A DIPOLE-DIPOLE RESISTIVITY SURVEY
OF THE ROOSEVELT HOT SPRING PROSPECT
BEAVER COUNTY, UTAH

submitted to

GETTY OIL COMPANY

September, 1976

by

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Project No. 76.120

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ABSTRACT

A dipole-dipole resistivity survey of 15 line miles was performed over the Roosevelt Hot Springs Prospect, Beaver County, Utah, for the Getty Oil Company by Geonomics, Inc.

Most of the prospect is underlain by high resistive strata corresponding to metamorphic or granitic rock. However, the prospect area is dissected by numerous faults, one of which may be a southern extension of the Dome Fault.

Three highly conductive (low resistivity) zones were located by the survey (Plate III), two of which appear very attractive in terms of geothermal potential. The third zone is not as attractive, but still merits further investigation.

Geonomics recommends that temperature gradient holes be drilled in all conductive zones found, in order to see if any warrant a deep exploration test hole.

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INTRODUCTION

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The survey consisted of 15 miles of dipole-dipole lines, which were taken on five different sections in T27S, R9,10W (see Plate I). The lines were surveyed using 300 m (984 ft) dipole separations, with an effective probing depth to N = 10.

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Dipole-dipole data provide a two-dimensional cross-section of the resistivity as a function of electrode separation and are presented as a series of pseudo-sections. The method of dipole-dipole presentation is to plot the resistivity value at the pseudo depth, representing the intersection of two 45° lines emanating from the center of the current dipole and the receiver dipole. As a result, the pseudo-section, although convenient, causes a distortion of geologic structure. Thus, a vertical fault is represented on the dipole-dipole pseudo-section as a slanted contact at an angle of 45°. Experience based upon computer modeling must be employed in the interpretation of dipole-dipole data. The five dipole-dipole section locations are shown on Plate I.

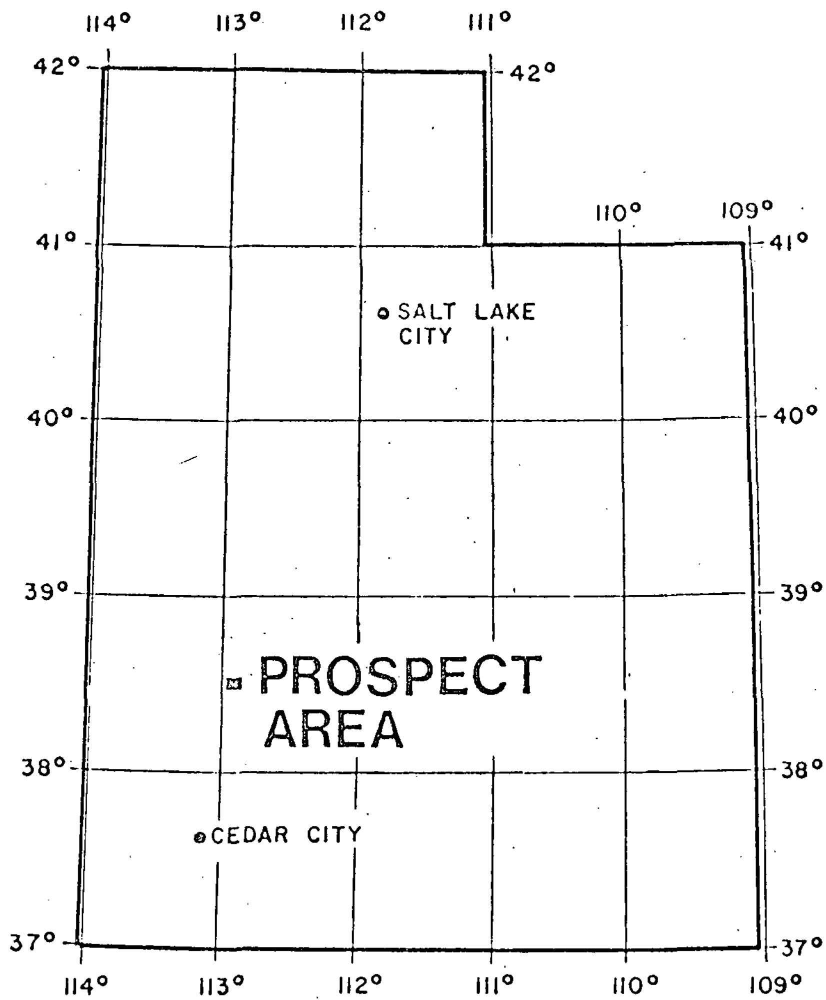

Figure 1. Location map of the Roosevelt Hot Spring Prospect.

GEOLOGICAL AND GEOPHYSICAL SETTING

The Roosevelt Hot Springs Prospect lies in the western foothills of the Mineral Range in Beaver County, Utah (see Figure 1). The town of Milford, Utah is located approximately five miles southwest of the prospect area.

The Roosevelt Hot Springs area is a part of the Basin and Range physiographic province which is characterized by broad sediment-filled valleys flanked by large, north-trending mountain ranges. This present configuration is due largely to Cenozoic block faulting which is primarily responsible for the north-south elongation of the ranges and valleys of Nevada. Thompson (1959) concludes that the faulting must be the result of an east-west extension, which may be on the order of 30 miles.

The most pronounced topographic and geologic feature of the Roosevelt Area is the Mineral Range (see Figure 2 and Plate II), which is an isolated, plutonic horst principally composed of white granite (Liese, 1957; and Earll, 1957). Radiometric age determinations have been made on the Mineral Range, indicating that parts of it are Miocene to early Pliocene (Park, 1968; and Armstrong, 1970). Silicic flows of late Pliocene age cap the Mineral Range, with two flows extending into the immediate area of Roosevelt Hot Springs (see Plate II). Liese (1957) and Earll (1957) have mapped a zone of Precambrian (?) metamorphic rocks at the base of the Mineral Range. Biotite gneiss makes up most of the exposed metamorphic rocks, but some schists and phyllites are also present. The age of Precambrian (?) was assigned to the rocks by Earll (1957) principally because of their metamorphic grade.

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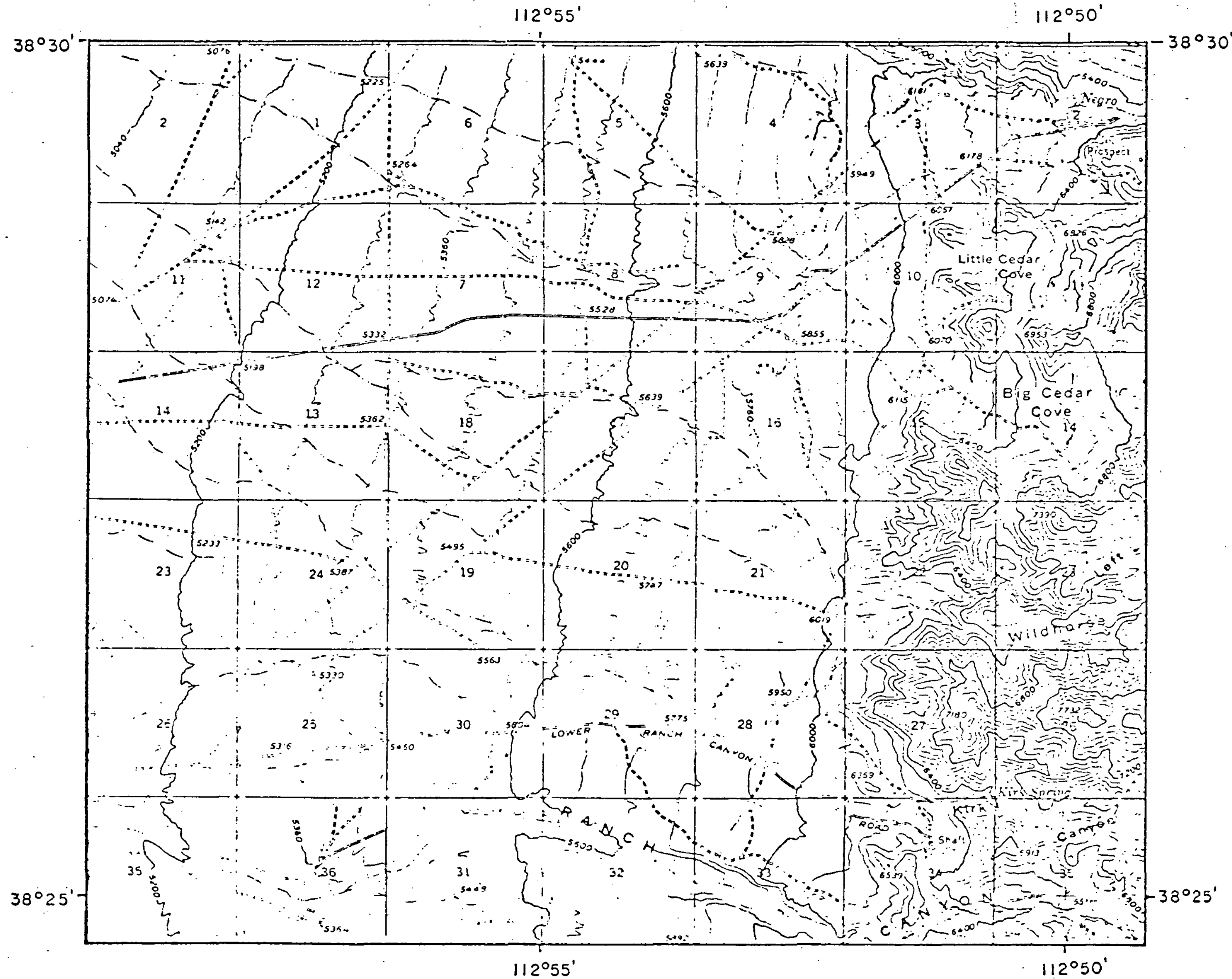

Figure 2. Topographic map of the Roosevelt Hot Spring prospect area.

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The Mineral Range is flanked on the east by Beaver Valley and on the west by the Escalante (or Milford) Valley. According to Cook and Mudgett (1966) the eastern side of the Escalante Valley graben contains a maximum of about 5,000 ft of valley fill.

The Roosevelt Hot Springs area is cut by numerous east-west and north-south trending faults, with the latter being the most predominant type of faulting (see Plate II). The Dome Fault, so named because its maximum offset is in some domed, siliceous hot spring deposits, is the most conspicuous fault in the Roosevelt area, extending in a north-northeasterly direction. The vertical displacement of the fault is at least 15 feet, with the west block moving up relative to the east block (Peterson, 1975).

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2. the hot springs dried up because the water table within the Escalante Valley has lowered in recent time.

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A hole drilled to a depth of 273 feet in the NW 1/4 NE 1/4 of Sec. 16, T27S, R9W encountered steam under some pressure with a temperature of 270°F. Two other drill holes are located on the upthrown block of the Dome Fault. One hole, which is still open to a depth of 50 feet, has a measured temperature of 140°F. The other, which was drilled to a depth of 80 ft, was known to have boiling water associated with it at depth.

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TABLE I

Analyses of Water from Roosevelt Hot Springs
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Potassium (K)	102	472	488
Bicarbonate (HCO_3)	30	158	156
Sulfate (SO_4)	90	65	73
Chloride (Cl)	87	3,810	4,240
Fluoride (F)	[1]	7.1	7.5
Nitrate (NO_3)	1.83	1.9	11
Boron (B)	[1]	[1]	38
Lithium (Li)	[1]	[1]	0.27
Iodide (I)	[1]	[1]	0.3
Residue on evaporation at 180°C	645	[1]	[1]
Calculated dissolved solids	[1]	7,040	7,800
pH	[1]	[1]	7.9
Location	sec.? T27S, R9W	sec.34 dcb, T26S, R9W	sec.34 dcb, T26S, R9W
Sampled and analyzed by	USGS	USGS	USGS
Source of data	Lee, 1908 p.50	Mundorff, 1970, p.16-17	Mundorff, 1970 p.16-17

[1]Not determined
dcb = SE 1/4 SE 1/4 NW 1/4

(from Peterson, 1975)

to a depth of 2,728 ft (762 meters), and encountered initial flows of steam in excess of 200,000 pounds per hole at a temperature of over 400°F.

