINITY OF SALT LAKE CITY

Geothermal resources in the Salt Lake City area probably have no potential for electric power generation, but they are discussed here because of their potential value for uses, such as space heating, associated with urban development. Two areas of hydrothermal potential have been mapped by Marine and Price (1964). Their map has been modified as a result of additional data that became available after its compilation in 1959. The modifications involved the enlargement of the hydrothermal areas, as shown on figure 28. The map is based on discharge temperatures of wells generally drilled to depths of 200 m or less.

The northern and larger of two areas extends generally westward from the faults along the Wasatch Range front at least to the Great Salt Lake. The western and northern boundaries of the warm-water body are generally unknown; however, the map shows approximate southern and eastern limits. The warmest water in the northern area issues from Becks (B–1–1)14dcbS and Wasatch Hot Springs (B–1–1)25dbS, 56°C and 42°C, respectively. Both issue from Paleozoic limestone at the Warm Springs fault. The distribution of water temperatures suggests that most of the warm water originates from faults in the eastern part of the warm-water area, then migrates westward in the alluvium.

The hydrothermal area at the south end of Jordan Valley has similar temperatures. Crystal Hot Springs (C–4–1)11 and 12bS (fig. 28) has a reported temperature of 58°C. It, like Becks and Wasatch Hot Springs, probably flows from a permeable fault zone. The heat in the water is probably the result of deep circulation.

The warm-water areas (fig. 28) may be enlarged to a greater extent as more data become available. The southern area might be extended farther northward into the area northwest of Sandy, and the northern area may be enlarged westward and possibly northward.

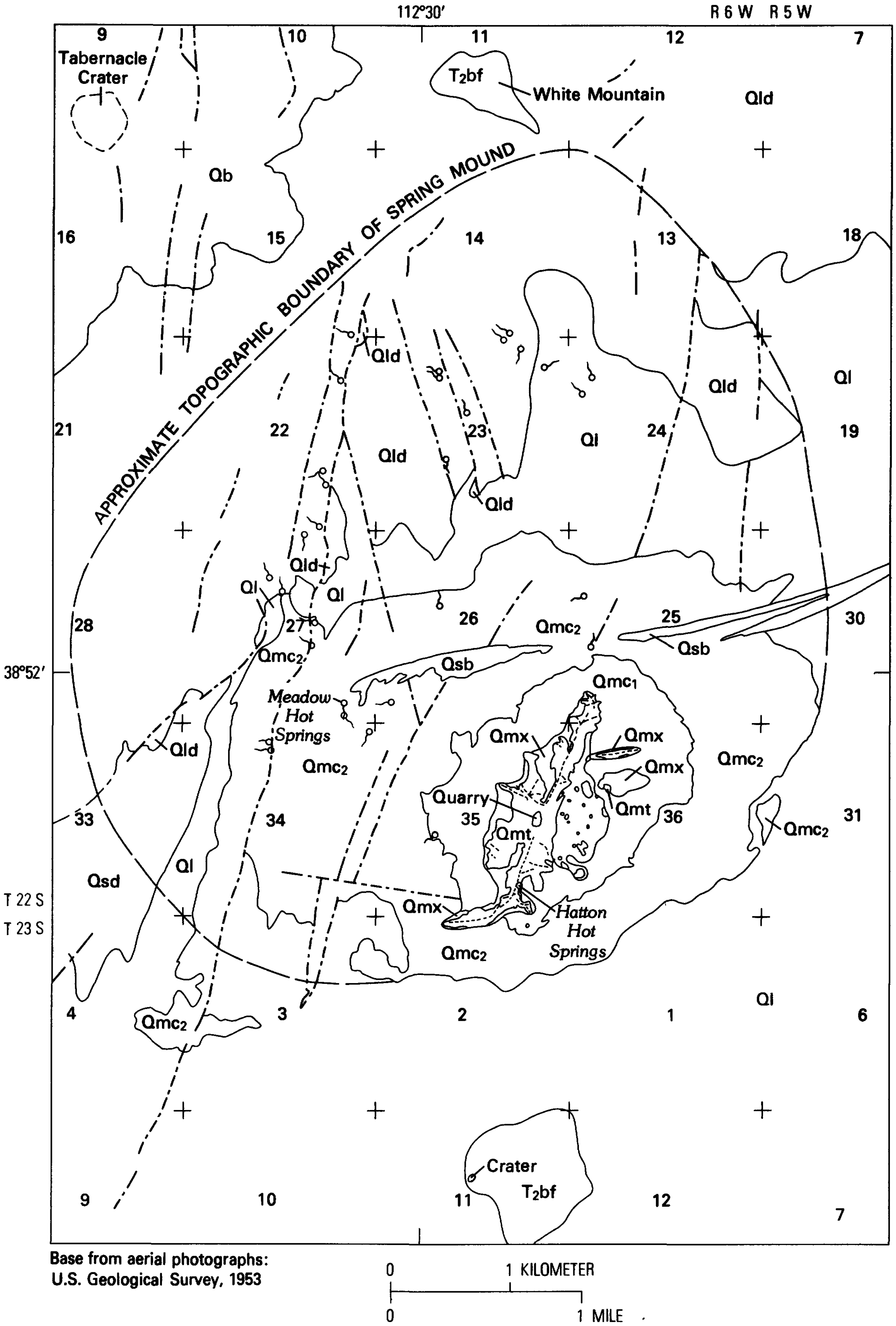
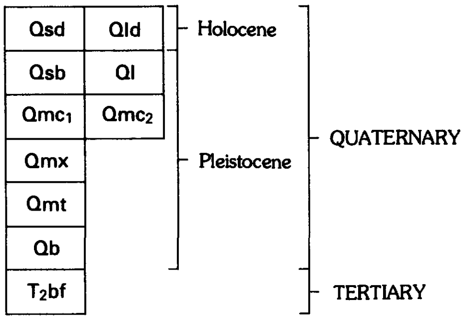


FIGURE 25.—Reconnaissance geologic map of the Meadow and Hatton Hot Springs area.

