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By

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Abstract

Hot springs are numerous and nearly uniformly distributed in northern Nevada. Most occur on the flanks of basins, along Basin and Range (late Miocene to Holocene) faults, while some occur in the inner parts of the basins. Surface temperatures of the springs range from slightly above ambient to boiling; some springs are superheated. Maximum subsurface water temperatures calculated on the basis of quartz solubility range as high as 252°C, although most are below 190°C. Flows range from a trickle to several hundred liters per minute.

The Nevada geothermal systems differ markedly from the power-producing system at The Geysers, Calif., and from those areas with a high potential for power production (e.g., Yellowstone Park, Wyo.; Jemez Mountains, N. Mex.). These other systems are associated with Quaternary felsic volcanic rocks and probably derive their heat from cooling magma rather high in the crust. In northern Nevada, however, felsic volcanic rocks are virtually all older than 10 million years, and analogous magmatic heat sources are, therefore, probably lacking.

Nevada is part of an area of much higher average heat flow than the rest of the United States. In north-central Nevada, geothermal gradients are as great as 64°C per kilometer in bedrock and even higher in basin fill. The high gradients probably result from a combination of thin crust and high temperature upper mantle.

We suggest that the geothermal systems of northern Nevada result from circulation of meteoric waters along Basin and Range faults and that their temperature chiefly depends upon (1) depth of circulation and (2) the geothermal gradient near the faults.

Introduction

The expanding interest in geothermal power as a supplement to conventional sources has resulted in an intensified study of known hydrothermal systems as well as increased effort to find concealed but exploitable geothermal systems. The abundance of hot springs in northern Nevada makes it a promising area and has served as a stimulus for this study. The work upon which this paper is based was begun in 1972 on a part-time basis, and was part of the U.S. Geological Survey's general program of geothermal research. Our major efforts were devoted to gathering and compiling a variety of geological, geophysical, and geochemical data, although the principal effort was an evaluation of the geological environment of areas considered to be favorable for geothermal energy. Of particular help in our work was a compilation of thermal spring data by Waring (1965). We found that some of the springs in this compilation were much cooler than reported and in some few cases nonexistent. In the course of this study we examined most of the hot springs of northern Nevada that we felt possessed the potential for geothermal power; that is, those that were reported to be the hottest or occurred in areas considered favorable geologic environments. On the basis of these examinations, chemical analyses would be performed in the field on selected high-potential springs while samples of the less promising springs were to be analyzed in the U.S. Geological Survey laboratories at Menlo Park, Calif. For the rationale of field analysis and for a summary of all these analyses, the reader is referred to Mariner and others (1974). Geologic mapping was done at the Beowawe geyser area.

This paper will briefly summarize the geologic setting of some of the hot-spring systems, relate them where possible to various geophysical and geochemical parameters, and propose a model to explain their origin. This model differs in only minor respect from those where heat is derived by conduction and convection from a magma.

Geologic Setting of Northern Nevada

Northern Nevada is in the north-central part of the Basin and Range province and is characterized by elongate northerly trending mountain ranges flanked by more or less flat-bottomed basins. The region contains thick sequences of rock ranging in age from Precambrian to Holocene. The Paleozoic and Mesozoic rocks are extraordinarily varied, principally because the region included both miogeosynclinal and eugeosynclinal depositional habitats. They include chert, graywacke, shale, sandstone, pillow lava, limestone, dolomite, evaporites, and others, as well as their metamorphosed equivalents. During the Antler orogeny of Late Devonian to Mississippian age, eugeosynclinal rocks were moved eastward over their miogeosynclinal correlatives on thrust faults having an aggregate displacement of up to 90 miles. During Pennsylvanian and Permian time the disturbed terrane was overlapped by marine sediment; to the east, in the miogeosyncline, deposition was continuous and the effects of the Antler orogeny are barely noticeable. The western part of the State was also subjected to tectonism during the Permian Sonoma orogeny and again during the Middle to Late Triassic(?). These events were interspersed with periods of marine sedimentation. The entire region was then affected by massive tectonism that occurred probably during the late Mesozoic to early Tertiary. The resulting terranes were intruded by quartz monzonite to granite magmas during the Triassic, Jurassic, and Cretaceous. Gabbroic rocks of Jurassic age are present in the southern part of the West Humboldt Range and in the northern part of the Carson Sink.

The Tertiary geologic history began with local accumulation of nonmarine strata to be followed in late Eocene time by the beginning of a period of volcanism that persisted into the Holocene.[1] The main volcanism took place from the Oligocene to middle Pliocene time and was so extensive that volcanic rocks now comprise nearly 25 percent of the outcrops of northern Nevada.

The Tertiary volcanic rocks are mainly silicic to intermediate in composition but include basaltic intercalations especially in the northern part of the province near the Snake River Plain. These rocks include lava flows, ash flows, some air-fall tuffs and volcanogenic sedimentary units. Although the vents or eruptive centers for some of the volcanic rocks have been recognized, the source for many is unknown. Some known source areas from which large volumes of rock were erupted are: the southern Fish Creek Mountains described as a volcanic center and collapse caldera (McKee, 1970); the Clan Alpine Mountains, a complex eruptive center (Noble and others, 1972); the Northumberland caldera (McKee, 1974) near Northumberland Canyon on the west side of the Toquima Range; a volcano-tectonic depression (caldera structure filled with eruptive rocks) in the northern part of the Toiyabe and Shoshone Ranges (Masursky, 1960); a caldera and associated volcanic rocks west of McDermitt, Nev. (Yates, 1942; Walker and Repenning, 1965, 1966); and a series of linear vents from which a large volume of ash-flow tuff was erupted in Washoe and Humboldt Counties, Nev. (Korringa and Noble, 1970).

The Pliocene to Holocene volcanic rocks, found mostly near the northern and western margin of the region are principally basaltic although there are bodies of rhyolite and andesite as well. Late Pliocene to Holocene basaltic rock distribution is shown on figure 1.

The present elongate northerly trending mountain ranges of high relief, flanked by more or less flat-bottomed valleys or basins, are the products of tectonism that probably began as long ago as the Oligocene in some places, but in most regions started in middle Miocene (15-17 m.y. ago), and continues today. This widespread rifting caused extension of the entire province and produced the steep fault scarps that occur on either or both sides of the fault-block mountain ranges. The total extension across the Basin and Range province has been estimated by Thompson and Burke (1973) on the basis of geologic and geophysical work in Dixie Valley, Nev., to be about 100 km. Since the ranges and basin are relatively uniformly distributed across the province, Stewart (1971) has suggested deep-seated extension of a plastic substrate.

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[1] We use the following time division of the Tertiary which is based on mammalian chronology (Evernden and others, 1964): Quaternary, 0-2 m.y.; Pliocene, 2-12 m.y.; Miocene, 12-26 m.y.; Oligocene, 26 to 37-38 m.y.; Paleocene and Eocene, 38-65 m.y.

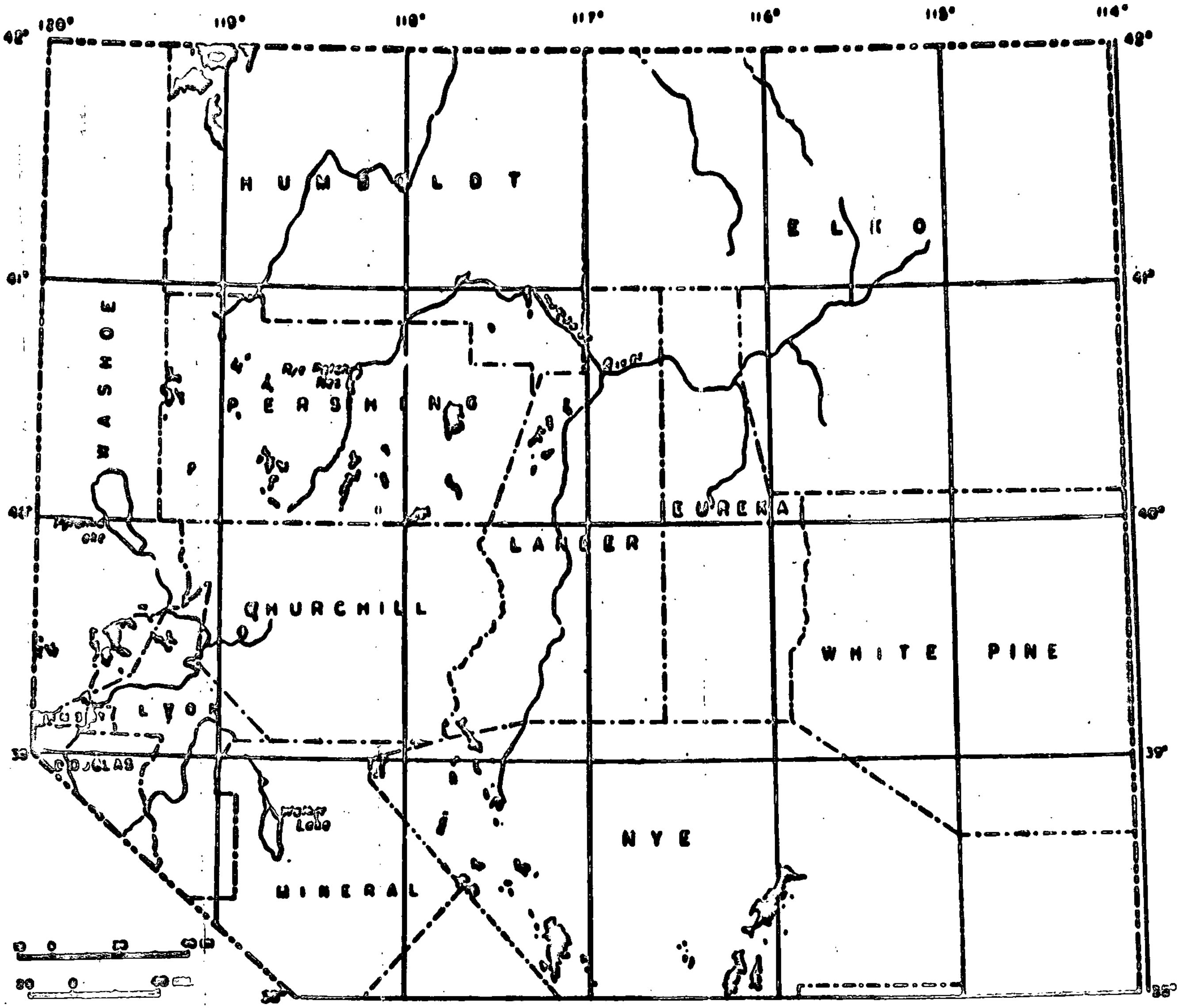

Fig. 1 Distribution of late Pliocene to Holocene basalts.