Electrical Surveys

An extensive dipole-dipole survey was performed in the Roosevelt Hot Springs area by Ward and Sill (1976). They reported that the area immediately north of this prospect area is extensively cut by numerous north-south and east-west trending faults (see Interpretation Map, Plate III). In their conclusions they state that the heat source of the convective hydrothermal system of Roosevelt Hot Springs area may lie to the west of the area. They also conclude that the hydrothermal system must be leaking westward along some unmapped east-west fractures. They have concluded this by the use of thermal gradients (see above) and the extremely low resistivities that were encountered in the western foothills of the Mineral Range.

THE PRESENT SURVEY

Five dipole-dipole lines, totaling over 15 miles, were surveyed for the Roosevelt Prospect (see Plate I). The lines, labeled Line 1 through Line 5, used 300 meters (984 feet) dipole spacings, and were surveyed in a manner allowing the greatest delineation of subsurface structure. An intuitive interpretation for each of the five dipole-dipole lines (or pseudo-sections) is shown on figures in the following order: Line 1 on Figure 3; Line 2 on Figure 4; Line 3 on Figure 5; Line 4 on Figure 6; and Line 5 on Figure 7. The Interpretation Map (Plate III) is a combination of the results of this survey, with the findings of other geological and geophysical studies.

The Interpretation Map (Plate III) and the interpreted pseudo-sections can be summarized as follows:

1. The study area is dissected by numerous faults or discontinuities that appear to trend in a predominantly northerly direction. Two major faults, one of which appears to be a southern extension of the Dome Fault, are seen intersecting Line 1 (Figure 3) and Line 2 (Figure 4). Other major faults or discontinuities are shown on Line 5 (Figure 7); two of these coincide with faults mapped by Ward and Sill (1976).

2. Three highly conductive subsurface zones have been located within the study area (Areas I, II, and III, Plate III) Area I occurs at the intersection of Line 1 and Line 5. Area II occurs near the northeast end of Line 5, and Area III is located at the northern end of Line 4.

3. A near surface resistive block intersects approximately 500 m of Line 3 and appears to intersect a portion of Line 1. The resistive block has the electrical expression of a subsurface dike (see Plate II), but the true nature of the block will be unknown until further work is accomplished in the area.

DISCUSSION

The results of the dipole-dipole electrical survey are presented as apparent resistivity pseudo-sections in Figures 3 through 7. Significant resistivity contrasts are denoted as faults or discontinuities on the pseudo-sections, with the approximated surface extension of the faults or discontinuities also denoted.

Line 1 and Line 5 (Figures 3 and 7 respectively), the longest dipole-dipole lines, along with Line 2 (Figure 4) indicate that the apparent resistivity values decrease to the southwest, and that the lowest apparent resistivity values (<15 ohm-meters) lie beneath the intersection of the two lines (Area I, Plate III). The highly conductive strata of this area can be caused by a variety of factors, including salinity variation, temperature variation, clay content, and water saturated sediments. Because of the close proximity of Area I to the Escalante Valley, all these factors have a probable influence on the resistivity of the area, with clay and water-saturated sediments considered to have the greatest effect. The effect of temperature on the apparent resistivity values of this area cannot be determined from the pseudo-sections. The highly conductive zone of Area I lies to the west of a major fault or discontinuity. The effects of this fault or discontinuity are most clearly seen from Line 1 (Figure 3) and Line 2 (Figure 4); further, the Interpretation Map (Plate III) indicates that it could be a southern extension of the Dome Fault. In any case, this fault or discontinuity appears to divide the highly conductive rock of the west from the highly resistive rock to the east in this area. Undoubtedly, the highly resistive strata is that of the Mineral Range, and is either granitic or metamorphic in character.

Area II (Plate III) is a near-surface conductive zone. It is located near a well drilled in opal deposits in the NW 1/4 NE 1/4 of Sec. 16. This well hit steam at a depth of 275 feet (see section on Geological and Geophysical Setting), at a temperature of 270°F. It is possible, therefore, that the conductive zone of Area II is an electrical expression of geothermal resource, although it is nearly one mile from the steam well. The hot water of this area could be migrating westward along unmapped east-west fractures.

Area III is a conductive zone which is found deeper in the subsurface. Since the northern limit of this area was not defined, it is possible that this zone is a southern extension of Area II. If this is the case, the entire area surrounding Area II and Area III may be considered as a possible geothermal area warranting further evaluation.

CONCLUSIONS AND RECOMMENDATIONS

Three areas of highly conductive strata have been found within the Roosevelt Hot Springs Prospect (Areas I-III, Plate III). Areas II and III appear to be the most attractive targets for further geophysical work because of a well-defined electrical expression and close proximity to a steam-producing well (NW 1/4 NE 1/4 of Sec. 16, see Plate II). Shallow thermal gradient holes in the immediate area of these two anomalies will give a more clear, and inexpensive, interpretation of the geothermal potential of the area.

Area I appears to expand toward the Escalante Valley and is probably due to water-saturated valley fill. However, the Dome Fault appears, from previous drill hole surveys, to be a channelway for migrating geothermal fluids. With this in mind, one or two shallow thermal gradient holes drilled along the Dome Fault in this area (see Plate III) may prove extremely profitable. The client would have to assess the economic feasibility of such a survey.

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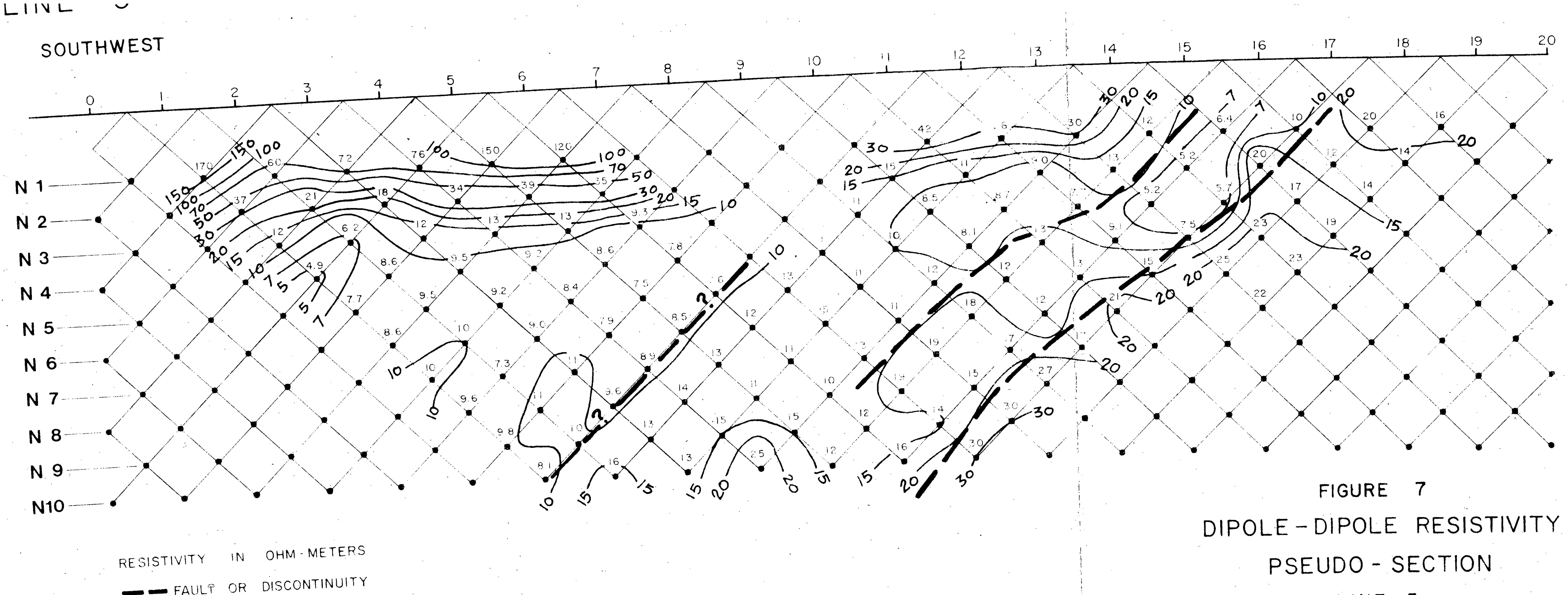
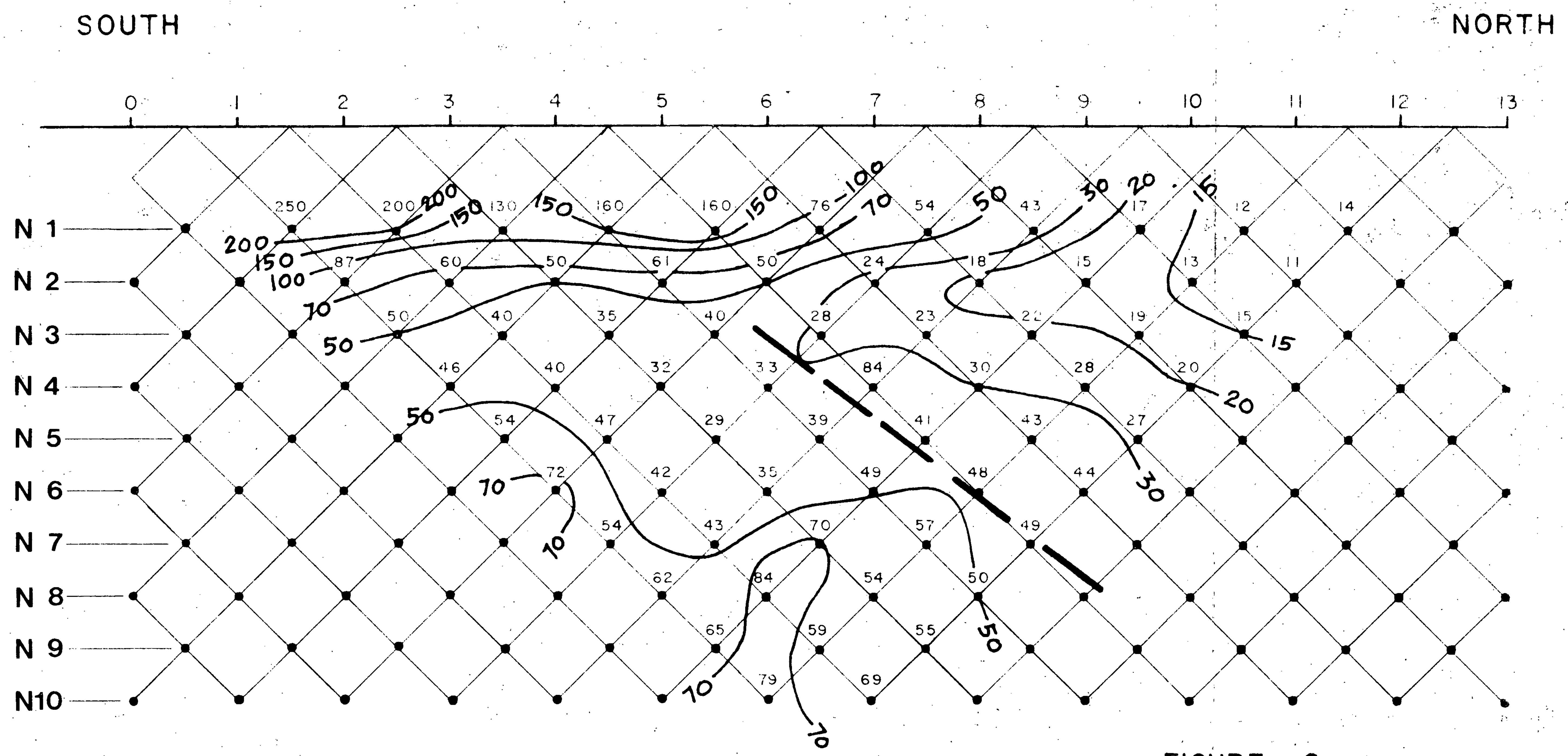