CORRELATION OF MAP UNITS



DESCRIPTION OF MAP UNITS

HYDROTHERMALLY RELATED DEPOSITS

Qmt — SPRING-MOUND TRAVERTINE—Highly porous to dense and banded calcium carbonate deposited during the cooling and evaporation of water discharged from the Hatton Hot Spring system. Commonly light tan, yellow or greenish gray. The mound of its crest extends 10 m above the adjacent valley floor

Qmx — SPRING-MOUND TALUS—Mostly angular fragments of travertine forming the flank of the spring mound, resulting from the disintegration of Qmt, described above. Mostly sand and gravel size; some larger blocks

Qmc_1 — SPRING-MOUND COLLUVIUM—Tan silty fine- to medium-grained sand. Dominantly travertine fragments carried by infrequent runoff from the mound. Minor salt crust present in some damp low-altitude areas. Surface generally is firm. Material saturated at shallow depths with ground water, having temperatures as high as 67°C

Qmc_2 — SPRING-MOUND COLLUVIUM—Mostly tan to brown sandy silt and silty sand derived characteristically by disintegration and erosion of spring mound travertine. Areas to the north and west have ground-water saturation generally to within 1 m of the land surface resulting in a surface salt crust or grassy wet areas. Locally, some areas have dry, firm surface

VALLEY—FLOOR DEPOSITS

Ql — LAKE BONNEVILLE SEDIMENTS—Mostly medium-brown sandy silt; relatively undissected. Land surface, usually near horizontal and flat, usually damp and hard; salt crust and marsh areas common near springs and seeps. Depth to ground-water saturation generally less than 1 m

Qld — DISSECTED LAKE BONNEVILLE SEDIMENTS—(See above description of Ql.) Dissected by infrequent runoff. Locally includes blow sand, sand dunes, and playa silt and clay. Depth to ground-water saturation variable and dependent on topographic position

Qsb — SAND BAR—Tan, medium- to coarse-sand and fine gravel. Mostly quartz, travertine, and volcanic-rock fragments. Bars extend about 1 m above the general land surface. Formed in Lake Bonneville at an altitude of about 1455 m above mean sea level

Qsd — SAND DUNES—Mostly fine- to coarse-grained, silty quartz sand transported mostly by southwesterly winds. Includes some dissected Lake Bonneville sediments (Qld)

CONSOLIDATED ROCKS

Qb — QUATERNARY BASALT—Dark gray, commonly vesicular

T_2bf — LATE TERTIARY BASALT AND BASALTIC ANDESITE

———— Contact

—— - —— Lineament on aerial photographs; may be a fault

············ Ridge line on spring mound

o— Spring or seep

FIGURE 25.—Continued.

Discovery of warm water at shallow depths elsewhere in the valley is unlikely.

Becks and Wasatch Hot Springs yield sodium chloride-type waters with fairly high concentrations of dissolved solids, 13,000–14,000 mg/L and 6,000–13,000 mg/L, respectively. Crystal Hot Springs yields water with lower dissolved-solids concentrations, in the range of 1,300–1,700 mg/L. Water flowing from faults to the alluvium near the springs probably has ion concentrations similar to the springs. As this water mixes with nonthermal water in the alluvium, the mix will have chemical characteristics intermediate between water types. In the northern area, the water in the alluvium will, therefore, have dissolved solids that are highly concentrated but less concentrated than the spring flow. Water from Becks, Wasatch, and Crystal Hot Springs or from wells nearby may be useful for space heating, but development of wells in the vicinity of the springs may stop the spring flow.

GREAT SALT LAKE DESERT

Thick beds of high-porosity clay underlie the Great Salt Lake Desert of northwestern Utah (fig. 1). Such beds have an insulating quality, impeding the conductive flow of heat to land surface. As a result, geothermal gradients must be high to discharge the heat flowing upward in the earth's crust. For example, if a regional heat flow of 2 HFU and a thermal conductivity of porous clay of 2×10^{-3} cal/cm/s°C are assumed to be reasonable values, the computation of the geothermal gradient would be:

$$I = \frac{HFU \times 10^2}{K}$$

where

I is the geothermal gradient in °C/km,
HFU is heat-flow units in μcal/cm²/s, and
K is thermal conductivity.

The calculation becomes

$$I = \frac{2 \times 10^2}{2} = 100°C/km.$$

If surface ambient temperatures on the desert are about 10°C, subsurface temperature at a depth of 1 km would be 110°C if the described bed of clay were also of that minimum thickness. The implication is that areas of thick clay accumulation, like the Great Salt Lake Desert, may have low-temperature geothermal potential for space heating without a near-surface source of heat or a permeable zone in which to circulate upward-flowing hot water from great depths. The temperature-gradient data on the Bonneville Salt Flats in the western part of the Great Salt Lake Desert, of Turk (1973) and Whelan and Petersen (1974), support a conclusion that a geothermal potential may exist in the area.