The flanking faults are the latest structures imposed on what was locally at least a structurally complex terrane of diverse lithology. The faults produced gouge and breccia zones with permeability to meteoric water dependent principally upon the original lithologies, but subsequent movement of chemically active waters presumably altered this permeability, increasing it in some places and decreasing it in others.

Hot Springs

Distribution and Setting

Hot springs in northern Nevada are numerous and, except for a few rather large areas that lack them, are fairly uniformly distributed (fig. 2). In the eastern part of northern Nevada, in southeastern Elko County and northeastern White Pine County (fig. 3), there is a large area that lacks hot springs. Two other areas lack springs, one in the northwestern corner of Nevada and the other a 30- to 50-mile-wide band extending from east of Pyramid Lake northeastward to Oregon and widening eastward into western Elko County.

The latter two areas are separated by a 30-mile-wide zone that contains several scattered hot springs, and this zone extends from just west of Pyramid Lake N. 30°-35° E. to the Oregon State Line and probably continues northward in Pueblo Valley, Oreg., to the Alvord Desert, another 50 miles.

In various areas in northern Nevada, the hot springs seem to occur in slightly different physiographic and geologic environments. Most are located close to the margins of the basins, but many springs occur more basinward. No hot springs are known to occur in the mountains. Most issue from Quaternary cover although some rise from bedrock, but the positions of all are believed controlled by Basin and Range faults.

In some places several springs are present and in others only one or two. Where several springs occur as part of one system, they are either disposed as clusters more or less equant in outline, or in a linear array up to several miles long. Many springs occur singly and have no neighbors.

Beowawe Geyser Area

Among the systems that can be characterized as linear is the Beowawe geyser area in Whirlwind Valley. Figure 4 is a reconnaissance geologic map that shows the distribution of hot springs, siliceous sinter, and altered and unaltered volcanic rock. The spring deposits, made up of opaline silica (siliceous sinter), cover nearly 0.6 of a square mile and are about 6,000 feet long and 2,800 feet wide. The sinter covers the valley floor and rises to a 300-foot-high terrace that rests against the northern edge of a ridge of volcanic rock. Sinter was deposited on the altered and unaltered volcanic rock and an inferred fault that provided the channelways or conduits for the springs.

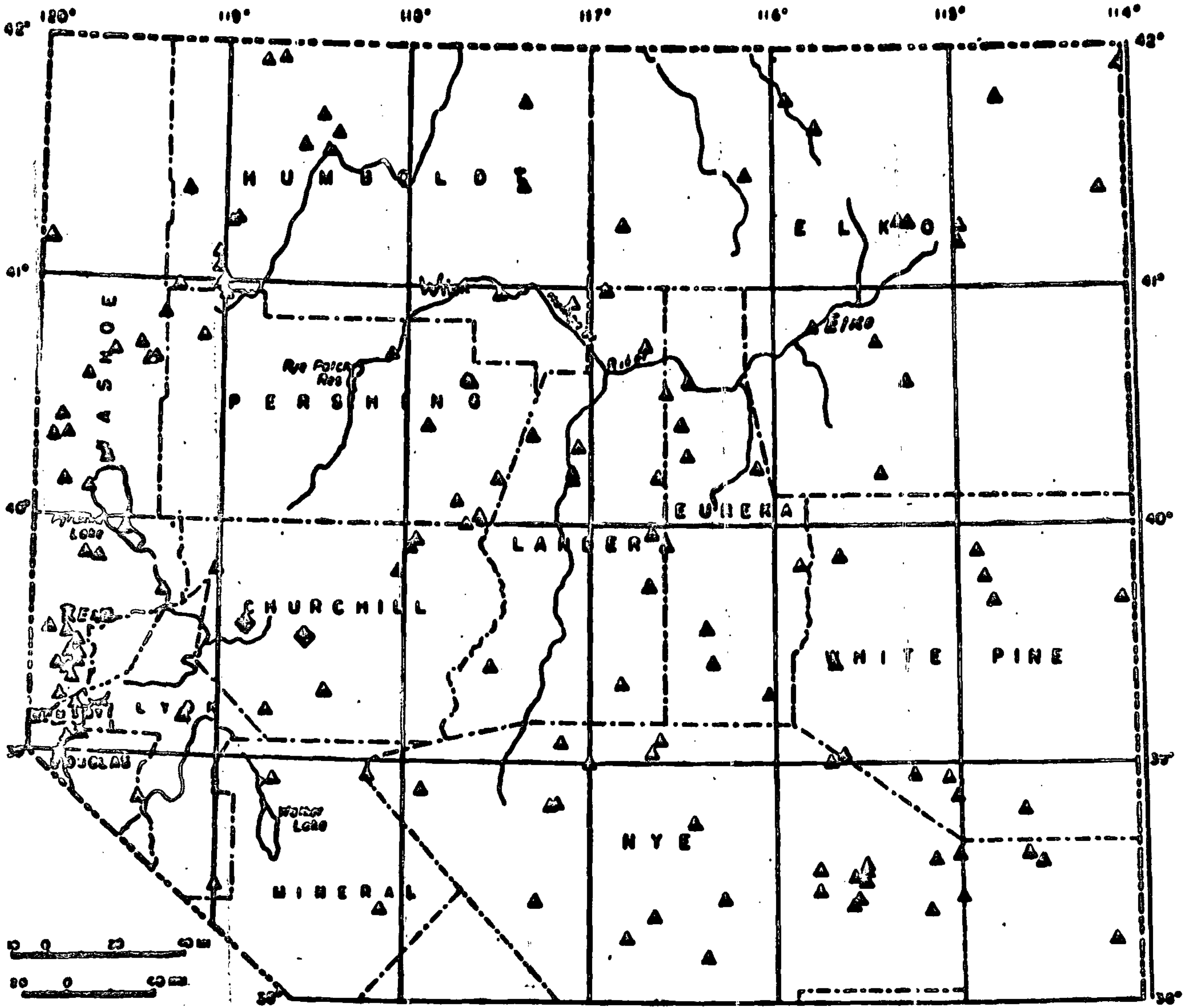

Fig. 2 Distribution of hot springs (▲) in northern Nevada,
See Table 1

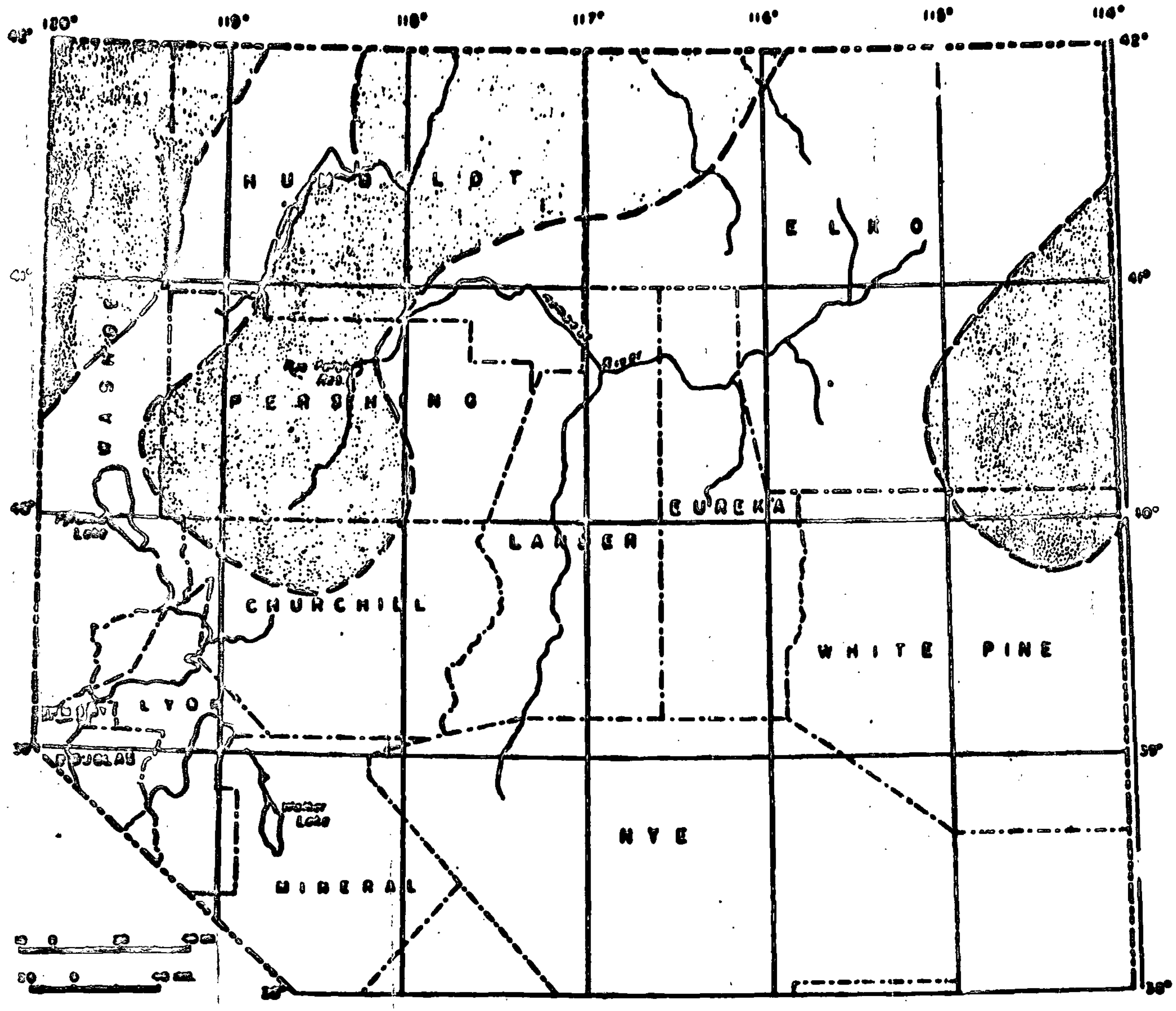

Fig. 3 Stippled areas lack hot springs

	Total depth	°C max. temp. during drilling	°C max. temp. on test
Beowawe No. 1	1918	—	—
Magma No. 1	715	212	168
Magma No. 2	683	203	171
Magma No. 3	715	203	179
Magma No. 4	767	210	182
Vulcan No. 6	—	—	—
Sierra Pacific Power No. 1	920	—	81
Sierra Pacific Power No. 2	430	—	56
Sierra Pacific Power No. 3	1180		193