FIGURE 7

DIPOLE-DIPOLE RESISTIVITY PSEUDO-SECTION

LINE 5

NV-1523-211-2



RESISTIVITY IN OHM-METERS

FAULT OR DISCONTINUITY

FIGURE 6

DIPOLE-DIPOLE RESISTIVITY

PSEUDO-SECTION

LINE 3

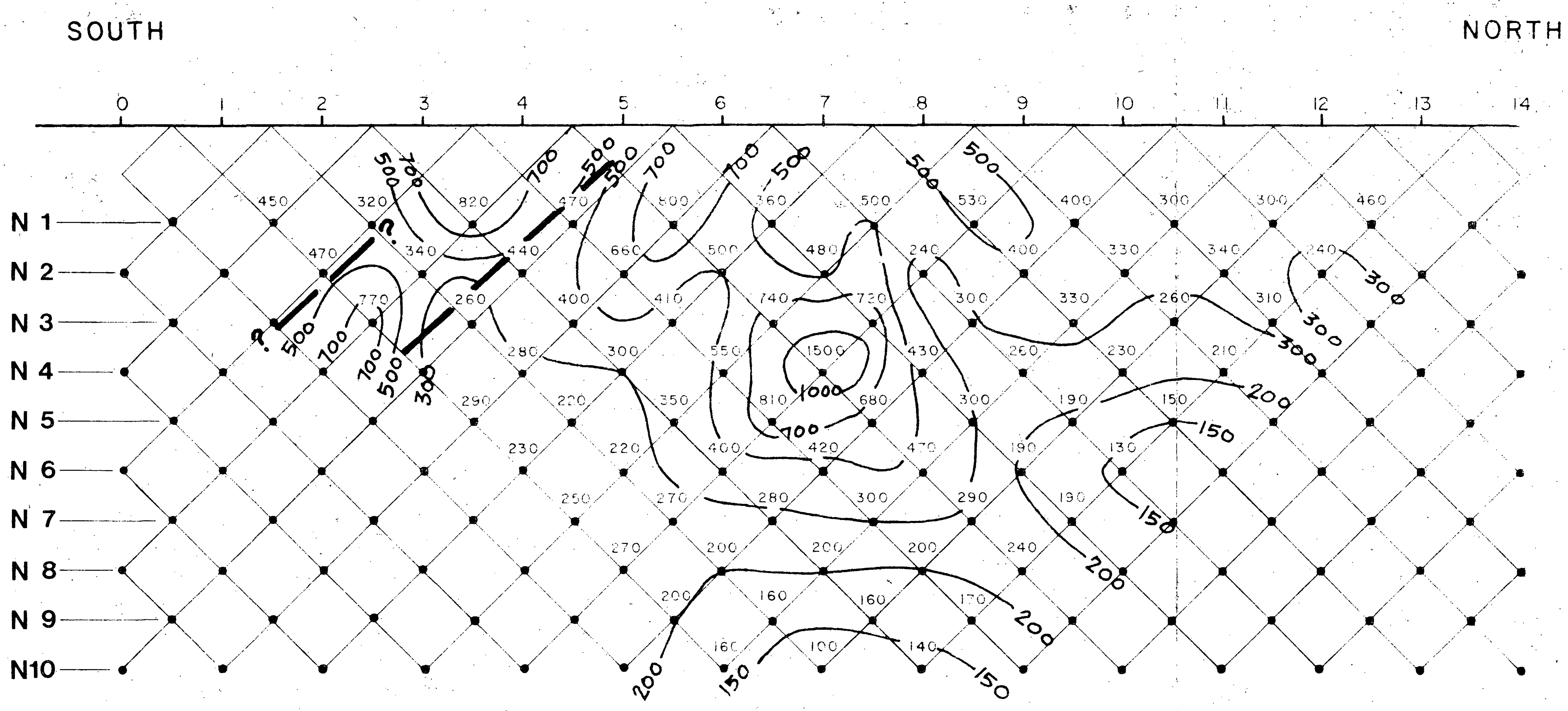


FIGURE 5

DIPOLE - DIPOLE RESISTIVITY

PSEUDO - SECTION

RESISTIVITY IN OHM-METERS

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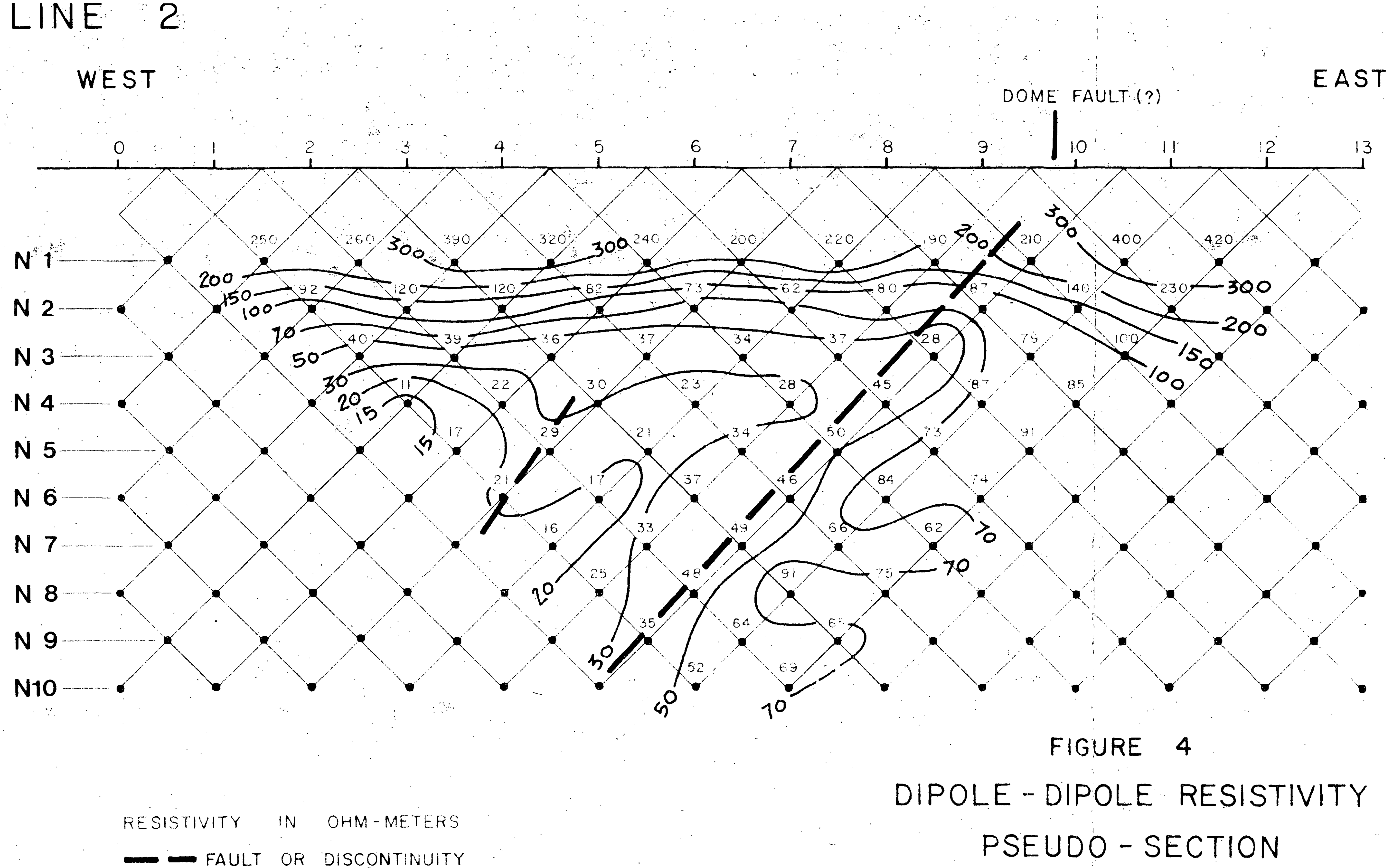
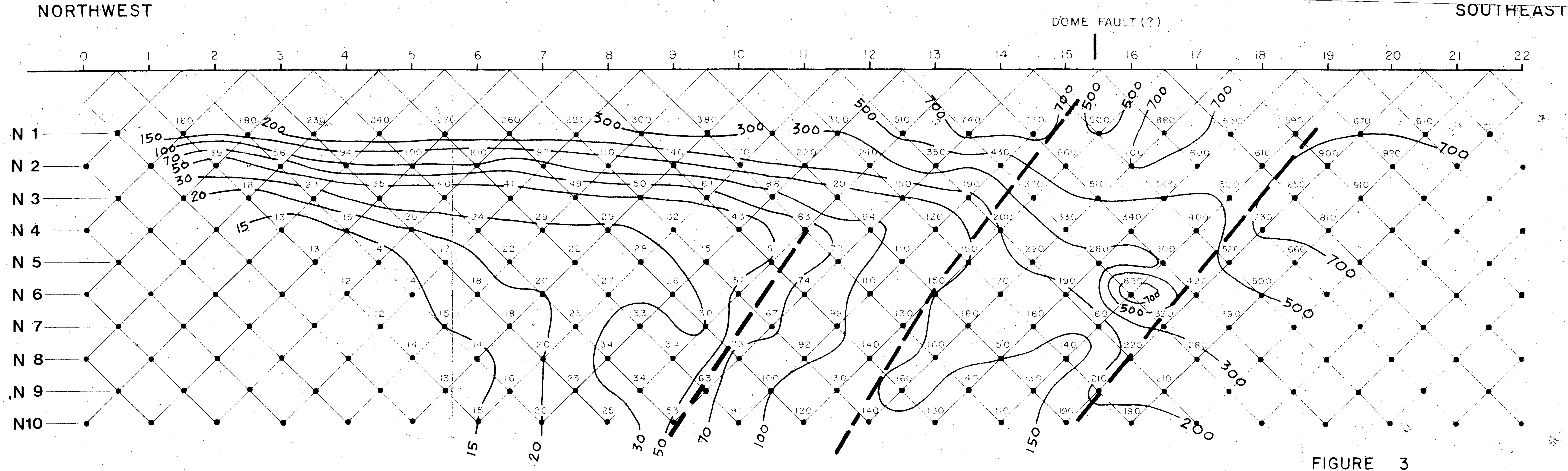


FIGURE 4

DIPOLE - DIPOLE RESISTIVITY

PSEUDO - SECTION



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DIPOLE - DIPOLE RESISTIVITY

PSEUDO - SECTION

9 p.
19 p

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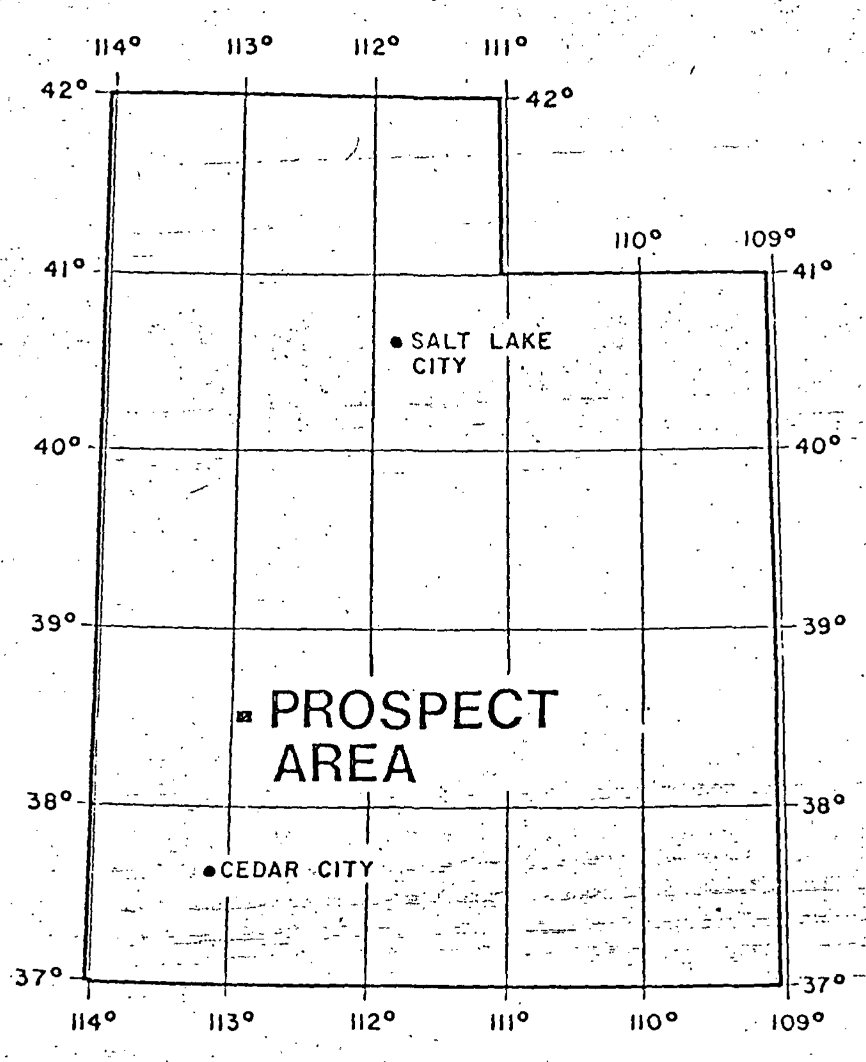