OTHER AREAS

The few geothermal areas described in this report were selected because they appear to have the highest development potential. However, other areas in western Utah may have important geothermal potential. Additional sources of geothermal data on these areas are the thermal-spring report by Mundorff (1970), computer-stored temperature data of the Water Resources Division of the U.S. Geological Survey, and other data in the files of the U.S. Geological Survey. (See table 18.) Table 18 lists temperatures and heat-flow data for nearly 100 wells and springs. It is not designed to be a comprehensive listing of such data for Utah, but rather a limited listing for widely spaced data points selected to provide information on wells and springs having mostly above-ambient temperatures.

SUMMARY AND CONCLUSIONS

1. Several publications have summarized data on hot and warm springs in Utah (Stearns, Stearns, and Waring, 1937; Waring, 1965; and Mundorff, 1970). As a result, no attempt was made to include in this report comprehensive tables of descriptions and data for ther-

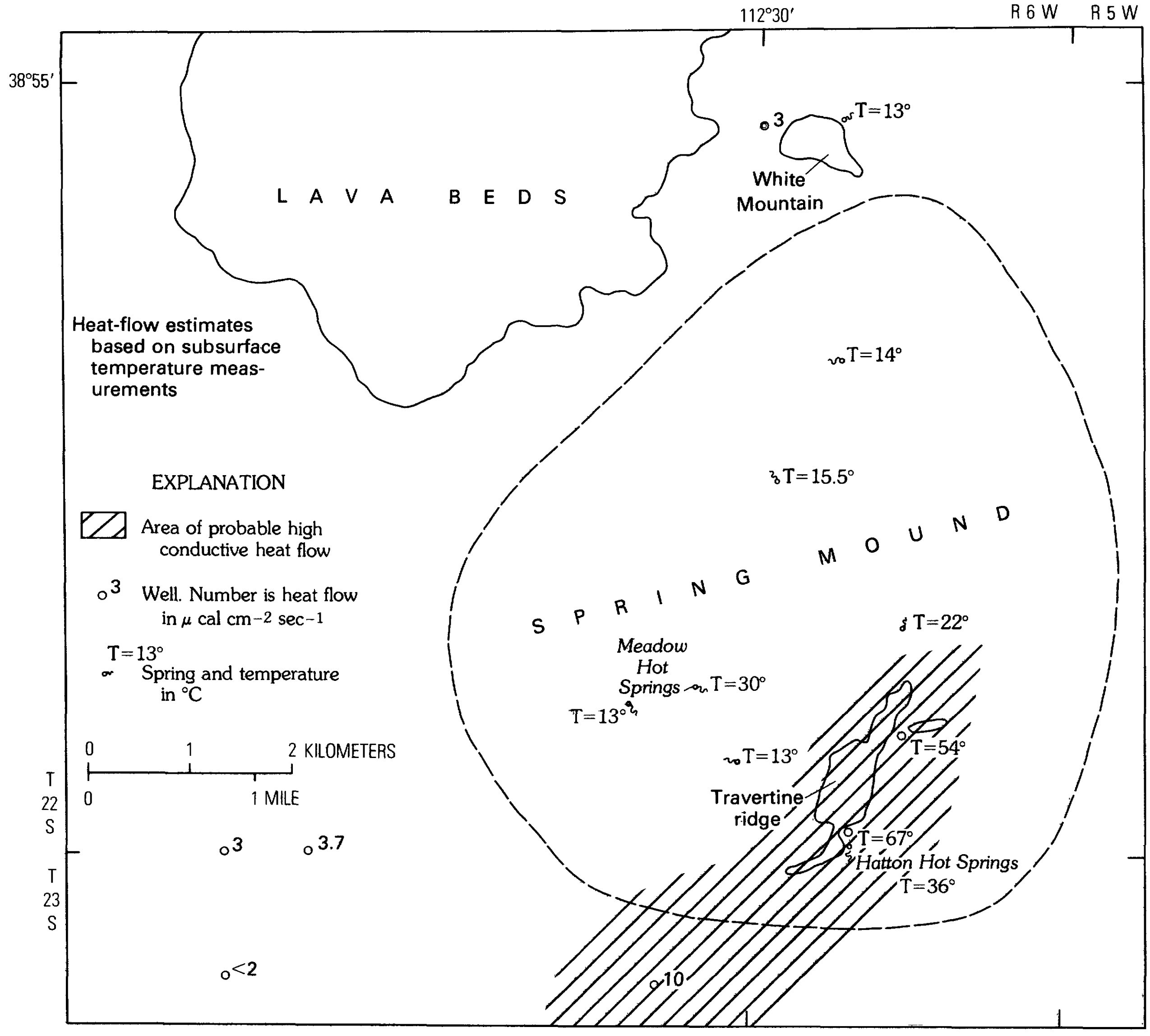


FIGURE 26.—Estimated heat flow and water temperatures in the Meadow and Hatton Hot Springs area.