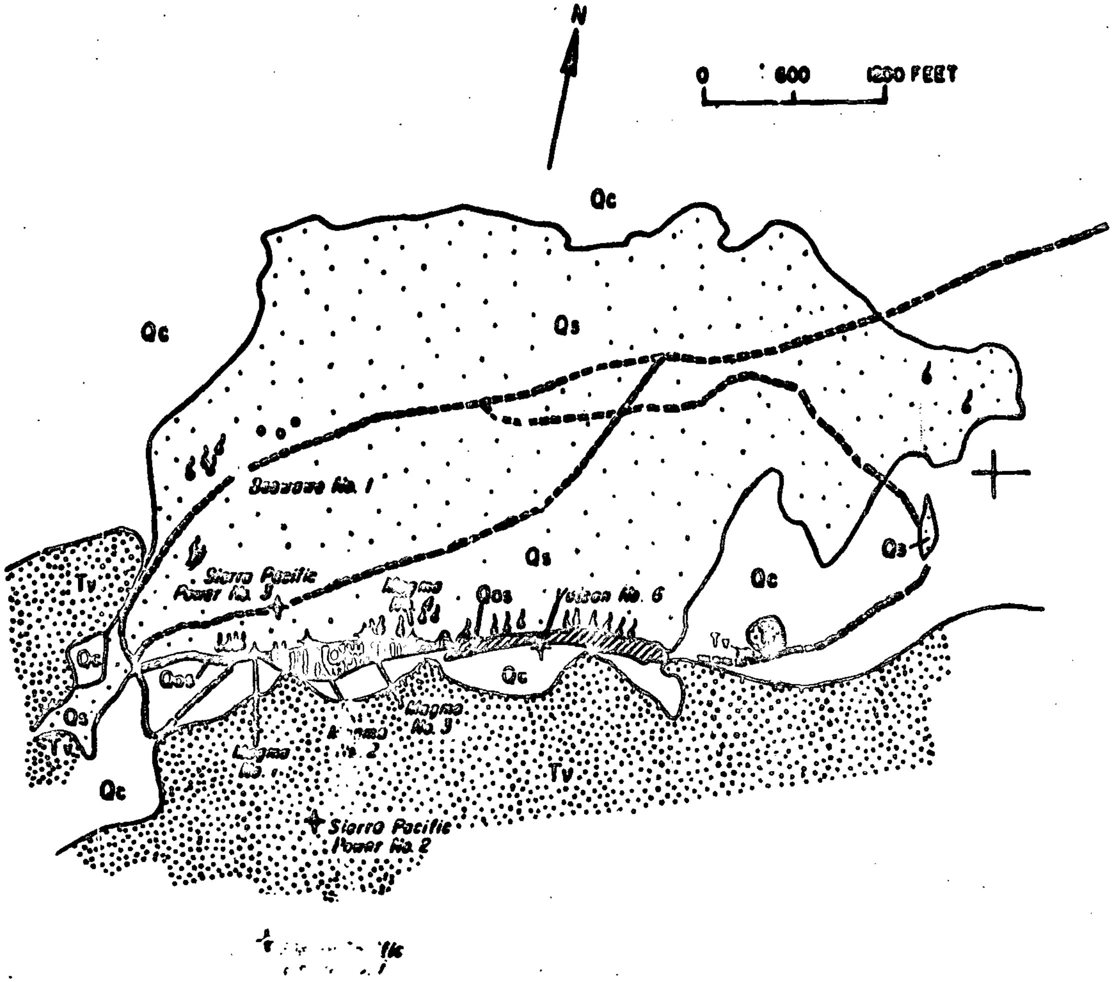

Fig. 4 Geologic map of Beowawe Geyser area showing location of boreholes, maximum depth, maximum and sustained temperatures in degrees C.

Most of the springs occur at the top of the terrace along a line about 0.5 mile long which trends about N. 75° E., although several occur to the north and northeast of the terrace.

When the Beowawe thermal area was first described by Nolan and Anderson in 1934, the natural system was much more active than it is now. Specifically, it included three geysers, two of which played to heights of 3 feet and one to 12 feet. In 1951, White (personal commun.) observed about the same level of hydrothermal activity as Nolan and Anderson.

Beginning in 1960, nine test holes were drilled by Sierra Pacific Power Company and by Magma-Vulcan. After extensive testing, all the wells were capped. The effects of the drilling and testing on the natural systems are unknown. Vandals blew the caps from four of the wells (Magma 1, 2, 3, and Vulcan 6) sometime prior to 1972, and one of these released steam and water in rather large volumes. One of the noticeable effects of this release of fluid and possibly the earlier drilling was the cessation of geyser activity. In August 1972, Magma 2 and 3, which had been earlier vandalized, began to vent and released large volumes of hot water and steam.[2] This resulted in the diminution of spring flow and cessation of some others. Magma 1 was capped during the winter of 1972-73.

Drilling data for the Beowawe area are summarized in figure 4 which shows location of the wells, total depth, maximum temperature during drilling, and maximum temperature on test. Maximum bottomhole temperature attained during drilling ranged from 208°C to 210°C, while subsequent tests of temperature after the wells were allowed to flow ranged from 165°C to 182°C for wells Magma 1 through 4. The drop in temperature results from a lowering of the boiling point as the pressure in the system is lowered by venting. A comparison of flow pressure and temperature at the bottom (767 feet) in 1961 with 1965 for Magma 4 shows a reduction from 119 psig to 46 psig and temperature at 767 feet reduced from 210°C in 1961 to 170°C in 1965.

Double Hot Springs

A linear fault-controlled hot spring array is present on the east side of the northwestern re-entrant of the Black Rock Desert, just west of the Black Rock Range. The fault which is Holocene, and which has negligible displacement, is identified by the northerly alinement of linear seeps that cut fans, small hills, and mounds at a high angle (fig. 5). The northern end of the system is marked by Double Hot Springs, which have a flow of about 80 gpm of 81°C water. Seven springs and seeps are present to the south

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[2] The reason for this spontaneous eruption is unknown, but it is possible that as a result of a very dry winter (1971 to 1972), the ground water had receded to the point where steam pressure was able to flow out the water remaining in the bore holes.

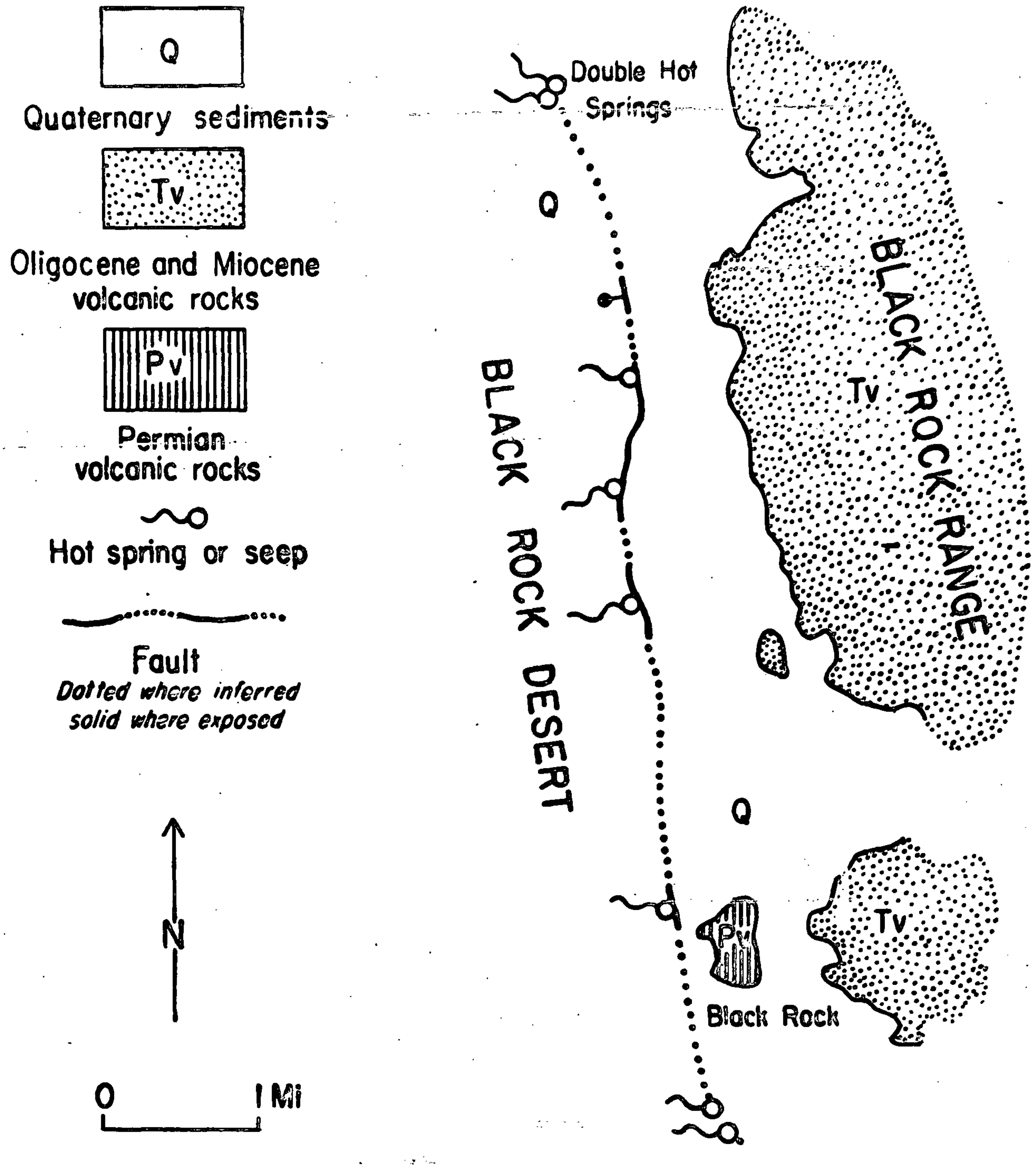

Fig. 5. Geologic map of the Double Hot Springs–Black Rock area.

for a distance of 7 miles. The southernmost end of the system is marked by an unnamed spring with a low flow of water at 90°C. The flow from the springs between the two ends is low. Extrapolating from what is known elsewhere of the style of Basin and Range faulting, a pre-Holocene fault could be inferred beneath the Holocene one. This is supported by the position of the Permian volcanic rock making up Black Rock itself, which is only a short distance east of the trace of the fault.