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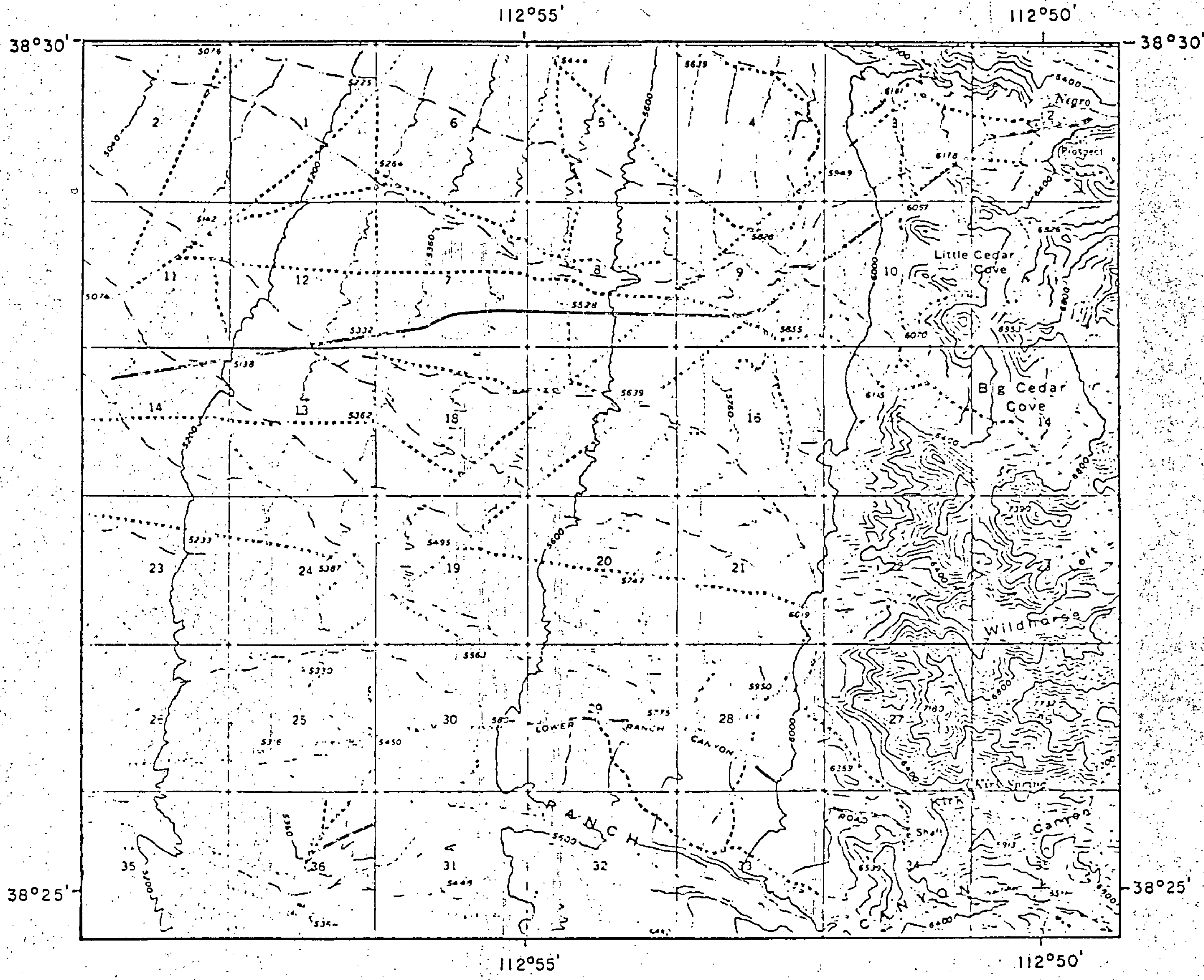

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Nitrate (NO_3)	1.83	1.9	11
Boron (B)	[1]	[1]	38
Lithium (Li)	[1]	[1]	0.27
Iodide (I)	[1]	[1]	0.3
Residue on evaporation at 180°C	645	[1]	[1]
Calculated dissolved solids	[1]	7,040	7,800
pH	[1]	[1]	7.9
Location	sec.? T27S, R9W	sec.34 dcb, T26S, R9W	sec.34 dcb, T26S, R9W
Sampled and analyzed by	USGS	USGS	USGS
Source of data	Lee, 1908 p.50	Mundorff, 1970, p.16-17	Mundorff, 1970 p.16-17

[1]Not determined
dcb = SE 1/4 SE 1/4 NW 1/4

(from Peterson, 1975)

to a depth of 2,728 ft (762 meters), and encountered initial flows of steam in excess of 200,000 pounds per hole at a temperature of over 400°F.

Electrical Surveys

An extensive dipole-dipole survey was performed in the Roosevelt Hot Springs area by Ward and Sill (1976). They reported that the area immediately north of this prospect area is extensively cut by numerous north-south and east-west trending faults (see Interpretation Map, Plate III). In their conclusions they state that the heat source of the convective hydrothermal system of Roosevelt Hot Springs area may lie to the west of the area. They also conclude that the hydrothermal system must be leaking westward along some unmapped east-west fractures. They have concluded this by the use of thermal gradients (see above) and the extremely low resistivities that were encountered in the western foothills of the Mineral Range.

THE PRESENT SURVEY

Five dipole-dipole lines, totaling over 15 miles, were surveyed for the Roosevelt Prospect (see Plate I). The lines, labeled Line 1 through Line 5, used 300 meters (984 feet) dipole spacings, and were surveyed in a manner allowing the greatest delineation of subsurface structure. An intuitive interpretation for each of the five dipole-dipole lines (or pseudo-sections) is shown on figures in the following order: Line 1 on Figure 3; Line 2 on Figure 4; Line 3 on Figure 5; Line 4 on Figure 6; and Line 5 on Figure 7. The Interpretation Map (Plate III) is a combination of the results of this survey, with the findings of other geological and geophysical studies.

The Interpretation Map (Plate III) and the interpreted pseudo-sections can be summarized as follows:

1. The study area is dissected by numerous faults or discontinuities that appear to trend in a predominantly northerly direction. Two major faults, one of which appears to be a southern extension of the Dome Fault, are seen intersecting Line 1 (Figure 3) and Line 2 (Figure 4). Other major faults or discontinuities are shown on Line 5 (Figure 7); two of these coincide with faults mapped by Ward and Sill (1976).

2. Three highly conductive subsurface zones have been located within the study area (Areas I, II, and III, Plate III) Area I occurs at the intersection of Line 1 and Line 5. Area II occurs near the northeast end of Line 5, and Area III is located at the northern end of Line 4.

3. A near surface resistive block intersects approximately 500 m of Line 3 and appears to intersect a portion of Line 1. The resistive block has the electrical expression of a subsurface dike (see Plate II), but the true nature of the block will be unknown until further work is accomplished in the area.

DISCUSSION

The results of the dipole-dipole electrical survey are presented as apparent resistivity pseudo-sections in Figures 3 through 7. Significant resistivity contrasts are denoted as faults or discontinuities on the pseudo-sections, with the approximated surface extension of the faults or discontinuities also denoted.

Line 1 and Line 5 (Figures 3 and 7 respectively), the longest dipole-dipole lines, along with Line 2 (Figure 4) indicate that the apparent resistivity values decrease to the southwest, and that the lowest apparent resistivity values (<15 ohm-meters) lie beneath the intersection of the two lines (Area I, Plate III). The highly conductive strata of this area can be caused by a variety of factors, including salinity variation, temperature variation, clay content, and water saturated sediments. Because of the close proximity of Area I to the Escalante Valley, all these factors have a probable influence on the resistivity of the area, with clay and water-saturated sediments considered to have the greatest effect. The effect of temperature on the apparent resistivity values of this area cannot be determined from the pseudo-sections. The highly conductive zone of Area I lies to the west of a major fault or discontinuity. The effects of this fault or discontinuity are most clearly seen from Line 1 (Figure 3) and Line 2 (Figure 4); further, the Interpretation Map (Plate III) indicates that it could be a southern extension of the Dome Fault. In any case, this fault or discontinuity appears to divide the highly conductive rock of the west from the highly resistive rock to the east in this area. Undoubtedly, the highly resistive strata is that of the Mineral Range, and is either granitic or metamorphic in character.

Area II (Plate III) is a near-surface conductive zone. It is located near a well drilled in opal deposits in the NW 1/4 NE 1/4 of Sec. 16. This well hit steam at a depth of 275 feet (see section on Geological and Geophysical Setting), at a temperature of 270°F. It is possible, therefore, that the conductive zone of Area II is an electrical expression of geothermal resource, although it is nearly one mile from the steam well. The hot water of this area could be migrating westward along unmapped east-west fractures.

Area III is a conductive zone which is found deeper in the subsurface. Since the northern limit of this area was not defined, it is possible that this zone is a southern extension of Area II. If this is the case, the entire area surrounding Area II and Area III may be considered as a possible geothermal area warranting further evaluation.

CONCLUSIONS AND RECOMMENDATIONS

Three areas of highly conductive strata have been found within the Roosevelt Hot Springs Prospect (Areas I-III, Plate III). Areas II and III appear to be the most attractive targets for further geophysical work because of a well-defined electrical expression and close proximity to a steam-producing well (NW 1/4 NE 1/4 of Sec. 16, see Plate II). Shallow thermal gradient holes in the immediate area of these two anomalies will give a more clear, and inexpensive, interpretation of the geothermal potential of the area.

Area I appears to expand toward the Escalante Valley and is probably due to water-saturated valley fill. However, the Dome Fault appears, from previous drill hole surveys, to be a channelway for migrating geothermal fluids. With this in mind, one or two shallow thermal gradient holes drilled along the Dome Fault in this area (see Plate III) may prove extremely profitable. The client would have to assess the economic feasibility of such a survey.