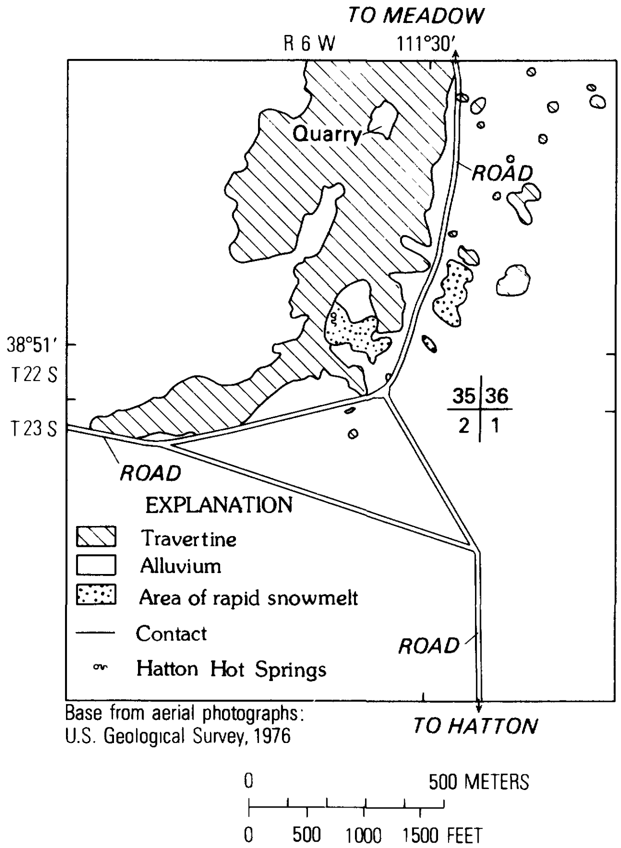


FIGURE 27.—Areas of rapid snowmelt near Hatton Hot Springs, March 1976.

mal springs. However, data for selected hydrothermal systems are presented and summarized in table 15.

2. Geologic factors are more favorable for geothermal resources in the Basin and Range province than on the Colorado Plateaus or in the Middle Rocky Mountains. The structure of the Basin and Range province is a result of sizable east-west crustal extension and crustal thinning during the last 17 million years. Igneous rocks less than 6 million years old crop out at many places, mostly in the southwestern part of Utah. Basalts as young as 10,000 years have been mapped.

3. The principal source of geothermal fluids is water stored in the hydrothermal reservoir and water entering the system as recharge from precipitation. Because of the semiarid climate of most of the area, most geothermal development for generation of electricity will remove fluids from storage at a higher rate than natural replenishment.

4. Conductive heat flows to the land surface at a generally high rate in the Basin and Range province of Utah are probably due to crustal thinning and possibly due to intrusion of young magmas into the earth's crust at shallow depths. The rates of conductive heat flow are commonly in the range of 1.5–2.5 HFU and probably average about 2.0 HFU. This is about normal for the Basin and Range province and considerably higher than the average of about 1.6 HFU for the entire earth.

5. High-temperature convection systems may be located by searching for high-silica volcanic and intrusive rocks that are of Quaternary age. Mapping of faults, hydrothermally altered rock, and thermal-spring deposits, along with the drilling of temperature-gradient holes, would be desirable components of an exploratory program.

6. Drilling for hot, high-silica, buried bodies of rock would best be pursued in the areas of recent volcanic activity.

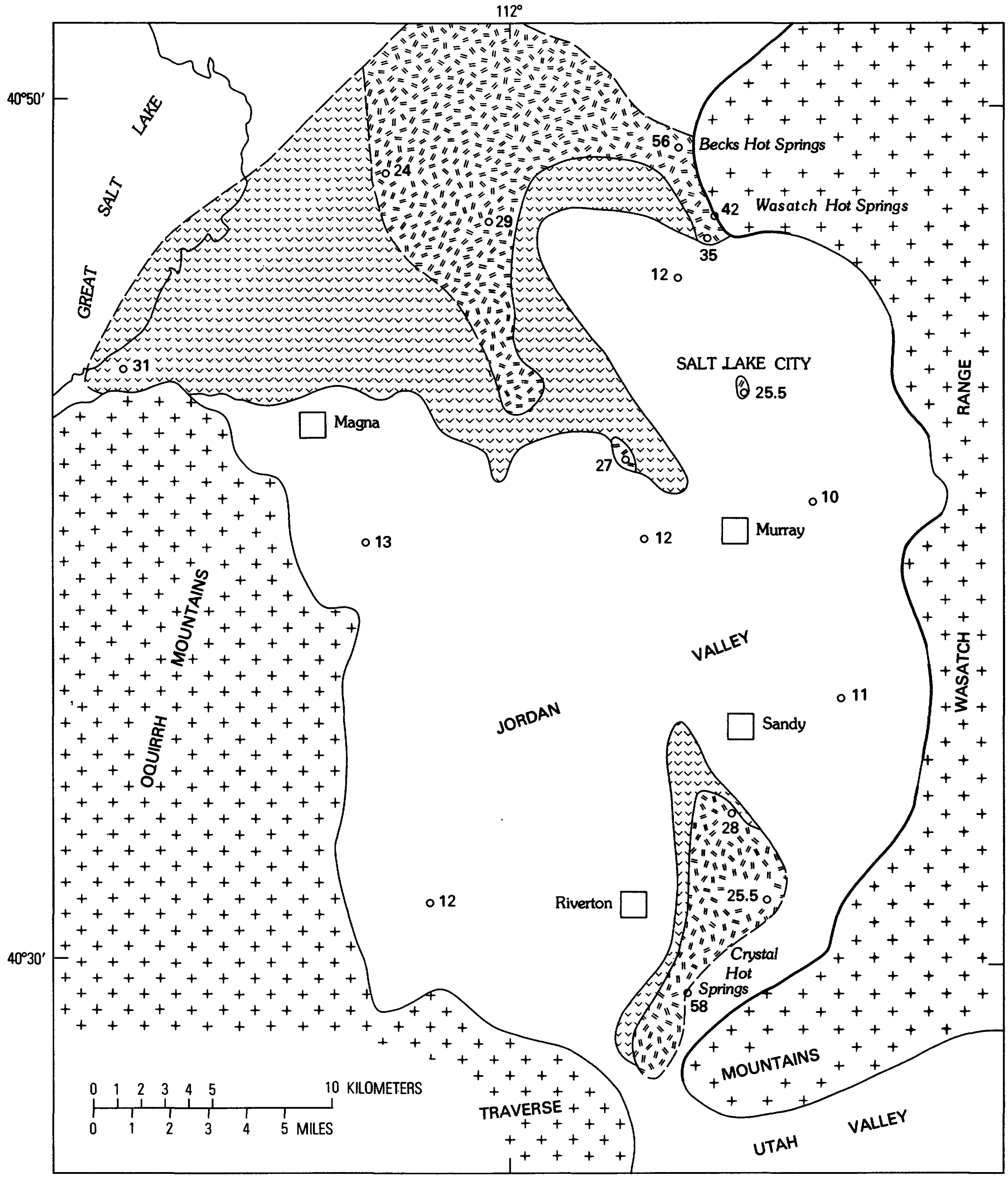
7. Some geothermal systems may be related to calderas because of their potential for eruption of large volumes of silicic rock. An example is the potential relation of Crater Hot Springs and Crater Bench to Thomas, Keg, and Desert calderas.