The Holocene fault then would have developed in response to a reactivation of the older fault if the latter is present as inferred.

Slemmons and others (1965) list an earthquake of magnitude 4.1 at lat 41°N., long 119°W. for the year 1936. Its proximity to this Holocene fault suggests a relationship although certainly not the cause of the springs since the Double Hot Springs were observed long before 1936 (Hague and Emmons, 1877).

Data in Mariner and others (1974) show that the water from Double Hot Springs at the north end of the fault is considerably more dilute than the spring waters south of Black Rock, containing 0.89 as much SiO_2, 0.14 as much Ca^{++}, 0.0025 as much MgO, 0.12 as much Na^+, 0.225 as much K^+, and comparable fractions for the anions. The larger amounts of dissolved solids in the southernmost springs are attributed to contamination by subsurface brines, the southernmost spring being closer to the topographically lowest point in the Black Rock Desert which would be the most favorable site for accumulation of saline water at various depths.

Calculations of subsurface temperatures, based on SiO_2 content, of the springs at either end of the fault despite the implied contamination, show reasonably consistent temperatures, 135°C-140°C for Double Hot Springs and 141°C-147°C for the unnamed spring at the south end.

Another prominent linear array of springs is Brady Hot Springs, a fault-controlled system in bedrock (altered and unaltered volcanic rocks) and associated with siliceous sinter.

Buffalo Valley has a subcircular group of many very low flow hot springs that typify the cluster array mode of occurrence. The cluster, which is located in the southeastern part of Buffalo Valley, is about 2 miles northwest of a line of 3-m.y.-old basalt cones (McKee, personal commun., 1974). It is about a quarter of a mile in diameter and emerges from a circular mound that is a few feet above the surrounding flat and made up of marly material. A few of the hottest springs deposit travertine, but others are so cool or have such a low flow that no discernible deposit accumulates. The springs emerge from what would be the distal edge of a large fan if its basinward half mile or so weren't reworked and covered by lake beds. Although the older terrane is obscured by a veneer of Quaternary sediment, it is inferred that the position of the cluster is controlled by pre-Quaternary faults.

Many other hydrothermal systems fall into the cluster category such as the springs near Soldier Meadow, Spencer Hot Springs, springs northeast of the Ruby Marshes, Hot Sulfur Spring in the northern Ruby Valley, Hot Pot, etc. The hot springs in Smith Creek Valley are similar in mode of occurrence, but are almost of a type intermediate between the linear and cluster arrays, and seem to emerge close to the basinward edge of a fan. Dixie Hot Springs, on the northwest margin of Dixie Valley, also emerge from the distal edge of fans but mainly where two fans coalesce.

A small-scale (AMS 1:250,000) map of western Humboldt County, Nev., discloses a prominent lineament beginning in the vicinity of a large cluster of hot springs close to Soldier Meadow, and trending N. 30° to 35° E. through Baltazor Hot Spring into Pueblo Valley, Oreg. and Nev., a distance of more than 65 miles (fig. 6). Attempts to understand the relationship of this lineament to the hot springs are based on unpublished reconnaissance geologic mapping by D. C. Noble (1969).

Noble's geologic map shows the lineament to be considerably more complex than the topographic map indicates. Just north of Summit Lake the lineament splits and both branches appear to be controlled partly, although not wholly, by faults. Perhaps less prominent structures, such as joints in the Tertiary lavas, contribute to the lineament. In any case, the faults are of only modest displacement. The eastern branch at the northern end is a concealed fault of very large displacement that truncated the Pueblo Mountains at an angle of about 50° from their trend. Baltazor Hot Spring rises from this part of the lineament. The western branch of the lineament extends along faults as far north as McGee Mountains and is inferred to extend farther north beneath what Noble mapped as Late Miocene tuffs to terminate at Bog Hot Springs.

The tremendous contrast in magnitude of faulting along the lineament suggests, to us at least, that the lineament existed as a large fault in the Early Tertiary terrane and that tectonism that occurred after the Oligocene and Miocene volcanic rocks were deposited resulted in modest renewed displacement that manifested itself in the volcanic cover.

Chemical analyses (Mariner and others, 1974) show that the springs in the vicinity of Soldier Meadow are dilute and very like Bog Hot Spring. Baltazor Hot Spring, on the other hand, includes a significantly greater proportion of dissolved material. The lineament seems to terminate in Pueblo Valley, perhaps against the fault that must be present along the east face of the Pueblo Mountains. Pueblo Valley extends north from Denio to the Alvord Desert, a distance of 50 miles; several hot springs occur centrally and along the margin of this valley system.

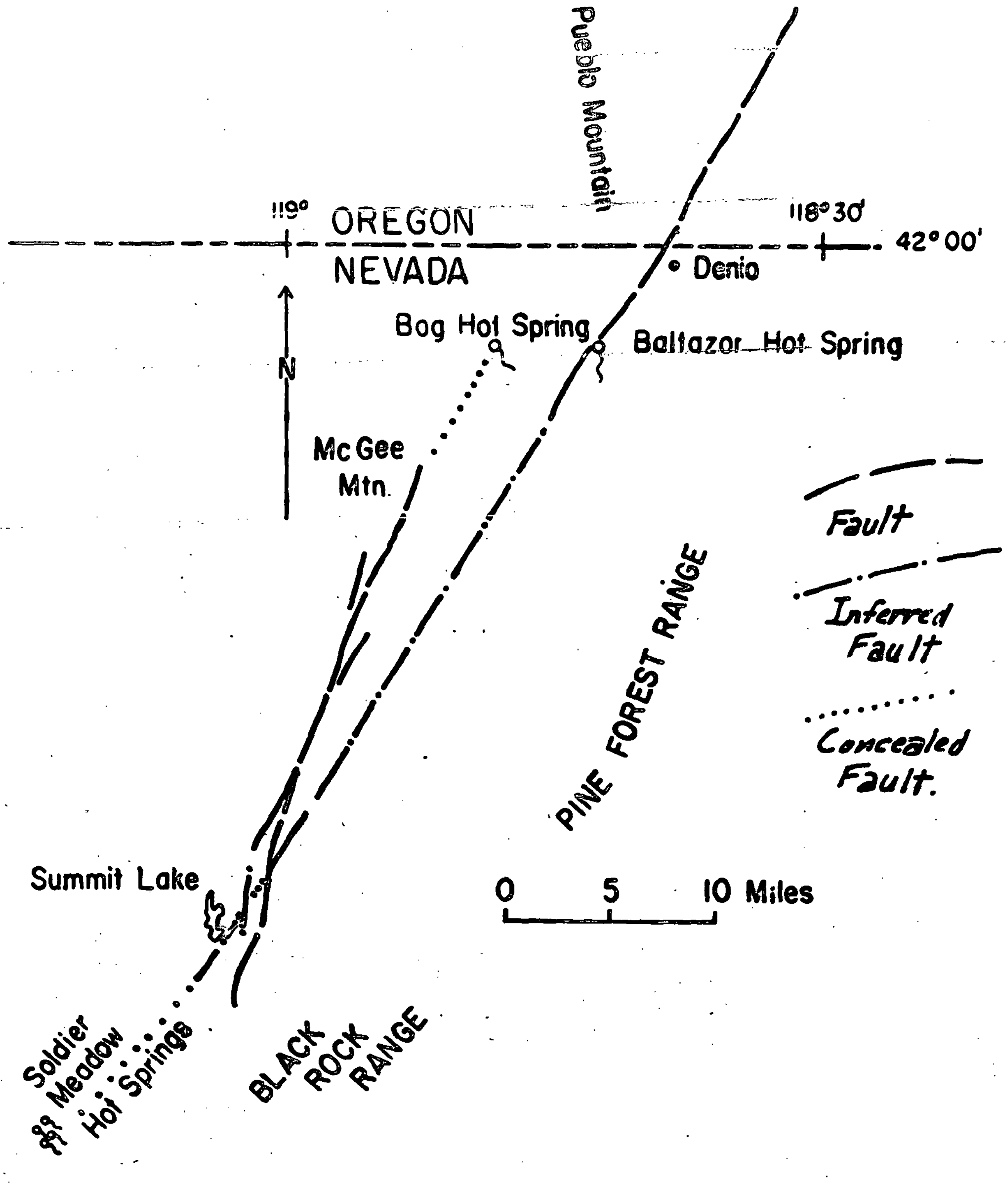

Fig. 6 Lineament in northwestern Humboldt County and its relation to hot springs.

Spring Deposits

The two principal deposits associated with hot springs of northern Nevada are travertine ($CaCO_3$) or siliceous sinter (amorphous hydrous SiO_2). Locally, mixtures of the two occur, either interlayered as in the older deposits at Pinto Hot Spring, or as silica disseminated in travertine, as at Kyle Hot Spring. Extensive silica is present at Steamboat Springs, Beowawe, and Big Sulfur Spring in Ruby Valley. A calcareous scum commonly occurs at many of the cooler hot springs in place of the more conspicuous indurated deposits.

The solubility of silica increases with temperature as the solubility of calcite decreases. This suggests that sinter deposits are indicative of higher temperature systems than those associated with travertine deposits. The observation of sinter aprons at some hot springs in table 1 having relatively high quartz temperatures supports the validity of this general rule-of-thumb.