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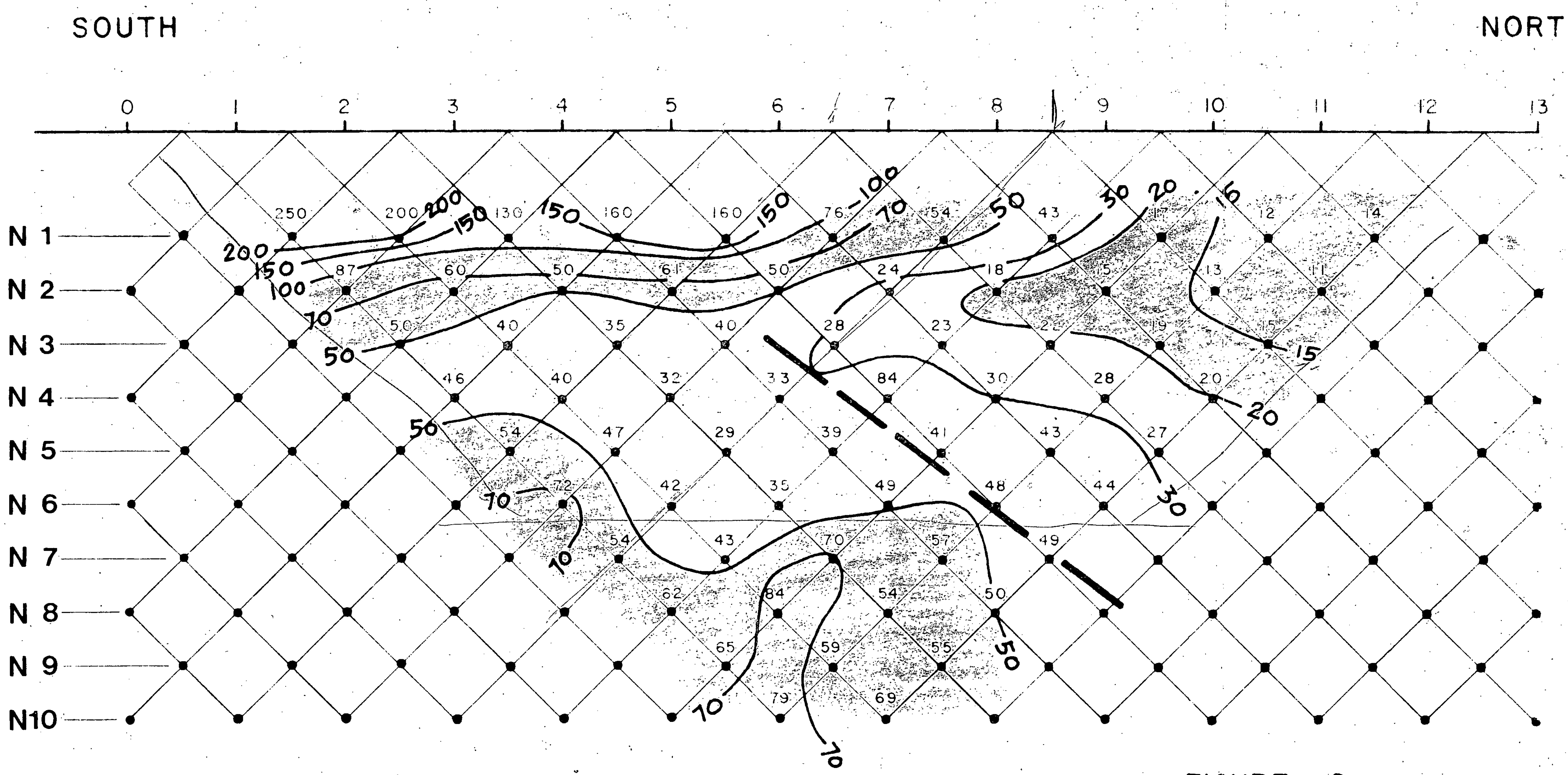


RESISTIVITY IN OHM-METERS

FAULT OR DISCONTINUITY

FIGURE 6

DIPOLE-DIPOLE RESISTIVITY PSEUDO-SECTION

LINE 4

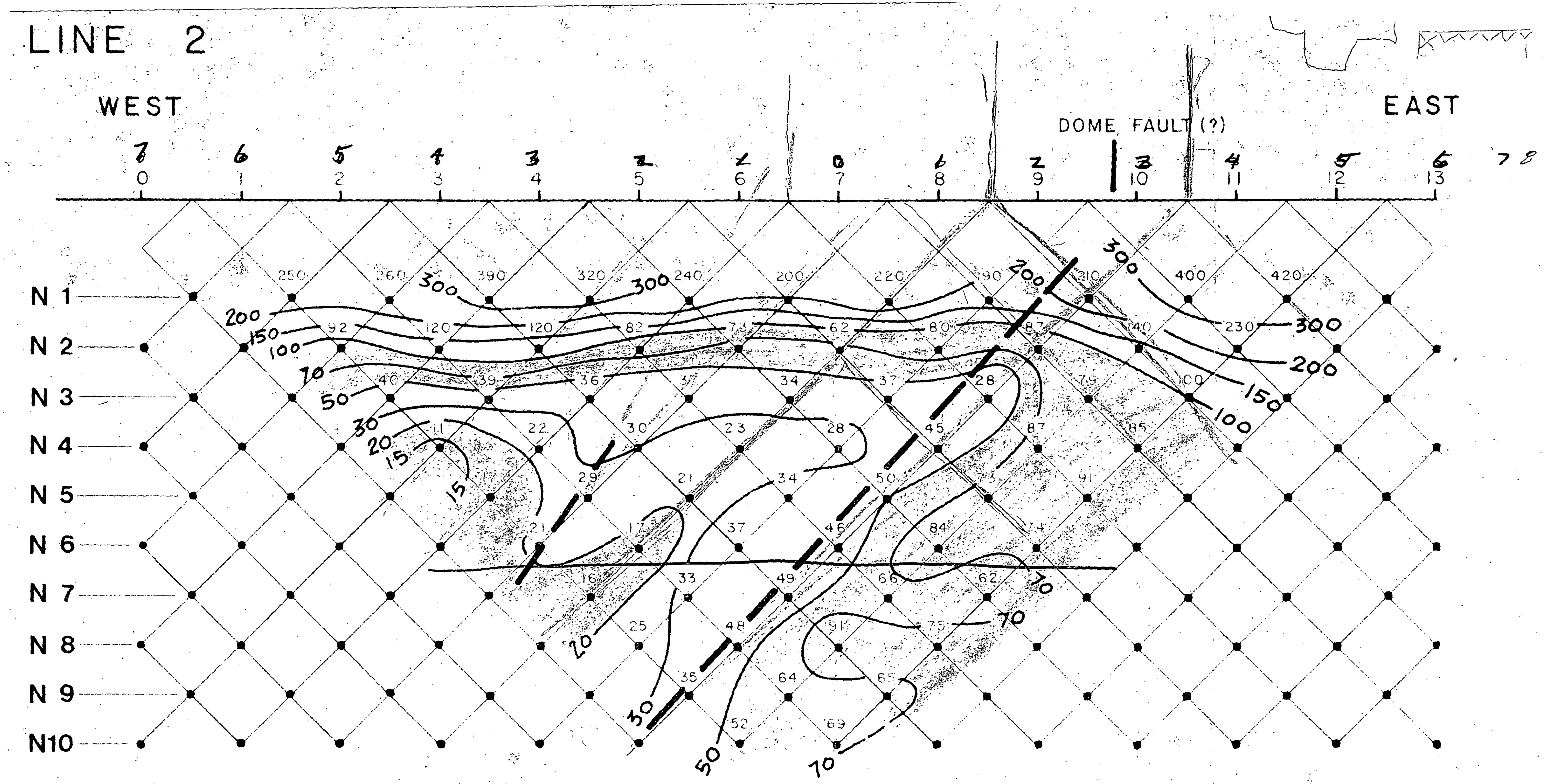


FIGURE 4

DIPOLE - DIPOLE RESISTIVITY

PSEUDO - SECTION

RESISTIVITY IN OHM - METERS

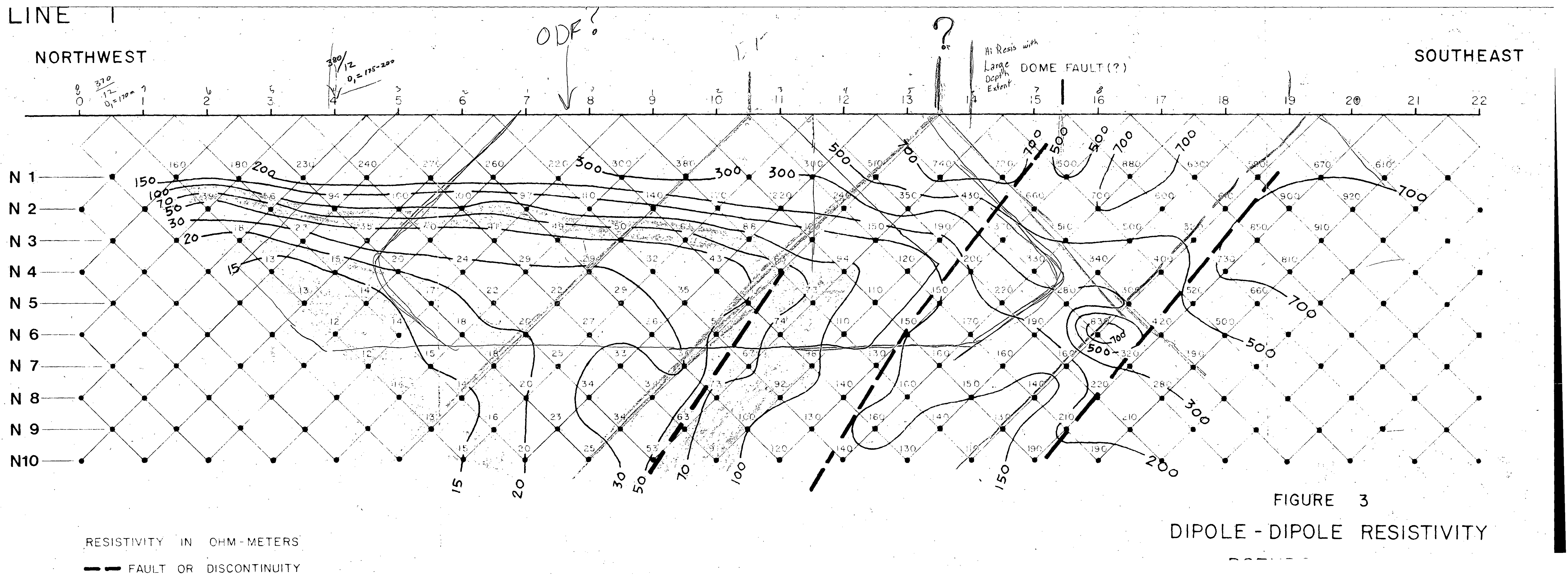
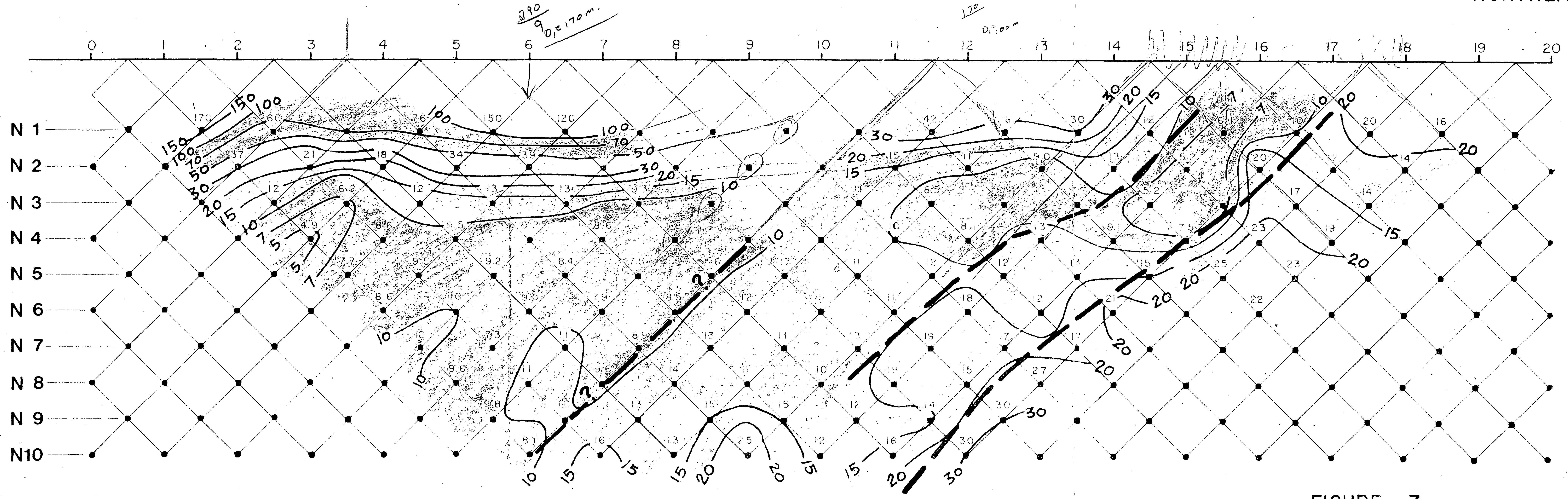