8. The southwestern part of Utah probably has the most promising geothermal potential, judged on the basis of spring temperatures, silica concentrations, and deposits such as siliceous sinter and sulfur.

TABLE 15.—*Summary of data for selected hydrothermal systems in Utah*

Hydrothermal system	Estimated water discharge from the hydrothermal reservoir ($\times 10^6 m^3/yr$): Evapo-transpiration	Spring flow	Ground-water outflow	Minimum total convective flow	Estimated total heat discharge ($\times 10^{13}$ cal/yr)	Temperatures (°C): Estimated reservoir	Spring flow maximum	Percent thermal water in mixed water	Estimated maximum depth of hydrothermal reservoir below land surface (km)
Roosevelt	Minor	Minor	Principal	Unknown	Unknown	260–290	85	----	6–7
Cove Fort-Sulphurdale	do	None	do	do	do	200±	----	----	5
Thermo	0.6	0.02	Probably small	0.6	15	140–200	82.5	40	3–4
New Castle	Minor	None	0.4	.4	7	140–170	[1]107.8	60	3–4
Monroe-Red Hill	.02	.2	Unknown	.2	3	100–160	75	50	2–4
Joseph	Minor	.02	do	.02	.3	100–170	65	35	2–4
Crater	2.5	([2])	do	2.5	36	110–140	87	50	1.3–1.7
Meadow-Hatton	4	Minor	do	4	43	70–120	[1]67	40	----

[1]Measured in well.
[2]Supports phreatophytes and evaporates. Included in the evapotranspiration estimate.



Modified from Marine and Price (1964)

FIGURE 28.—Areas of warm ground water in the Jordan Valley.

9. Deep exploratory drilling near Roosevelt Hot Springs has demonstrated that this KGRA has high potential for electric power generation. Reservoir temperatures are at least 260°C (table 15), and well testing demonstrates high reservoir permeability. The heat source may be related to Pleistocene rhyolites as young as 490,000 years.

10. The Cove Fort-Sulphurdale area may have reservoir temperatures as high as 200°C. No thermal water is known to discharge at land surface in the area, but sulfur deposits, altered ground, and gaseous emissions indicate past hydrothermal activity. Quaternary basalt flows are abundant in the area. The area extending northward 60 km to Neels, including Roosevelt Hot Springs and the Cove Fort-Sulphurdale KGRA's, probably has the best potential for geothermal development in Utah.

11. Thermo Hot Springs discharge from a hydrothermal system having an estimated reservoir temperature between 140°C and 200°C. Estimated hot-water circulation through the hydrothermal reservoir is at a rate of 18 L/s.

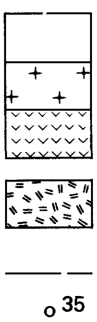
12. The Newcastle area has many thermal water wells but no thermal springs. The estimated reservoir temperature for the hydrothermal system is between 140°C and 170°C. An irrigation well has pumped boiling water at a rate of 108 L/s. Thermal water is discharged from a range-front fault from which it flows northward into an alluvial aquifer. The thermal water discharges its heat mostly by conduction to the land surface.

13. The Monroe-Joseph KGRA contains two hydrothermal systems, one at Joseph Hot Springs and the other at the Monroe-Red Hill Hot Springs complex. Reservoir temperatures appear to be at least 100°C but may be as high as 160°C–170°C.

14. Crater, Meadow, and Hatton Hot Springs (table 15) which discharge from hydrothermal reservoirs having estimated temperatures less than 150°C, could be considered for space and process heating. Despite their probable low reservoir temperatures, these prospects have the largest estimates of heat and water discharge of those systems listed in table 15.