The observation of travertine in hot spring deposits indicates that the Na-K-Ca concentrations of the spring water will not be the equilibrium values reached at depth in the system. Hence, the values given by the simple log (Na/K) geothermometer will be abnormally high. This is also supported by the data in table 1.

For calcium carbonate precipitation to occur, the following relationship must hold: $Ca^{+} \times HCO_3^{-} \times 10^{pH} > 4$ Keq (I. Barnes, oral commun., 1973). This, of course, enables $CaCO_3$ to be deposited in an acid environment. Silica solubility is a function of temperature, thus temperature drop alone is enough to precipitate sinter if initial SiO_2 content is high enough (250 ppm or higher, indicating subsurface temperatures of 180°C, is generally required).

When springs issue directly from exposed igneous-derived sediment or igneous rock, the latter two are sometimes kaolinitized, as at Steamboat Springs, Brady Hot Springs, and Beowawe, plus the fumarole 2 miles or so north of Soda Lake. It is quite probable that concealed volcanic rocks in the vicinity of hot springs are partially kaolinitized above the water table.

Temperature

The temperature of hydrothermal systems, particularly in the subsurface, is one of the principal elements of interest in connection with their potential as energy sources. Although a rough correlation apparently exists between surface temperature and the temperature in the subsurface, no method short of drilling can accurately determine the maximum temperature of a system. However, the relative abundance of certain cations in spring waters, particularly Na^{+}, K^{+}, and Ca^{++} (Fournier and Truesdell, 1973), and the amount of SiO_2 (Mahon, 1966; Fournier and Rowe, 1966) have been found to be functions of temperature and the subsurface temperatures can be estimated from chemical analyses.

TABLE 1. TEMPERATURES AND DEPOSITS OF SOME NORTHERN NEVADA HOT SPRINGS

Name	Location		Temperature1/, °C						Type of Deposit
			Surface	°Silica Conductive	Adiabatic	°log(Na/K)	°log(Na/K) + 1/3 log(√Ca/Na)	°log(Na/K) + 4/3 log(√Ca/Na)	(where recorded)
ELKO COUNTY									
Unnamed	41°59', 114°22'	sec. 9, T. 47N., R. 67E.	30	-	-	-	-	-	none
Niles Warm Spring	◆ 41°55', 114°03'	sec. 30, T. 47N., R. 70E.	43	81	84	543	220	44^	travertine
Unnamed	41°53', 114°10'	sec. 4, T. 46N., R. 69E.	45	-	-	-	-	-	travertine
Contact Mineral Spring	o 41°49', 114°49'	sec. 16, T. 45N., R. 64E.	60	127	124	78	129	103	
Rizzi Ranch Hot Spring	◆ 41°48', 115°55'	sec. 20, T. 45N., R. 54E.	41	69	74	163	166	63	
Unnamed	41°39', 115°48'	sec. 4, T. 43N., R. 55E.	42-54	92	94	255	197	111	
Unnamed	41°16', 115°16'	sec. 18, T. 39N., R. 60E.	47						travertine
Hot Sulfur Spring	◆ 41°10', 115°00'	sec. 8, T. 41N., R. 62E.	80	123	125	190	101	153	none
Unnamed	◆ 41°10', 116°00'	sec. 17, T. 38N., R. 62E.	61	140	135	188	101	124	travertine
Tuscarora	◆ 41°25', 116°03'	sec. 20, T. 39N., R. 62E.	90	167	158	184	104	133	none
Unnamed	◆ 40°42', 116°03'	sec. 33, T. 33N., R. 52E.	79	110	117	395	216	81	none
Elko Hot Spring	o 40°50', 115°46'	sec. 16, T. 34N., R. 55E.		116	113	360	234	127	
Stenier Lake	o 40°35', 115°17'	sec. 11, T. 31N., R. 60E.	45	190	176	139	170	160	sinter?
Hot Sulfur Spring	o 40°35', 115°17'	sec. 11, T. 31N., R. 60E.	93	183	171	140	181	190	sinter
East Ruby Marsh	◆ 40°15', 115°24'	sec. 2, T. 27N., R. 58E.	40-65	102 92	103	314	202	86	travertine
HUMBOLDT COUNTY									
Baltazor Spring	o 41°56', 118°42'	sec. 13, T. 46N., R. 28E.	81	165	156	163	193	125	sinter
Baltazor Well	o 41°56', 118°42'	sec. 13, T. 46N., R. 28E.	91	161^	152	159	140^	117	sinter
Bog Ranch	o 41°55', 118°47'	sec. 17, T. 46N., R. 28E.	54	163	100	28	169^	128	none
Unnamed	41°43', 118°30'	sec. 4, T. 44N., R. 31E.	43						none
Howard Hot Spring	◆ 41°43', 118°30'	sec. 4, T. 44N., R. 31E.	56	128^	125	121	110	81	sinter
Dyke Hot Spring	o 41°34', 118°34'		66	120^	125	73	137^	136	
Unnamed	◆ 41°25', 117°23'	sec. 20, T. 41N., R. 41E.	58	106	103	203	209	107	travertine
Pinto Mtns. Spring, West	o 41°22', 118°43'		91	166^	156	187^	192	203	sinter;travertine
Pinto Mtns. Spring, East	o 41°22', 118°43'		99	161^	153	145	176	169	travertine;sinter
Soldier Meadow Hot Spring	o 41°22', 119°13'	sec. 23, T. 42N., R. 24E.	51-53	113	112	34	98	65	none
Double Hot Spring	o 41°03', 119°02'	sec. 4, T. 38N., R. 26E.	81	140^	135	64	127	113	$CaCO_3$ cement
Unnamed	41°00', 119°03'	sec. 22, T. 38N., R. 26E.	74-80						
Golconda	o 40°52', 117°28'	sec. 34, T. 36N., R. 40E.	74	116	114	235	201	121	
Hot Pot	◆ 40°55', 117°07'	sec. 11, T. 35N., R. 43E.	58	125	123	300	193	150	travertine
Tipton Ranch	o 40°45', 117°23'	sec. 5, T. 33N., R. 40E.	65	160^	144	172	180	139	

WASHOE COUNTY

Fly Ranch	o 40°51', 119°19'	sec. 12, T. 34N., R. 23E.		126	124	116	160	120	tufa
Gerlach	o 40°40', 119°23'	sec. 16, T. 32N., R. 23E.	86	167$^{\Delta}$	168	170$^{\Delta}$	205	220	
Unnamed Springs	40°16', 119°22'	sec. 26, T. 31N., R. 22E.	86	-	-	-	-	-	
Pyramid Lake Well	o 40°[illegible]', 119°41'		86	149$^{\Delta}$	137	232	214	184	
Steamboat Springs	o 39°23', 119°45'	sec. 29, T. 18N., R. 20E.		201	186	180	204	203	

PERSHING COUNTY

So. Black Rock Desert	o 40°57', 118°58'		90	148	142	20	117	101	
Trego Hot Spring	40°41', 119°07'		86	128	126	51	120	111	
Leach Hot Spring	o 40°36', 117°38'	sec. 31, T. 32N., R. 39E.	97	166	147	161	176	130	sinter
Kyle Hot Spring	+ 40°24', 117°53'	sec. 12, T. 29N., R. 36E.	59	171$^{\Delta}$	161	199	194$^{\Delta}$	164	travertine
Jersey Valley	o 40°11', 117°30'	sec. 28, T. 27N., R. 40E.	34	143$^{\Delta}$	137	196	182	119	sinter
Lower Ranch, Dixie Valley	+ 40°02', 117°37'	sec. 16, T. 26N., R. 39E.	40	94	96	162	164	100	travertine

LANDER COUNTY

Unnamed	40°24', 117°03'	sec. 8, T. 29N., R. 44E.	cool						travertine
Buffalo Valley Springs	o 40°22', 117°19'	sec. 24, T. 29N., R. 41E.	65-63	97	99	164	170	227	travertine
Unnamed	+ 40°12', 117°06'	sec. 23, T. 27N., R. 43E.	63	92	94	263	207	132	travertine
Unnamed	+ 40°11', 117°05'	sec. 26, T. 27N., R. 43E.	64						
Sou Hot Spring	+ 40°05', 117°44'	sec. 29, T. 26N., R. 38E.	73	116	113	244	190	100	travertine
Spencer Hot Spring	+ 39°19', 116°50'		70	128$^{\Delta}$	125	228	204	142	travertine
Smith Creek Valley	o 39°19', 117°33'	sec. 25, T. 17N., R. 39E.	64-68	143	137	114	157	139	

EUREKA COUNTY

Beowawe Hot Spring	o 40°34', 116°35'	sec. 8, T. 31N., R. 48E.	98	214$^{\Delta}$	196	165	194$^{\Delta}$	237	sinter
Beowawe Steam Well	o 40°34', 116°35'	sec. 17, T. 31N., R. 48E.		262	226$^{\Delta}$	238$^{\Delta}$	242	292	
Hot Spring Point	o 40°24', 116°31'	sec. 11, T. 29N., R. 48E.	64	116$^{\Delta}$	115	325	233	159	
Walti Hot Spring	o 39°54', 116°35'	sec. 33, T. 24N., R. 48E.	76	117$^{\Delta}$	115	375	212	79$^{\Delta}$	travertine
Bartine Ranch Springs	39°34', 116°22'	sec. 4, T. 20N., R. 50E.	40						travertine
Hot Spring Ranch	o 39°24', 116°21'	sec. 28, T. 18N., R. 50E.	67	128	125	19	92	79	none