FIGURE 3

DIPOLE - DIPOLE RESISTIVITY

SOUTHWEST

NORTHEAST



RESISTIVITY IN OHM-METERS

FAULT OR DISCONTINUITY

FIGURE 7

DIPOLE-DIPOLE RESISTIVITY

PSEUDO-SECTION

LINE 3

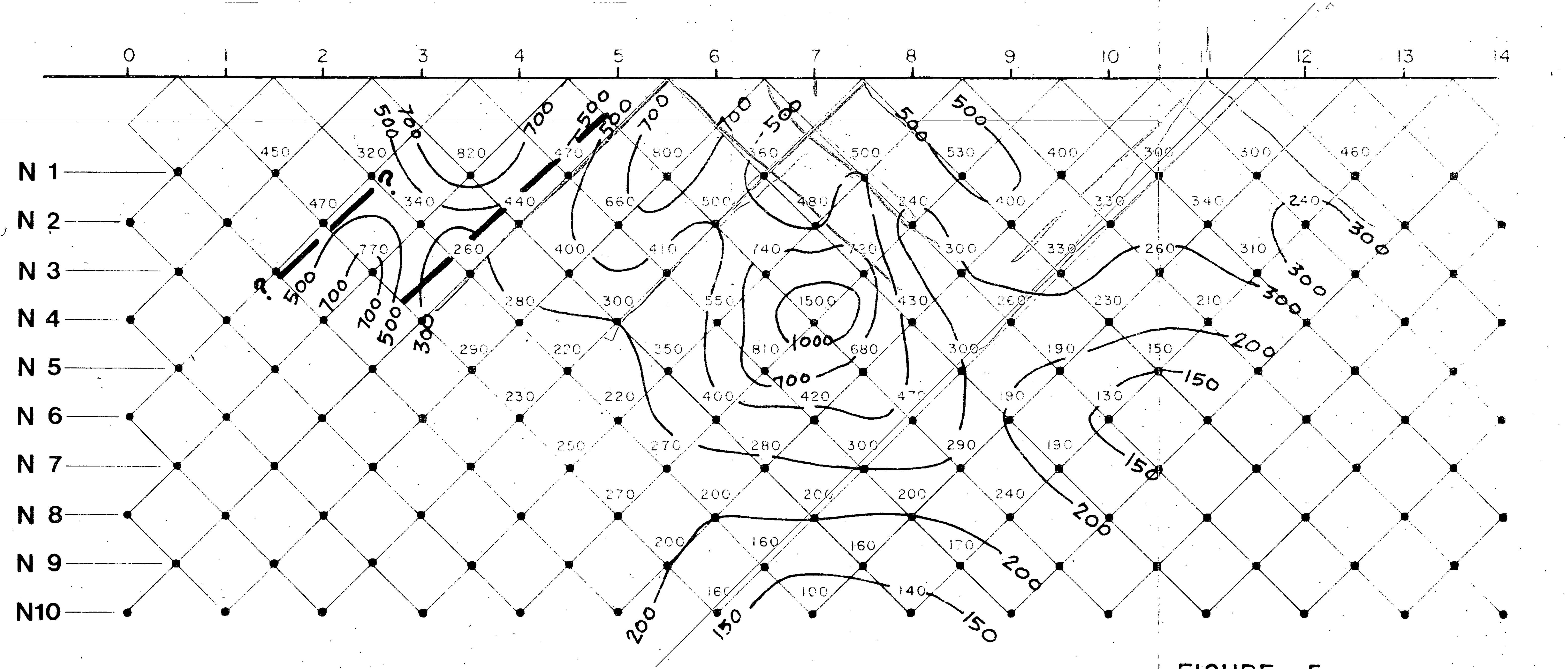


FIGURE 5

DIPOLE - DIPOLE RESISTIVITY

PSEUDO - SECTION

LINE 3

RESISTIVITY IN OHM-METERS

FAULT OR DISCONTINUITY

Abstracts - 1981 AAPG San Francisco
May 31- June 3

chanoid dune deposition. Facies C is characterized by tabular-planar cross-beds, 3 to 4 m thick, interlayered with flat laminated fluvial arenites. It probably formed by migration of solitary transverse dunes across emergent parts of the braidplain. Paleocurrents in all facies are unimodal and parallel to the paleoslope, but commonly show a strong mode perpendicular to the paleoslope. The lack of duricrusts, silcretes, and ephemeral lake deposits suggests a semi-arid to humid paleoclimate.

The eolianites are distinguished from fluvial sediments by: (1) tabular-planar and trough cross-beds bounded by low-angle to horizontal planar surfaces, the cross-beds being composed internally of wedge-shaped intrasets that dip in the direction of the megaset foresets; (2) ripple cross-lamination perpendicular to the megaset foreset dip; and (3) large-scale cross-beds. An eolian origin is substantiated by association with braided river facies, dissimilarity of cross stratification compared to established fluvial facies models, and the tectonic setting.

ROSS, HOWARD P., and WILLIAM E. GLENN, Univ. Utah, Salt Lake City, UT, and CHARLES M. SWIFT, Chevron Resources Co., San Francisco, CA

Reflection Seismic Surveys for Basin and Range Geothermal Areas—An Assessment

Several state-of-the-art reflection seismic surveys have been completed in high-temperature geothermal areas of the northern Basin and Range province. The survey data have been made public through the Department of Energy/Division of Geothermal Energy Industry Coupled and Exploration Technology programs. Data were studied for the Stillwater, Dixie Valley, Beowawe, San Emidio, and Soda Lake resource areas.

Reflection quality, and hence usefulness of the reflection method, can be highly variable in the complex basin and range environment. Certainly survey design and proper processing are required to enhance the quality of the data. The most severe geologic condition appears to be the presence of surface, or near surface, layered volcanic rocks. These result in strong early reflections, substantial ringing and poor energy penetration to depth, as at Beowawe. In areas of thick alluvial cover, or Tertiary gravels and lake bed sedimentation (San Emidio, Soda Lake, Stillwater), data quality is often sufficient to map basin border faults and major displacements on volcanic or bedrock surfaces beneath 2,000 to 4,000 ft (609 to 1,219 m) of cover. Faulting is indicated primarily by the systematic termination of coherent reflections. Diffraction patterns are sometimes recognized but commonly obscured by the complex faulting and lithologic variations. The identification of a given reflector across major structures and accurate time-to-depth conversion are difficult interpretational problems. Excellent data quality at Stillwater and Dixie Valley should contribute to the development of these resources.

ROSS, HOWARD P., DENNIS L. NIELSON, WILLIAM E. GLENN, et al, Univ. Utah, Salt Lake City, UT

Roosevelt Hot Springs, Utah Geothermal Resource—Integrated Case Study

The Roosevelt Hot Springs geothermal resource is located along the western margin of the Mineral Mountains, approximately 19 km northeast of Milford, in southwestern Utah. To date, seven producing wells have been drilled by Phillips Petroleum Co. and Thermal Power Co. Construction will soon begin on the first stage of a 120-megawatt power plant.

Detailed geologic mapping and the study of well logs and drill cuttings indicate that the geothermal reservoir is a fracture-controlled, liquid-dominated system. The host rocks of the reservoir are Precambrian metamorphic rocks and various Tertiary intrusives. The reservoir is mainly localized between the range front and an alluvial covered horst block, along which fluids have migrated to the surface forming an elongate north-trending dome of siliceous sinter. The reservoir is an area of high heat flow (over 1,000 mW/sq mi) and low near-surface electrical resistivity (less than 10 ohm-m). Aeromagnetic, gravity, and reflection seismic data help define the geologic structure within and around the alluvium covered reservoir. Trace element geochemistry shows that arsenic, lithium, and mercury are enriched along fluid pathways of the geothermal system. Mercury concentrations greater than 20 ppb occur only at temperatures less than 225°C and reflect the present thermal configuration of the field.

The system was efficiently explored using detailed geologic mapping in combination with thermal gradient studies and dipole-dipole resistivity.

ROWELL, H. CHANDLER, Exxon Co., U.S.A., Houston, TX

Biostratigraphy of Monterey Formation, Palos Verde Hills, Southern California

Three members comprise the Monterey Formation in the Palos Verdes Hills: the Altamira Shale, The Valmonte Diatomite, and the Malaga Mudstone. Following the diatom zonation of Barron (in press), the middle to upper Altamira Shale ranges from Subzone b of the *Denticulopsis lauta* Zone through Subzone b of the *Denticulopsis hustedtii-D. lauta* Zone (14.5 to 12 m.y.B.P.), the Valmonte Diatomite ranges from Subzone b of the *D. hustedtii-D. lauta* Zone into the lower *Thalassiosira antiqua* Zone (13 to 8 m.y.B.P.), and the Malaga Mudstone ranges from the lower *T. antiqua* Zone into the lower *Thalassiosira oestrupii* Zone (8 to 4 m.y.B.P.), transgressing the Miocene-Pliocene boundary (5 m.y.B.P.). The overlap of up to one million years along the Altamira-Valmonte contact is not surprising since this contact is characterized by a diagenetic change of Opal-A to Opal-CT at most sites.

The age distribution of outcrops reflects northwest-southeast-trending anticlinorium structure of the Palos Verdes Hills, but local sections are discontinuous and deformed due to slumping, folding, and faulting during Pliocene uplift of the hills. This is best seen at Malaga Cove where folds, faults, and slumps are visible along the sea cliffs, and a short hiatus marks the Valmonte-Malaga contact.

The siliceous biostratigraphy of the Palos Verdes Hills correlates to that of the Monterey Formation at Newport Bay. However, the correlation of the siliceous zonation to the benthic foraminiferal stages (assigned by Woodring et al, and Warren) differs for the two areas.

RUBIN, DAVID M., and RALPH E. HUNTER, U.S. Geol. Survey, Menlo Park, CA

Dune Size in Paleodeserts of Colorado Plateau

Where dunes migrate during deposition, they move upward (climb) with respect to the generalized depositional surface. Sediment deposited on each lee slope and not eroded during passage of a following trough is left behind as a cross-stratified

DRAFT COPY

GEOLOGY AND GEOPHYSICS OF THE ROOSEVELT GEOTHERMAL FIELD, UTAH

This paper will be included as a section in the O'Brien-Thermal Power-AMAX application for an ERDA loan guarantee. The figures and plates which accompany this draft are not in every case the final illustrations which will be included in the finished text, but they do contain the information which will be shown on the final illustrations. Plate I, Geologic Map of the Central and Northern Mineral Mountains compiled by Evans was not included with this rough draft copy.

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January 7, 1977

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INTRODUCTION

The Roosevelt Hot Springs thermal area is located on the western flank of the Mineral Mountains in Beaver County, Utah (figure 1). In 1975 Phillips Petroleum Company announced the discovery of a geothermal steam field in the Roosevelt Hot Springs Known Geothermal Resource Area. As currently defined by drilling the Roosevelt steam field lies primarily in T 27 S and R 9 W. The geothermal resources contained within section 16 in this township were leased by the state of Utah to O'Brien Resources and Thermal Power Co. of Utah.

Exploration of the Roosevelt Hot Springs thermal area began in 1968 when Dr. Eugene N. Davie, Mr. Austin B. Smith and Mr. Louis Cooper obtained a geothermal lease from the state of Utah covering section 16, T 27 S and R 9 W. Dr. Davie and his associates drilled a 4 inch rotary hole in the north central part of section 16. The surface location of the hole is immediately to the west of the Dome fault at the eastern edge of the opal mound. The hole was drilled to a depth of 275 feet when steam blew the drilling equipment out of the hole. Uncontrolled steam discharge continued for about 6 weeks until the hole was plugged and abandoned by filling the hole with cement. The steam temperature was 270°F and increased slightly as discharge continued (Petersen, 1975). Today at the surface location of this hole a valve can be observed to leak a small quantity of steam vapor.