15. Areas that may have hydrothermal potential but are inadequately defined are all in the Escalante Desert northwest of Beryl and Lund and east of Table Butte.

EXPLANATION



Alluvium

Consolidated rocks

Areas where ground-water temperature is commonly between 16°C and 21°C

Area where ground-water temperature is commonly above 21°C

Contact, dashed where location is indefinite

o 35 Well; number is temperature of water in °C

FIGURE 28.—Continued.

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TABLES 16, 17, AND 18

Table 16.—*Inventory of Thermo Hot Springs*[1]

[Flow rate: Estimated total visible flow of the orifices. Elevation: Reference to adjacent land surface (LS). Specific conductance is in μmhos/cm at 25°C. + indicates very small flows. See fig. 10 for map no.]

Map no.	Number of orifices	Flow: L/s	Flow: Direction	Principal orifice: Length (m)	Principal orifice: Width (m)	Principal orifice: Depth (m)	Principal orifice: Elevation (m)	Principal orifice: Temperature (°C)	Principal orifice: Specific conductance	Remarks
West spring mound										
Springs in south meadow area:										
1	1	+	Southwest	0.7	0.4	0.1	At LS	39	2,100	Southernmost spring on mound.
2	1	0.01	Southwest	.3	.3	1.0	At LS	70	----	Located 3 m north-northeast of No. 1 along mound axis.
3	4	+	Southwest	.3	.3	.3	At LS	66	----	20 m north of No. 2. Sinter deposits nearby.
4	4	+	Southwest and east	.3	.3	1.0	At LS	----	----	North end of meadow at shoulder of mound. Sinter deposits nearby.
Springs flowing east in mostly grassy channels:										
5	2	+	East	.4	.3	.6	At LS	56	----	15 m north of No. 4. Located slightly east of mound axis. Deposits of salt on surface.
6	2	+	East	.4	.4	----	At LS	54	----	Springs are oriented east-west. Only east spring supports grass; on east flank of mound.
7	4	+	East	.3	.3	1.7	+0.3	72	----	Mound is sinter, 1.5 m in diameter. Other three orifices are smaller and 3–6 m northeast.
8	5	+	East	.3	.2	.6	At LS	----	----	Water flows to and ponds on lowlands near mound.
Springs flowing in poorly developed channels:										
9	5	+	----	.3	.3	1.5	At LS	61	----	At southwest edge of north-south elongated grassy area, 11 m north of No. 8.
10	4	+	East	.4	.3	.6	At LS	70	----	Channel extends part way down flank.
Springs flowing east in grassy channels to lowlands:										
11	9	.1	East	.5	.3	.6	−.7	65	1,900	Spring area has three distinct arms at mound axis.
12	3	.1	East	.2	.2	.3	−.8	74	----	Principal spring is middle spring of group.
Spring flowing west:										
13	5	.3	West	.3	.3	----	−.8	67	----	On mound axis near south edge of trail crossing mound. Flow ponds on lowlands.
Springs north of trail crossing mound:										
14	1	+	East	.3	.2	.1	At LS	66	----	At east margin of mound.
15	1	3	West	.8	.7	.8	−.8	82.5	----	Located slightly west of mound axis in southern part of a 20-m diameter tule area.
16	2	2	East	.2	.2	----	−.5	54	1,650	At east margin of mound.
17	1	+	East	.2	.05	----	At LS	61	----	On east flank of mound. Farthest north of springs on mound.
Summary of springs on west mound: Number of orifices: 54; Total visible flow: 1–2 L/s						Maximum temperature: 82.5°C; Specific conductance range: 1,650–2,100 μmhos/cm				
East spring mound										
Springs in south area:										
1	1	+	East	.3	.2	.3	At LS	64	----	Half way up east flank of mound. Wet grassy area 10 m north, but no orifice.
2	2	+	East	.2	.2	----	−.4	57	----	Half way up east flank of mound.
3	1	----	----	.2	.2	.6	At LS	61	----	Slightly higher on mound flank than No. 2. Sinter deposits nearby.
4	1	.1	East	.3	.3	.6	At LS	67.5	----	At west edge of tules area.
5	1	+	East	.4	.3	.3	−.4	56	1,700	On mound axis. Flow is to large tules area.
Spring in tules area (30 m × 150 m):										
6	4	+	East	----	----	----	----	----	----	Half way up east flank of mound. Three grassy channels carry minor flow to lowlands.
Springs north of tules area:										
7	1	.2	East	2	1	>4	−.7	70	2,000	Two-thirds down east flank of mound and 20 m north of tules area.
8	1	+	East	1	1	.2	−.7	36	----	At southwest corner of man-made reservoir.
9	0	+	East	----	----	----	At LS	----	----	Seep; 150 m northeast of reservoir. Sinter deposits nearby on lowlands.
Spring at north end of mound:										
10	3	.1	Northeast	----	----	2	At LS	68	----	Three-fourths down north end of mound.
Summary of springs on east mound: Number of orifices: 15; Total visible flow: 0.5–1.0 L/s						Maximum temperature: 70°C				

[1]Inventory made in April and May 1976.