				WHITE PINE COUNTY					
John Salvi's Hot Spring	39°53', 115°03'	sec. 7, T. 21N., R. 63E.	66						travertine
Monte Neva Hot Spring	39°40', 115°08'	sec. 24, T. 21N., R. 63E.	70;71						travertine
				CHURCHILL COUNTY					
Dixie Valley Hot Spring	39°43', 118°00'	sec. 4, T. 20N., R. 35E.	60-73	145	153	65	144	137	none
Unnamed	39°34', 118°51'	sec. 28, T. 20N., R. 28E.	69			none (orifice covered by concrete)			
Stillwater Hot Well	o 39°32', 118°32'	sec. 6, T. 19N., R. 31E.	-	169	159	73	140	160	
Leo Hot Spring	o 39°13', 118°43'	sec. 34, T. 16N., R. 29E.	93	173$^{\Delta}$	162	128	162*	138	sinter; tufa domes
				LYON COUNTY					
Wabuska Hot Spring	+ 39°10', 119°10'	sec. 15, T. 15N., R. 25E.	97	145	139	120	152	111	none
				DOUGLAS COUNTY					
Genoa Hot Spring	o 39°00', 119°51'	sec. 9, T. 13N., R. 20E.	-	103	103	64	119	85	
Nevada Hot Spring	38°49', 119°24'	sec. 16, T. 12N., R. 23E.	-	104	104	64	119	86	
				MINERAL COUNTY					
Mina Hot Spring	o 38°22', 118°06'	sec. 8, T. 6N., R. 35E.		95	-	122	163	-	
				NYE COUNTY					
Potts Ranch Hot Spring	o 39°05', 116°30'	sec. 2, T. 14N., R. 47E.	45	87$^{\Delta}$	90	344	205	79	
Diana's Hot Spring	o 39°02', 116°40'	sec. 22, T. 14N., R. 47E.	51	98$^{\Delta}$	99	334	207	88$^{\Delta *}$	
Diana's Punch Bowl	o 39°02', 116°40'	sec. 22, T. 14N., R. 47E.	58	98$^{\Delta}$	99	341	208	86$^{\Delta}$	
Darrough Well	o 38°49', 117°10'	sec. 8, T. 11N., R. 42E.	94	140	133$^{\Delta}$	68	131	122	
Darrough Hot Spring	o 38°49', 117°10'	sec. 8, T. 11N., R. 42E.	95	136	132$^{\Delta}$	61	127	120	none
Hot Creek Valley	+ 38°23', 116°16'	sec. 30, T. 7N., R. 51E.	61	63	69	282	161	18	travertine
Unnamed	38°26', 115°47'		43	-	-	-	-	-	travertine
Warm Spring	o 38°10', 116°16'	sec. 20, T. 4N., R. 50E.	-	111	110	225	192	122	

1/ Modified from Mariner et. al. (1974)

† term sinter used here to mean opaline silica

Δ preferred (see text)

o partial analysis in field; remainder of analysis: Menlo Park, California

+ total analysis performed in Menlo Park, California

* calculated minima at depth (rounded to whole number); see Fournier and Truesdell, 1973.

Surface temperature.--Surface temperatures listed in table 1 were measured with a maximum recording mercury thermometer, generally from 10 to 50 cm below the water surface and where possible in an orifice. Values range from slightly above ambient, through boiling to superheated in a few springs.

Subsurface temperatures.--As noted above, the relative cation abundance and amount of SiO_2 in spring waters enable calculations of maximum temperature of the systems. Such calculations may reflect the last attainment of chemical equilibrium between the spring water and the conduit material encountered in the rise to the surface. Thus, it is possible that the true maximum temperature may be masked by these late interactions. Similarly, dilution by cool shallow ground waters will markedly change the chemical composition and distort the calculations of maximum temperature.

Table 1 also lists the calculated minimum subsurface temperatures based on (1) quartz solubility as developed by Mahon (1966) and Fournier and Rowe (1966), and (2) various cation ratios (Na^+, K^+, and Ca^{++}) as described by Fournier and Truesdell (1973). The chemical data from which the calculations are derived were reported by Mariner and others (1974). Calculations were performed on an IBM 360/65 computer, using a program written by Yousif Kharaka.

The temperatures calculated from the SiO_2 content assume no dilution of the ascending geothermal waters and the calculations allow for two conditions, (a) adiabatic cooling of the spring water during its ascent, or (b) conductive cooling. Generally the actual temperature of the system should lie within the limits defined by the two calculations. The adiabatic cooling-model temperature is preferred when the spring is boiling at the surface, especially if within a cluster of other near-boiling springs, and the rate of ascent to the surface is sufficiently rapid to exclude heat loss by conduction (discharge greater than ~100 liters per minute).

Contributions to the silica content of geothermal waters by amorphous silica (e.g., diatomite, volcanic glass, etc.) lead to errors in temperature calculations owing to the fact that all silica in solution is assumed to result from quartz-water reaction. Hence, knowledge of the rock column through which thermal waters circulate is required for the judicious use of this geothermometer.

Temperatures calculated from silica content of the spring waters analyzed range from 95°C to 252°C, although most are below 200°C. To a first approximation, these temperatures are roughly twice their respective measured surface values.

Fournier and Truesdell (1973) have pointed out that the simple Na/K geothermometer must be used cautiously as a temperature indicator even when the quartz temperature for the system is high. In lower and moderate-temperature systems, however, Ca enters into silicate reactions involving K and Na, and use of the simple Na/K ratio leads to temperature estimates that are commonly too high. Accordingly, the correlation between quartz- and Na/K temperatures is generally better when the quartz temperature is comparatively high.

The calcium-corrected temperatures are, in general, better indicators than the Na/K ratio except in those instances where: (1) calcium carbonate has been deposited as the thermal water ascends toward or reaches the surface, or (2) where waters are in contact with limestone or dolomite. This may result in anomalously high temperatures, and may explain some of those listed for the $\log (Na/K) + \beta\sqrt{Ca/Na}$ geothermometers ($\beta = 1/3$ or $4/3$) in table 1.

Ion exchange reactions between rocks and ascending geothermal waters result in temperature estimates that are too low. These reactions may be quite rapid, but their relative importance should, to a large degree, depend on the extent to which the rocks are prohibited from further reaction by an armor of minerals.

Geophysical Relations

One important thing to evaluate in geothermal exploration is the relationship of various geophysical parameters to geothermal systems, that is, do geophysical anomalies relate directly to known geothermal systems, and if so how can geophysics be employed to locate hidden geothermal systems?

Gravity

The gravity measurements in Nevada, generalized in figure 7, are all negative ranging from -80 to -250 milligals. Southern Nevada and northwestern Nevada have relatively high gravity, whereas the east-central part of the State has relatively low gravity. These regional differences can be interpreted as reflecting variations in crustal thickness, the crust being thinner in the southern and northwestern parts of the State. In a general way, the chemically determined maximum spring temperatures are somewhat higher in the area of relatively higher gravity, although Hot Sulfur Springs in the Ruby Valley is a noteworthy exception. Also, it should be noted that part of the area in northwestern Nevada lacking springs (fig. 3) is associated with the highest measured gravity field as well as greatest occurrence of Cretaceous granodiorite intrusive rock.

Aside from the broad regional variations in gravity, there is a clear association of gravity highs with mountain ranges and gravity lows with basins. The steepest gradients are approximately coincident with the mountain fronts which are in themselves an expression of faults. Hot springs have a high frequency of occurrence along steep local gradients, suggesting that they rise along concealed Basin and Range faults that flank the mountain ranges.

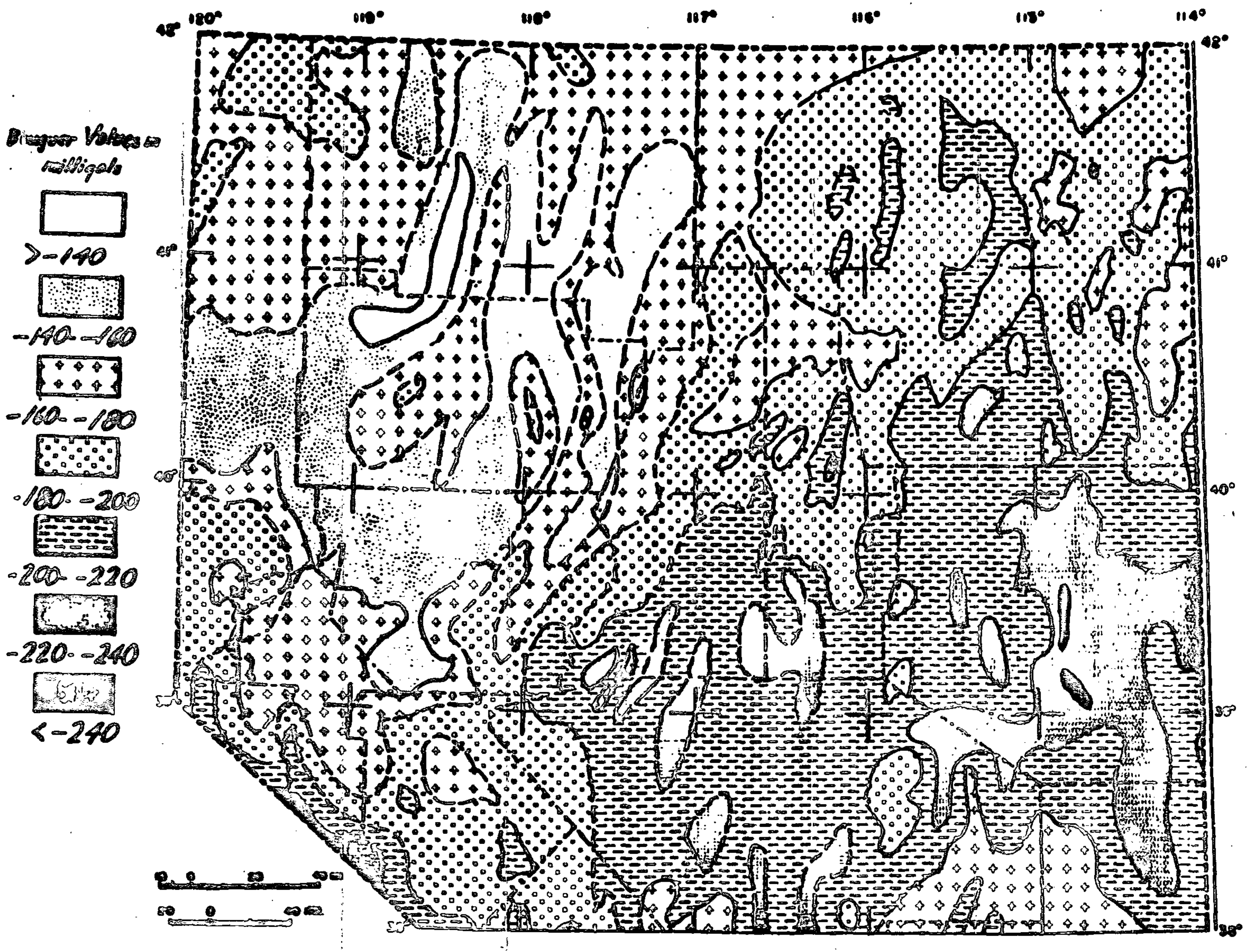

Fig 7 Bouguer anomaly map of northern Nevada, adapted from Woollard and Joesting (1964)

Detailed gravity measurements by D. L. Peterson at Buffalo Valley hot springs and Leach Hot Springs and by Peterson and D. L. Mabey at Hot Pot (D. L. Mabey, personal commun., 1973) disclose a positive 5-to-10-milligal anomaly associated with the springs. Mabey suggests the anomaly results from subsurface silicification. Addition of calcium carbonate to the valley fill material might also contribute.