In 1972 O'Brien Resources began geothermal exploration in the Roosevelt area and in 1973 O'Brien purchased the geothermal interest held by Mr. Smith and Mr. Cooper in section 16. At the same time O'Brien Resources entered into a joint venture exploration program with Dr. Davie and two companies controlled by him, Davon, Inc. and Thermal Power Co. of Utah. O'Brien Resources and joint venture partners began to acquire geothermal leases and they conducted a geothermal exploration program in the immediate area of the Roosevelt Hot Springs thermal anomaly.

In 1973 and 1974 Thermal Power Co. of Utah, acting as operator for the joint venture partners, retained Dr. David D. Blackwell of Southern Methodist University to conduct temperature gradient and gravity surveys in the immediate vicinity of the Roosevelt Hot Springs thermal area. Temperature gradient holes were drilled in section 16 and the surrounding area. Dr. Blackwell's report has been substantially verified by later workers in the area. He correctly (in the author's opinion) placed the heat source to the east of section 16 below the Mineral Mountains. He also indicated a high probability that geothermal fluids above 400°F would be found below section 16 and the immediately surrounding land (Blackwell, 1974a and 1974b).

Phillips Petroleum Company began their exploration program in the Roosevelt area in 1972. Initially Phillips conducted geologic, geochemical and geophysical studies in the area. In 1975 Phillips Petroleum Co. announced the discovery of the Roosevelt steam field following the drilling of several successful wells (Lenzer, Crosby and Berge, 1977).

Thermal Power Company of Utah made an agreement with Thermal Power Company of California whereby Thermal Power Company of California by drilling a successful geothermal well could earn a half interest in the Thermal Power Company of Utah leases. Thermal Power Company and O'Brien Resources entered into a joint venture exploration program on section 16 in 1976. The Thermal Power-O'Brien Resources program consisted of the drilling of one geothermal well on section 16. The ownership of this well and the exploration expenditures were 2/3 O'Brien Resources and 1/3 Thermal Power Co. Upon the successful completion of this well in 1976 Thermal Power Co. earned a 1/6 interest in section 16 and Thermal Power Co. of Utah subsequently sold the other 1/6 interest to AMAX Exploration, Inc. The ownership in section 16 is now 2/3 O'Brien Resources, 1/6 Thermal Power Co. and 1/6 AMAX Exploration, Inc.

For several years a team of geologists and geophysicists from the University of Utah has ~~have~~ been studying the Roosevelt thermal area and the surrounding region including the nearby Cove Fort thermal area. These studies have been funded by the National Science Foundation and The Energy Research and Development Administration. Nearly all of the publicly available information on the Roosevelt thermal area has resulted from studies by the University of Utah faculty and students. This research has developed valuable scientific information on geothermal systems and it has had practical application in exploration and development of the geothermal resources in the Roosevelt steam field. The geologic and geophysical information contained in this report is based almost entirely on the information contained in the reports of the University of Utah geothermal study group.

GEOLOGY

REGIONAL GEOLOGY

Summary descriptions of the Precambrian, Paleozoic and Mesozoic regional stratigraphy for southwest Utah were published by Woodward (1973) and Welsh (1973). Petersen (1973) and Evans (1977, Plate 1 this report) described the stratigraphy and geology in the Mineral Mountains.

Precambrian metasediments and Paleozoic and Mesozoic sediments in eastern Nevada and southwestern Utah were thrust eastward during the Mesozoic. Imbricate thrusting caused older sediments to be thrust upon younger strata. A period of normal faulting began in the Tertiary which created the northward trending mountain ranges (horsts) and valleys (grabens). The alternating series of horsts and grabens formed the Great Basin physiographic province, which may have formed as a result of east-west extension of the Earth's crust. The Roosevelt thermal area lies at the eastern edge of the Great Basin, where this province meets the less deformed strata of the Shelf, Plains and Plateau provinces. In the Mineral Mountains a period of intrusive activity during the Tertiary resulted in the implacement of granitic magma. The Mineral Mountains granitic pluton is the largest exposed body of granitic rock (approximately 250 km^2) in Utah (Crosby, 1973 and Parry, et. al., in press). A bimodal volcanic assemblage of rhyolite (erupted 0.8 to 0.5 m.y. ago) and basalt are the latest exposed phase of igneous activity in the Mineral Mountains (Nash, 1976). Thermal activity in the Roosevelt thermal area and the nearby Cove Fort thermal area attests to the possible presence of magma at depth in the area. Thermal activity at Roosevelt and Cove Fort and the young volcanism in the Mineral Range and extending eastward to Cove Fort could represent a still evolving silicic volcanic center.

Stratigraphy in the Mineral Mountains

The Mineral Mountains are a horst of Precambrian metasediments and Paleozoic and Mesozoic sediments intruded by Tertiary granite. The granitic pluton is no older than 35 m.y. (Lipman, et. al., 1977). K-Ar dates of 14 and 9 m.y. probably record a later period of igneous activity which resulted in the intrusion of the granite pluton by rhyolite (Armstrong, 1970; Park, 1970, and Parry, et. al., in press). Uplift of the Mineral Mountains horst and accompanying erosion were followed by the volcanism which produced the youngest rhyolitic and basaltic rocks of the Mineral Range. Mountains.

(1) Precambrian Rocks

In the central and west-central portions of the Mineral Mountains dark colored biotite gneiss is exposed which has tentatively been assigned a Precambrian age (Evans, 1977; Petersen, 1973 and 1975). These rocks are assigned a Precambrian age based on their lithologic similarities to similar appearing Precambrian rocks in other parts of Utah. The principle Precambrian metamorphic rock is biotite gneiss with a composition of 50% biotite, 40% quartz, 8% orthoclase, 2% magnetite and minor muscovite and zircon (Earll, 1957). Liese (1957) mapped a zone of Precambrian metamorphic rock that had been partially assimilated by the granitic pluton. Petersen (1975) reported metamorphic rocks in intrusive contact with the granitic rock of the Mineral Mountains.

(2) Cambrian

Three Cambrian sedimentary units outcrop at the northern end of the Mineral Mountains. The oldest sedimentary unit exposed is the Lower Cambrian Prospect Mountain Quartzite. The lithology is a medium grained, massive thick-bedded gray to pink colored quartzite. Evans (1977) reports a maximum thickness of 30 meters.

The Pioche shale overlies the Prospect Mountain quartzite. The Pioche shale is an olive drab to green micaceous shale having a thickness of 8 to 30 meters.

The third Cambrian formation exposed is an undifferentiated limestone unit. The limestone is dark gray in color weathering light to dark gray. Abundant chert lenses occur throughout the unit. The thickness of the exposed portion of the undifferentiated Cambrian limestone is 370 meters.

The Cambrian rocks at the northern end of the Mineral Range are cut by both normal and thrust faults.

(3) Permian

Two Permian formations outcrop at the southern end of the Mineral Mountains. The Coconino Sandstone is a white to brownish-yellow to medium brown orthoquartzite. The Permian Kaibab Limestone is composed of undifferentiated limestone, marble and hornfels.

(4) Triassic

Earll (1957) included 3 rock units in the Moenkopi Group. The lower unit is a red to brown sandstone and sandy shale. The middle unit is a medium bedded to massive gray limestone with thin bedded gray and brown limestone and calcareous shale interbeds. The upper unit is a reddish brown sandy shale with minor limestone beds.

(5) Jurassic

At the southern end of the Mineral Mountains Earll (1957) mapped two Jurassic formations. The Navajo Sandstone is a buff colored to pink, cross-bedded sandstone. The Carmel Formation is a calcareous shale and limestone sequence.

(6) Cretaceous

A conglomerate of subangular to subrounded clasts of limestone and quartzite with minor chert and sandstone in a limestone matrix has been tentatively assigned to the Indianola Formation.

(7) Tertiary Igneous Rocks

The granitic pluton of the Mineral Mountains is composed of three phases: (1) granite, (2) granodiorite and (3) a dark granite phase. The granodiorite and dark granite phases are confined to the northern end of the pluton.

Granite, the predominate phase of the pluton, is a light colored medium to very coarsely crystalline rock that in outcrop forms bold rounded faces. The granite is composed of quartz, microcline, alkali feldspar and plagioclase with minor biotite and magnetite.

The granodiorite is a light colored medium to coarsely crystalline unit that is a cliff former. The mineralogy of the granodiorite consists chiefly of quartz, microcline and plagioclase with minor biotite and hornblende.

The dark granite phase consists of a dark colored, fine to medium crystalline migmatitic textured rock. The principle minerals of the dark granite phase are quartz, microcline, alkali feldspar, plagioclase, biotite and hornblende. Locally all three phases of the granite show alteration of biotite to chlorite and feldspar to sericite. The granitic pluton is intruded by light colored aplite dikes and dark colored mafic dikes.

A series of northwest trending rhyolite porphyry dikes intrude the northern end of the granite. Two outcrops of rhyolite surrounded by alluvium are exposed in Corral Canyon. Evans (1977) reports biotite from this unit yielded a K-Ar date of 8.02 m.y. Nash (1976) states that two young K-Ar dates of 15 and 9 m.y. for the granite are supported by preliminary Rb-Sr isotopic data, which indicates that the granitic rock can be no older than mid-Tertiary (35 m.y.).

(8) Quaternary Volcanic Rocks

The present rugged topography of the Mineral Range existed prior to the onset of Quaternary volcanism in the Mineral Mountains. Nash (1976) divided the rhyolitic rocks in the Mineral Mountains into 3 stratigraphic units. Two nearly non-porphyritic obsidian-rich lava flows comprise the lower most unit. These are overlain by a pyroclastic sequence of ash-fall and ash flow tuffs. The final phase of rhyolitic volcanism was the eruption of porphyritic rhyolite lava domes from 9 separate vents along the crest of the Mineral Mountains.

Two rhyolite and rhyolitic obsidian flows located immediately to the east of the surface thermal anomaly (Sill and Bodell, 1977) are the oldest of the Quaternary silicic volcanic rocks. One flow is exposed along the south side of the Negro Mag Wash and along Bailey Ridge. The other flow is exposed along Wildhorse Canyon, west of Bearskin Mountain. Both flows were originally approximately ~~40~~ 80 meters thick and were deposited /80 along pre-existing valleys which drained the western side of the Mineral Mountains. The basal few meters of the flows are black obsidian which grades upward into flow banded obsidian and light gray or brown finely crystallized rhyolite which is in turn overlain by gray structureless devitrified rhyolite 10 to 30 meters thick forming the interior portion of the flow. The upper parts of the flows are flow banded obsidian interlayered with devitrified rock which passes upward into dense obsidian or grades into a frothy tan perlitic pumice breccia. Uncontorted planar flow banding is typical of both flows and indicates a low viscosity at the time of extrusion. Both flows probably were vented from the upslope side of the outcrop areas or in the case of the Wildhorse Canyon flow the vents may be located beneath younger lavas to the east. The oldest age date obtained from the Quaternary Mineral Mountain volcanic rock is a K-Ar date of 0.77 $\pm$ .08 m.y. from the Bailey Ridge flow which also has a reverse paleomagnetic polarity ~~pole position~~. The Wildhorse Canyon flow was not dated radiometrically, but also shows reversed polarity and the geologic relationships indicate it probably is of approximately the same age as the Bailey Ridge flow (Nash, 1976).

Pyroclastic rocks in the vicinity of North Twin Flat Mountain and South Twin Flat Mountain are the lowest exposed volcanic rocks. The pyroclastic rocks are poorly exposed, weakly consolidated, light colored ash-fall and ash-flow tuffs. At South Twin Flat Mountain the tuff is at least 60 meters thick and possibly as much as 180 meters thick. Air-fall pumice and ash at least 10 meters thick and possibly thicker are the lower most pyroclastic rocks. These are usually overlain by ash-flow deposits at least 50 meters thick and possibly thicker. Geologic relationships indicate the pyroclastic rocks are probably younger than the rhyolite of Bailey Ridge and Wildhorse Canyon. The pyroclastic deposits have normal magnetic polarities and an obsidian clast from an ash-flow tuff in Ranch Canyon yielded a K-Ar age of 0.68 $\pm$ .48 m.y. (Nash, 1976).