TABLE 17.—*Inventory of Crater Hot Springs, February 1976*

[Directions, distances, dimensions, and flow rates are estimated. See figure 28 for locations of larger springs. Size: Small, length and width less than 0.2 by 0.2 m. Flow: Small, less than about 0.05 L/S]

Pool or orifice	Location	Size (length × width × depth; in m)	Flow (L/s)	Maximum temperature (°C)	Remarks
Main pools:					
South (S)	Southwest flank of spring-mound crest	32 × 7 × 0.3	0	28	Many small orifices, mostly in southeast corner of pool; undrained.
North (N)	Northwest flank of spring-mound crest	30 × 4 × 1	<.1	58	Drains in ditch to northeast.
Along southwest ditch:					
D1	10 m southwest of main pool-south	0.3 × 0.3 × 0.2	1–2	63	
D2	5 m southwest of D1	Small	Small	74	
D3	15 m southwest of D2	0.2 × 0.2 × 0.1	Small	--	
D4	2 m southwest of D3	----	?	80	Group of three orifices, 3–4 m west of ditch.
D5	9 m southwest of D4	1.5 × 1 × 1	----	77	Group of three orifices, 1 m west of ditch.
D6	3 m southwest of D5 near bush 2 m high	----	----	--	1 m southeast of ditch.
D7	1 m southwest of D6	----	----	--	
D8	9 m southwest of D7	----	Small	69	11 m northwest of swamp.
D9	9 m southwest of D8	Small	?	--	5 orifices in ditch and many to southeast.
D10	Southwest of fence	0.5 × 0.5 × 0.5	----	84	Two orifices; temperature is of northeast orifice.
East of main pools:					
E1	20 m east of main pool-south	6 × 5 × 0.5	0	38	No visible orifices.
E2	12 m northeast of E1	7 × 4 × 0.2	0.3–0.6	82	Ditch extending east southeast. At end of ditch, flow = 5 L/s and temperature = 65°C.
E3	100 m east of E2	1 × 0.5 × 0.5	.3	64	Similar orifice 2 m northeast.
E4	140 m east southeast of main pool-north	1.5 × 17 × 0.5	.3	87	Trenched east; flow dissipates.
E5	90 m east of E4	1.5 × 1.5 × 0.2	0	--	
Recycling pools east of main pools:					
R1	20 m southwest of E5	1 × 1 × 0.6	Small	71	Swamp on south; brush on north.
R2	25 m east of R1	5 × 0.7 × 1	3	70	Source orifice in west end of pool.
R3	15 m southeast of R2	----	5	29	Source is surface flow from swamp. Several other recycling orifices to east.
Northeast of main pools:					
N1	30 m east of main pool-north	5 × 5 × 0.5	0	25	
N2	70 m northeast of main pool-north	3 × 3 × 1	.1	59	
N3	30± m west of N2	1 × 1 × 0.2	Small	77	
N4	15± m west of N3	----	.3	--	Several orifices; along ditch draining northeast from main pool-north.
Main drain; spring:					
----------	500 m east of main pool-north	3 × 4 × 1	72	52	Drains in ditch to northeast and then north toward set of buildings. Specific conductance was 5,800 μmhos.

Summary:

Observed flow:	[1]90± L/s	Observed orifices and pools:	40± 5
Flow from seeps and the like (approximate):	[2]45± L/s	Maximum observed temperature:	87°C
Total (rounded):	140± L/s		

[1]May include some recycled water.
[2]Estimated as half the observed flow.

TABLE 18.—*Selected subsurface temperature and heat-flow data not summarized on maps*

[Temperature description: M, maximum in well; probably bottom-hole temperature in most wells; D, discharge temperature of well or spring. Temperature gradient: Estimated, computed from listed temperature, land-surface ambient temperature, and well depth. Estimated heat flow: HFU, heat-flow units, in $\mu cal/cm^2/s^1$ based on average temperature gradient and generalized thermal conductivity by lithology]