Magnetics

Aeromagnetic surveys of northern Nevada delineate many intrusive bodies and expanses of volcanic rocks. These areas generally are so large that any anomalies resulting from associated thermal systems are completely masked. It is possible that detailed work in specific areas might disclose an understandable relationship, but none is detectable at this time.

In general, there seems to be a gross correlation between the aeromagnetic anomalies and the gravity. That is, the areas of lowest gravity have a lower density of aeromagnetic anomalies, and the areas of high gravity have a higher density of magnetic anomalies.

Seismicity

Epicenters of earthquakes for the period 1932-1968 show a limited correlation with hot spring occurrence. There are essentially three clusters of earthquake epicenters for this period, one in the Reno-Carson City area (1 on fig. 8), one southeast of Hawthorne (2), and the Fairview Peak-Dixie Valley area (3). A few obvious relationships include: (1) Steamboat Springs and Moana Hot Spring and a few other springs occur within the Reno-Carson City cluster; (2) hot springs are present in Dixie Valley and Lee Hot Spring occurs along the western edge of the Fairview Peak-Dixie Valley cluster; (3) no high-temperature hot springs are located within the Hawthorne cluster of epicenters; and (4) an epicenter lies on the fault which is the locus of the Double Hot Spring system. Otherwise, there is apparently little coincidence between hot spring occurrence and seismicity; indeed, some of the hottest calculated SiO_2-temperature waters, such as Big Sulfur Hot Springs, Baltazor, Leach, and others, are far removed from earthquake epicenters of the 1932-1968 period. Since most of the springs appear to be fault controlled, however, the probability of their association with prehistoric seismicity (tectonism) is high.

Ward and others (1969, 1971) have suggested a relationship between microearthquakes and geothermal systems based on studies of Iceland and El Salvador. These areas have relatively high seismicity and active volcanism, factors that undoubtedly bear on the association. Any relation between the hydrothermal systems of Nevada and microearthquakes or microseisms remains untested.

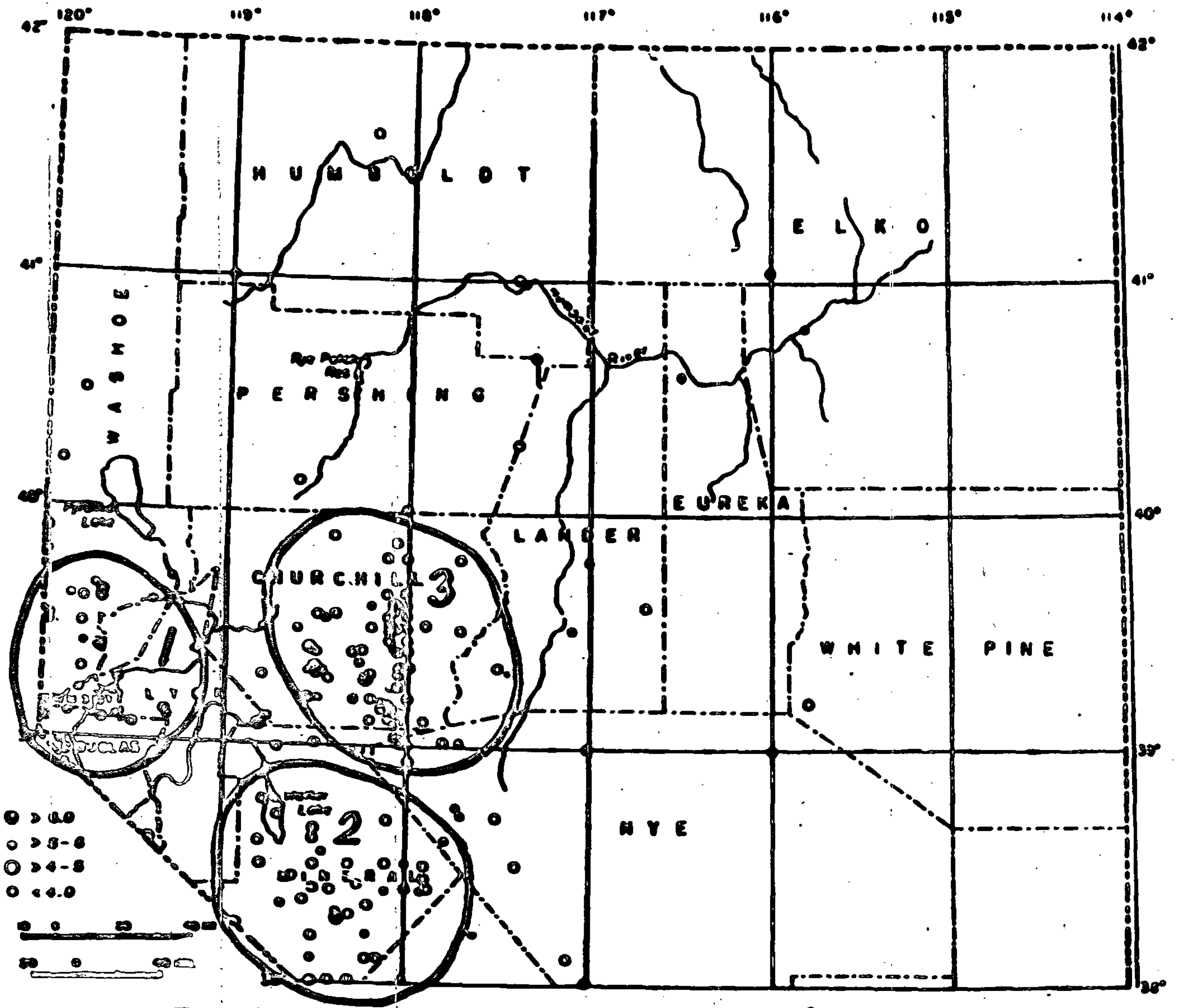

Fig. 8 Earthquake epicenters 1932 to 1968 (from Slemmons et al. 1965, and U.S. Coast and Geodetic Survey tabulation of U.S. Earthquakes, 1962 – 1968)

Heat Flow

Roy and Blackwell (1966), Roy and others (1968), Blackwell (1971), and Sass and others (1971) have established firmly that Nevada is a region of greater than average heat flow for continents. Sass and others (1971) have shown that a large region of abnormally high heat flow (2.5-3.8 heat-flow units)[3] is present in the vicinity of Battle Mountain in north-central Nevada (fig. 9), and that a large area south of Eureka has a low mean heat flow less than 1.5 hfus. This low heat flow is attributed to cooling by moving ground water (Sass and others, 1971).

Nearly all of the published heat-flow values were obtained from boreholes drilled in bedrock within the mountain ranges. There is a variation in the heat flow, conductivities, and gradient from range to range as expressed, for example, by gradients of 27°-34° C/km in the Shoshone Range, through 31°-35° C/km in Battle Mountain, to 42°-64° C/km in the Sonoma-Tobin Range area, all within the region of anomalously high heat flow.

Holes in the basins have been drilled but the heat flow results have been published for only one (Roy and others, 1968, Lovelock). This hole has a heat flow comparable to the mountain ranges but a low conductivity results in a gradient of 91.5°C/km, representing a heat flow of 2.5. A hole in the northern part of the Carson Sink has a corrected heat flow of 1.9 and a geothermal gradient of 77°C/km (Sass, oral commun., 1973). The higher gradient in the basins undoubtedly results from the much lower conductivity of the less well indurated basin fill. Variation of heat flow from place to place testifies strongly to the anistropy of the structures, and rocks and resultant thermal refraction. Geothermal gradients in the mountain ranges, because of higher conductivities, are generally lower than gradients in the fill portions of the basins.

Roy and others (1968) attribute the high heat flow to a thin crust and high-temperature mantle, while Blackwell (1971) attributes about 10 percent of the heat flow to "penetrative convection of material from the mantle to a shallow level in the crust."

Characteristics of Known Geothermal Systems

The known geothermal power-producing areas of North America as well as those with a known high potential for geothermal power production (e.g., Yellowstone Park, Wyo.; Jemez Mountains, N. Mex.; Imperial Valley, Calif. and Mexico; Long Valley, Calif.; and The Geysers, Calif.) have three

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[3] 1 heat-flow unit (hfu) = 10^{-6} cal cm^{-2} sec^{-1}, heat flow = conductivity x geothermal gradient.