Approximate well or spring location	Total depth (m)	Temperature		Average Temperature gradient °C/km	Estimated conductive heat flow (HFU)	Remarks
		Degrees °C	Description			
Beaver area						
(C-28-7)15bb	300	10	D	Very small	<1	Irrigation well.
(C-28-7)31ad	34	11.2	D	Very small	<1	Unused.
(C-29-7)15cd	46	11.6	D	Very small	<1	Public-supply well.
(C-29-8)9ba	46	18	D	140	3.6	Stock well.
(C-29-8)31ad	94	12.1	D	Very small	<1	Irrigation well.
(C-29-8)35ab	157	18.5	D	45	1.3	Do.
(C-29-8)36ac	110	22	D	96	3	Do.
(C-30-7)5cd	245	22	D	43	1.3	Do.
East of Cedar City						
(C-36-7)33	2,016	----	----	----	----	Mud temperatures were reported as high as 43°C.
Eastern Utah						
(D-5-22)22ac	1,311	46	D	27	1.6	
(D-11-24)8ca	2,002	25	D	7	<1	
(D-12-21)19bdS	----	19.5	D	----	----	Sulphur (SIC) Spring.
(D-22-6)4ca	492	26.5	D	33	1.3	
Milford-Minersville area						
(C-27-10)6da	30	13.5	D	49	1.5	Stock well.
(C-27-10)31dc	213	27	D	70	2	Irrigation well.
(C-28-10)18ac	138	21	D	65	2	Do.
(C-28-10)14bb	78	20.5	D	110	3.3	Stock well.
(C-28-10)31dd	59	13.5	D	25	<1	Irrigation well.
(C-28-11)10ac	69	16.5	D	65	1.5	Stock well.
(C-28-11)23cb	29	14	D	69	2.1	Irrigation well.
(C-29-10)27bb	45	12.5	D	Very small	<1	Stock well.
(C-29-11)27ad	36	14.5	D	69	2.1	Irrigation well.
(C-30-9)7ad	22	33.5	D	980	29	Near warm springs, east of Minersville. Reflects convective flow.
(C-30-10)19ab	89	21	D	100	3	Irrigation well.
Northwestern Utah						
(A-12-1)16dd	74	22	D	150	4	
(B-1-1)31d	183	28.5	M	72	<2	Near Salt Lake City airport.
(B-1-9)24cd	79	24	D	160	4	
(B-1-18)29cc	50	28	D	350	7	
(B-1-18)31cb	70	24	D	200	4	
(B-4-3)19ca	146	24	D	96	3	
(B-5-1)30ad	274	55	D	160	5	
(B-5-13)31ac	61	22	D	180	3.6	Temperature data from Stephens (1974, p. 44).
(B-6-3)19aa	67	19	D	130	4	
(B-6-5)21aaS	----	21	D	----	----	
(B-7-5)15cbS	----	25	D	----	----	
(B-7-5)22cdS	----	22	D	----	----	
(B-8-5)5cdS	----	22	D	----	----	
(B-10-6)9bbS	----	22	D	----	----	
(B-10-15)6cdS	----	20	D	----	----	Warm Spring No. 2.
(B-11-11)6dbS	----	19	D	----	----	Black Butte Spring.
(B-12-5)22daS	----	20	D	----	----	
(B-12-6)33dbS	----	20.5	D	----	----	
(B-13-12)30caS	----	25	D	----	----	
(B-13-13)27ddS	----	21	D	----	----	
(B-13-13)34cbS	----	21	D	----	----	
(B-13-13)35bbS	----	23	D	----	----	
(B-13-14)21ddS	----	19.5	D	----	----	
(B-13-14)24dcS	----	23	D	----	----	
(B-13-16)23ccS	----	21	D	----	----	Head Spring.

TABLE 18.—*Selected subsurface temperature and heat-flow data not summarized on maps*—Continued

Approximate well or spring location	Total depth (m)	Temperature Degrees °C	Temperature Description	Average Temperature gradient °C/km	Estimated conductive heat flow (HFU)	Remarks
Northwestern Utah—Continued						
(B–14–9)4bb	107	22	D	110	2.8	Stock well.
(B–14–9)9bb	110	21	D	100	2.5	Irrigation well.
(B–14–10)33bcS	----	43	D	----	----	Coyote Spring.
(B–15–9)28cb	122	24	D	110	2.9	Irrigation well.
(B–15–9)30ab	124	21.5	D	90	2.8	
(C–1–8)6ab	20	26.5	D	790	20	
(C–1–17)34ba	1,298	67	M	44	<2	
(C–1–19)1bb	50	24	D	260	5	
(C–1–19)3dc	53	24	D	240	5	Temperature data from Stephens (1974, p. 45).
(C–1–19)9db	27	24	D	480	10	Do.
(C–1–19)10ba	33	31	D	610	12	
(C–1–19)34cd	351	32	D	60	1.2	Do.
(C–2–6)23cb	64	20	D	140	4	
(C–2–19)24c	499	88	M	160	4	Temperature date from Turk (1973, p. 9).
(C–4–19)7S	----	29	D	----	----	Blue Lake Spring.
(C–5–1)23bd	32	21	D	310	9	
(C–5–1)24dc	27	22	D	400	12	
(C–5–1)25ab	30	24	D	430	13	
(C–5–1)25ba	30	23	D	390	12	
(C–5–1)25cb	45	35	D	540	16	
(C–5–1)25cc	32	46	D	1,100	33	Near Saratoga Springs.
(C–5–1)26bd	152	50	D	120	3.7	
(C–6–1)18dc	85	27	D	180	5	
(C–10–2)15 and 22	----	----	M	104	----	Temperature gradient is average for 31 values (Lovering and Goode, 1963, table 11).
(C–10–2)15 and 22	----	----	M	140	----	Do.
(C–10–2)15dd	----	54	----	----	----	Mine effluent.
(D–2–5)32bb	53	21	D	190	6	
(D–7–3)28bd	103	32	D	200	6	
(D–8–2)28cc	84	33	D	260	8	
(D–8–2)23dc	174	59	D	280	8	
Southwestern Utah						
(C–12–5)31	114	29.5	D	135	4	Heylmun (1966, p. 21).
(C–18–4)31db	159	51.5	D	250	8	
(C–41–15)32ac	183	43.5	D	130	3.8	
Near Thermo Hot Springs						
(C–29–11)4ba	17	13	M	Very small	<2	Martin well.
(C–29–11)17aa	18	12.5	M	Very small	<1	Windmill.
(C–39–11)19db	21	15.5	M	170	4	Do.
(C–30–13)34bb	20	13	M	Very small	<1	Do.
(C–31–12)30cd	----	17.5	D	----	----	May indicate significant heat flow.
(C–31–13)18aa	28	15	D	100	<2	
(C–31–13)23bb	----	12.5	D	----	<1	Windmill.
(C–32–12)6cb	21	9	M	Very small	<1	Do.