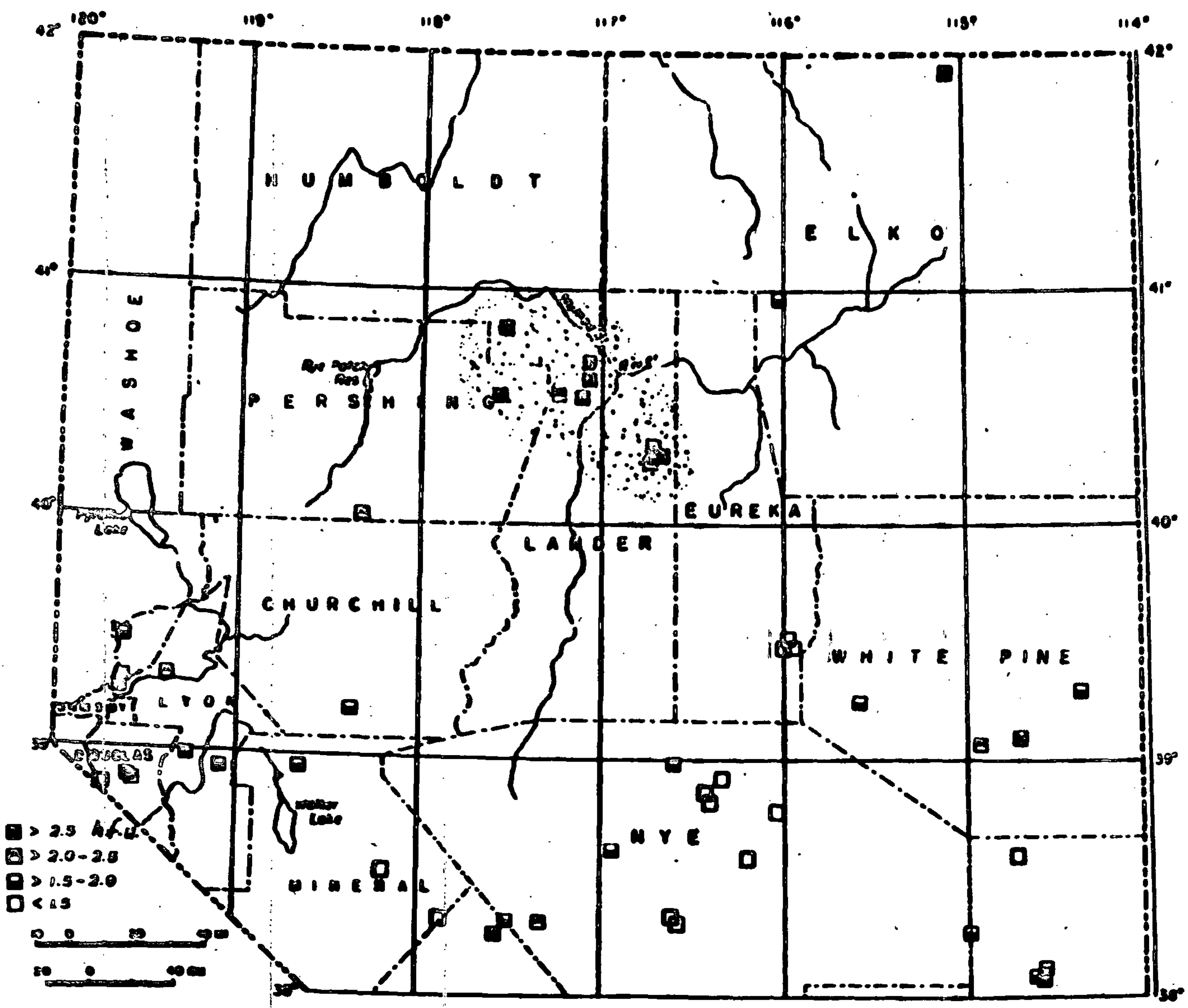

Fig. 9 Heat flow, modified from Sass et al. (1971) and Roy et al. (1968); Stippled area is the Battle Mountain heat flow

fundamental characteristics in common. They are (1) directly related to volcanic rocks of Quaternary age, (2) the volcanic rocks are felsic to intermediate in composition, and (3) all areas have relatively high seismicity. In addition, the thermal systems of Yellowstone Park, the Jemez Mountains, and Long Valley are principally within calderas, a tectonic feature commonly related to eruption of large volumes of silicic rock from shallow levels in the crust. It is reasonable to assume, because of the first two associations, that the heat is derived from either a molten magma or one only recently frozen but that is still hot (or from hot wall-rocks). At The Geysers, Calif., Garrison (1972) has implied a magma body as the source of the volcanic rocks at Clear Lake. An open-file gravity map (Calif. Div. Mines and Geology, 1966) shows a regional circular low over The Geysers Steam Field and a short text accompanying that map postulates a magma body as the source of the low. Conventional explanations of the origin of hot springs involve the circulation of meteoric waters down along faults to the limit of permeability where by convection they remove heat from the rock. The cool descending water more than counterbalances the more buoyant warmed waters and the latter rise to the surface. The permeable and underlying relatively impermeable rocks are presumably heated by conduction from a deeper magma. This magma would be rather high in the crust at depths of more than 6-8 km and would generate a very high local thermal gradient and high heat flow.

Origin of Northern Nevada Hot Springs

On the assumption that many of the silicic volcanic rocks of northern Nevada were erupted from shallow magma chambers and that hot magma may exist within or below the caldera usually formed after some eruptions, attempts have been made to calculate the time required for magma chambers of various sizes and depths in the crust to cool to the original temperature of the country rock. Calculations by A. H. Lachenbruch (oral commun., 1973) indicate that all magma chambers for the 10-m.y. or older silicic rocks have long ago lost their heat and should now have temperatures in the range required by the present Basin and Range thermal gradients. Assuming that the intrusives have cooled and are "dead," they cannot now be considered as an important heat source.

The abundant Quaternary and late Pliocene basalts in northern Nevada give rise to the question as to whether or not still-hot magma chambers could be associated with them. Basalts are the most common volcanic rock in the world, yet their plutonic coarse-textured equivalent, gabbro, is volumetrically uncommon. Most plutonic rocks are of felsic composition despite the fact that their eruptive equivalents are less common than basalt. These relationships, plus their chemical similarity to mantle rocks and their lack of crustal-rich elements such as K and Rb, imply that basaltic magmas find their way to the surface directly from the mantle without developing magma chambers of significant size at intermediate or shallow levels in the earth's crust. Harris and others (1970) provide a summary of the physical characteristics of basaltic magmas that enable them to erupt.

Although basaltic magmas have considerably greater heat content than their felsic counterpart, it is unlikely that their low viscosity would allow them to form magma chambers.

In the main, we subscribe to the conventional explanations for hot springs, but with one significant difference. Since there are probably no "still-hot" plutons or magma bodies high in the crust of northern Nevada, a heat source other than shallow intrusive bodies must be sought. We suggest that meteoric waters moving downward deeply along Basin and Range faults would, in areas of high geothermal gradients cited earlier, encounter high-temperature rock, would become heated, and rise to the surface. The fundamental difference being that the heat is derived from rock that is hot by virtue of high regional thermal gradient, imposed by a high-temperature upper mantle only partially blanketed by the thin (29-30 km) crust, rather than from a magma at some level within the crust.

The depths at which the fault gouge becomes impermeable to meteoric waters under hydrostatic pressure is critical to our model. Intuitively, it is reasonable to assume that owing to lithostatic pressure the rocks at some depth are impermeable to water under lesser hydrostatic pressures, but to our knowledge data bearing on this critical depth do not exist. If such a depth were real, it would probably be variable from place to place and be dependent on the genre of the rocks in the gouge zone, the time at which the faulting took place, the load, and to some extent on the chemistry of the water.

On the basis of rather indirect evidence we feel the critical depths of meteoric water penetration in Nevada is at least 3 km.

A 13,000-foot-deep well in southern Nevada has a low thermal gradient to a depth of 10,000 feet and a higher gradient below. The lower gradient is interpreted by Sass and others (1971) as resulting from the cooling effect of ground-water movement through basin fill and older rocks. The steeper gradient may be interpreted as resulting from an impounded static body of water in permeable rock or it may result from impermeability. In either case the rock is permeable to meteoric ground water at least to a depth of 10,000 feet (3 km). Differences in the inherent permeability of fault gouge as compared to other rocks could be expected to modify the absolute maximum depth of penetration of meteoric waters.

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The Tertiary geologic history began with local accumulation of nonmarine strata to be followed in late Eocene time by the beginning of a period of volcanism that persisted into the Holocene.[1] The main volcanism took place from the Oligocene to middle Pliocene time and was so extensive that volcanic rocks now comprise nearly 25 percent of the outcrops of northern Nevada.

The Tertiary volcanic rocks are mainly silicic to intermediate in composition but include basaltic intercalations especially in the northern part of the province near the Snake River Plain. These rocks include lava flows, ash flows, some air-fall tuffs and volcanogenic sedimentary units. Although the vents or eruptive centers for some of the volcanic rocks have been recognized, the source for many is unknown. Some known source areas from which large volumes of rock were erupted are: the southern Fish Creek Mountains described as a volcanic center and collapse caldera (McKee, 1970); the Clan Alpine Mountains, a complex eruptive center (Riehle and others, 1972); the Northumberland caldera (McKee, 1974) near Northumberland Canyon on the west side of the Toquima Range; a volcano-tectonic depression (caldera structure filled with eruptive rocks) in the northern part of the Toiyabe and Shoshone Ranges (Masursky, 1961); a caldera and associated volcanic rocks west of McDermitt, Nev. (Yates, 1942; Walker and Repenning, 1965, 1966); and a series of linear vents from which a large volume of ash-flow tuff was erupted in Washoe and Humboldt Counties, Nev. (Korringa and Noble, 1970).

The Pliocene to Holocene volcanic rocks, found mostly near the northern and western margin of the region are principally basaltic although there are bodies of rhyolite and andesite as well. Late Pliocene to Holocene basaltic rock distribution is shown on figure 1.

The present elongate northerly trending mountain ranges of high relief, flanked by more or less flat-bottomed valleys or basins, are the products of tectonism that probably began as long ago as the Oligocene in some places, but in most regions started in middle Miocene (15-17 m.y. ago), and continues today. This widespread rifting caused extension of the entire province and produced the steep fault scarps that occur on either or both sides of the fault-block mountain ranges. The total extension across the Basin and Range province has been estimated by Thompson and Burke (1973) on the basis of geologic and geophysical work in Dixie Valley, Nev., to be about 100 km. Since the ranges and basin are relatively uniformly distributed across the province, Stewart (1971) has suggested deep-seated extension of a plastic substrate.

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[1] We use the following time division of the Tertiary which is based on mammalian chronology (Evernden and others, 1964): Quaternary, 0-2 m.y.; Pliocene, 2-12 m.y.; Miocene, 12-26 m.y.; Oligocene, 26 to 37-38 m.y.; Paleocene and Eocene, 38-65 m.y.