Extending for 15 kilometers along the crest of the Mineral Mountains is a group of at least 9 separate lava domes and small porphyritic rhyolite flows. The domes are 0.3 to 1 kilometer in diameter across their bases and they are up to 250 meters high. The more deeply dissected domes of North and South Twin Flat Mountains allow for the internal structure of these domes to be observed. A black 5 to 10 meters thick basal vitrophyric zone grades upward into gray devitrified rhyolite containing lithophysae and gas cavities with lithophysae fillings. The outer surface of the dome was a frothy perlite layer. The interior portion of the domes exhibit a crude flow banding which is subhorizontal in the basal portion, becoming steeper in the upper parts of the dome. The domes exhibit steeply dipping flow banding and ramp structures, which were not present in the older rhyolite flows of Bailey Ridge and Wildhorse Canyon. The dome forming rhyolites contain up to 10% phenocrysts in contrast to less than 1% phenocrysts for the Bailey Ridge and Wildhorse Canyon flows. The phenocrysts consist of feldspar, quartz, biotite, Fe-Ti oxides, sphene, zircon, apatite, and allanite. K-Ar ages on obsidian are: Bearskin Mountain, 0.58 -.78 $\pm$ .07, South Twin Flat Mountain, 0.49 $\pm$ .07, and the un-named northernmost dome 0.53 $\pm$ .06 m.y. Magnetic-polarity determinations for several domes yielded normal magnetic polarities. (Nash, 1976 and Parry, et. al., in press).

Geologic relationships and K-Ar age dates for the rhyolites of the Mineral Range indicate that these rocks formed between 0.8 and 0.5 m.y. ago (Nash, 1976). To the north and east of the Mineral Range, Quaternary basalt flows and cinders are approximately the same age or younger than the rhyolite of the Mineral Mountains (Condie and Barsky, 1972 and Hoover, 1974).

Structure

Crosby (1973) places the Mineral Mountains in the disturbed belt, a transition zone between the undeformed strata of the Shelf, Plains and Plateau provinces and the severly deformed rocks of the fold and thrust belt in the Great Basin. He defines the boundary between the disturbed belt and the main fold and thrust belt as the trace of the easternmost thrust which places Cambrian or Eocambrian rocks over younger strata. Figure 2, a tectonic map of southwestern Utah, shows the relationship of the Mineral Mountains to the disturbed belt. Within the disturbed belt a major zone of transition occurs at the north end of the Mineral Mountains. North of the Mineral Mountains the traces of major normal faults and overthrusts are subparallel while south of the Mineral Mountains the normal fault trends diverge from the trace of the major thrusts by about 35° (Crosby, 1973). A thick pile of Tertiary silicious volcanic rock with associated granite, quartz monzonite, and granodiorite stocks extends in a line trending west-southwest from the Marysvale volcanic center and across southern Nevada. Economic mineral deposits occur associated with the plutonic rocks located within this transverse igneous belt. Several calderas and probable calderas also occur along the transverse igneous belt.

The Mineral Mountains are located at the extreme eastern edge of the Basin and Range physiographic province. A belt of seismic activity, known as the Intermountain Seismic Belt extends from Flathead Lake in Montana southward through Yellowstone Park and along the Wasatch front to the Beaver-Cedar City, Utah area. In the vicinity of Cedar City the zone of seismic activity bends abruptly and extends in a west-southwest direction across southern Nevada (Smith and Sbar, 1974 and Crosby, 1973). The west-southwest trending belt of seismic activity corresponds with a belt of Cenozoic volcanic activity and aeromagnetic anomalies which extends across southwestern Utah and southern Nevada (Stewart, Moore and Zietz, 1977). The Mineral Mountains are located at the east end of the west-southwest trending alignment of seismic activity, transverse igneous belt, Cenozoic volcanism and aeromagnetic anomalies where this trend intersects the north-south trend of the Intermountain Seismic Belt.

North of the Mineral Mountains the leading edge of thrust faulting has been displaced eastward relative to the leading edge of thrust faulting in the Mineral Mountains and to the south. Crosby (1973) termed this the Black Rock offset. In addition to offsetting the eastward edge of the thrust faulting, the Black Rock offset is located at the northern edge of the transverse igneous belt and the southern end of the Intermountain Seismic Belt where the bend in seismic activity occurs. The Black Rock offset also marks the divergence of strike between the major thrusts and normal faults, discussed above. Crosby (1973) suggests that the Black Rock offset is a zone of right lateral strike-slip faulting.

Fault plane solutions were prepared by Olson and Smith (1976) in the Cove Fort area. Their data tends to support the suggestion by Smith and Sbar (1974) that the Basin and Range subplate is moving in a westward direction relative to the more stable eastern portion of the North American plate. Olson and Smith also state that fault plane solutions indicate strike-slip faulting to be present in the southern Utah and southern Nevada area. Gravity data (Crebs and Cook, 1976) aeromagnetic data (Ward, 1976) and ground magnetics (Lenzer, Crosby and Berge, 1976) provide additional evidence for right lateral offset along eastward trending faults in the Mineral Mountains.

GEOLOGY, GEOCHEMISTRY AND GEOPHYSICS OF THE ROOSEVELT THERMAL AREA

Geology

The Roosevelt Thermal Area is located on the western flank of the Mineral Mountains primarily in Township 27 South and Range 9 West and the southern portion of Township 26 South and Range 9 West. The Roosevelt Thermal Area occurs in the central portion of the Mineral Mountains where no Paleozoic or Mesozoic sediments are present. The lithology is biotite gneiss, schist and phyllite, tentatively assigned a Precambrian age, intruded by granite no older than 35 m.y. (Lipman, et. al., 1977). Armstrong (1970) and Park (1970) obtained K-Ar dates of 14 m.y.

and 9 m.y. which probably record reheating of the granitic pluton during subsequent igneous activity. During the middle to late Tertiary the Mineral Mountain horst was uplifted along northerly trending normal faults believed to be concealed beneath the alluvium on the eastern and western flanks of the Mineral Mountains. Erosion during the late Tertiary and Pleistocene produced a rugged relief similar to the present topography.

Volcanic activity has occured fairly regularly during the last 20 million years in the Mineral Mountains and surrounding region. About 20 million years ago Mid-Tertiary volcanic activity occurred just to the south of the Mineral Mountains, and calc-alkalic lavas are exposed on the southern flank of the Mineral Mountains and in the Black Mountains farther south. Near the Thermo KGRA there are exposures of rhyolite dated 9.7 m.y. In the Mineral Mountains at Corral Canyon ~~9.7~~ 8.0 m.y. old rhyolite overlies granitic alluvium. The exposures located south of the Roosevelt thermal area in Corral Canyon indicate that erosion of the Mineral Mountains had occured prior to about 8 m.y. ago. On the southern end of the Mineral Mountains there are basalt outcrops dated at 7.6 m.y. The younger rhyolitic volcanism of the Mineral Range began 0.8 m.y. ago when the rhyolite of Bailey Ridge and Wildhorse Canyon erupted near the present eastern side of the Roosevelt thermal area. Later activity from 0.7 to 0.5 m.y. ago produced ash-fall and ash-flow tuffs which in turn are overlain by at least 9 rhyolite domes extending along the crest of the Mineral Range for approximately 15 kilometers (Nash, 1976, and Parry, et al., in press). On the eastern flank of the Mineral Mountains and extending for 70 miles north of the Mineral Mountains there occur a series of basalt flows with age dates of 6.18 m.y. to as recent as 6,800 years B.P. (Condie and Barsky, 1972; Hoover, 1974, and Parry, et. al., in press).

(1) Thermal Manifestations

The thermal manifestations in the Roosevelt and Cove Fort thermal areas are probably related to the same igneous activity responsible for the young rhyolitic volcanism in the Mineral Mountains and possibly the basaltic volcanism to the east of the Mineral Mountains. In the Roosevelt area the thermal manifestations include: recently active hot springs, opal and opaline sinter deposits, high thermal gradients, and geothermal wells. Most of the thermal phenomena in the Roosevelt area occur along a north-northeast trending fault named the Dome fault (Petersen, 1975). Petersen (1975) and Crebs and Cook (1976) indicate the movement on the Dome fault is upthrown on the west side.

In historic times the hot spring activity in the Roosevelt Thermal Area has been confined to the vicinity of the former Roosevelt Hot Springs Resort at the north end of the Dome fualt. Lee (1908) described the hot springs as having a flow rate of 10 gpm with a temperature on the pipe leading from the spring of 88°C (190°F). Lee also stated that the spring was actively depositing silica at the time of his visit. Mundorff (1970) reported that U.S. Geological Survey personnel reported a discharge of 1 gpm and a temperature of 85°C (185°F) in November, 1950, and a temperature of 55°C (131°F) in September, 1957. By May, 1966, the spring was dry and appeared not to have discharged for several years. Bryant and Parry (1977) report one small seep 500 meters (1640 feet) northwest of the early hot springs with a temperature 25°C (77°F). Table 2, compiled from several sources, contains chemical analyses of surface waters and samples from geothermal wells collected in the Roosevelt Thermal Area. The waters are of the sodium chloride type with an increase of magnesium and calcium in the surface waters over those seeping from the geothermal wells. Silica shows a slight decrease in the surface seeps over analyses from the geothermal wells, possibly indicating shallow deposition of silica by water from the natural seeps. The opal deposits and quartz crystals lining fractures in the reservoir (Lenzer, Crosby and Berge, 1976 and Plate III) indicate silica deposition by the Roosevelt thermal waters.

Deposits of opal sinter and opal cemented alluvium occur along the Dome fault and along inferred northwest trending faults at the north end of the Dome fault (Bryant and Parry, 1977). The opal deposits are the result of hot spring activity located along these faults and Brown (1977a) made a gross estimate that the length of depositional activity on the opal dome was confined to the last 350,000 years. Opal deposition probably required 5,000 to 10,000 years per 1 meter of thickness (Brown, 1977a). Using obsidian hydration rind data Brown (1977b) estimated surface hot spring activity at 0.22, 0.257 and 0.33 m.y. for the measured rims. The nature of the opal deposits are seen in a small opal mine on the opal mound in section 16. The opal occurs as bands a few millimeters to a few centimeters in thickness. Each band is brightly colored and consists of red, brown, green, orange, white, grey, pink and transparent bands. Associated with the laminated opal and siliceous sinter of the opal mound there occurs opal cemented alluvium.

Bryant and Parry (1977) describe a diamond core hole located at the crest of an opal cemented alluvium ridge. The upper 33.5 meters (110 feet) of this hole is altered granitic alluvium. The altered alluvium consists of quartz pebbles, altered feldspar, clay, hematite, green mica and patches and veinlets of opal. The alteration, which will be discussed later in this paper, consisted primarily of alunite, opal, and kaolinite. Parry, Benson, and Miller (1976) and Bryant and Parry (1977) present analytical data for the opal cemented alluvium in the Roosevelt KGRA.

Thermal gradient measurements (Blackwell, 1974a, Sill and Bodell, 1977 and Lenzer, Crosby and Berge, 1976) in the Roosevelt

Hot Springs KGRA have consistently shown the same pattern of high thermal gradients located in a north-south trending elliptical shaped area extending from the southern edge of section 16, township 27 south, range 9 west to the north for approximately 19 kilometers (12 miles) (figure 3). The configuration of the surface thermal anomaly is apparently due to the circulation of high temperature hot water along faults which trend north-south, northwest, east-west. The system of faults containing high temperature geothermal water and associated hydrothermal alteration produces, in additional to the thermal anomaly, an anomalous zone of low resistivity (Ward and Sill, 1976a and 1976b [figure 4] and Lenzer, Crosby and Berge, 1976).

The most spectacular thermal manifestation in the Roosevelt Hot Springs thermal area are the recently completed geothermal wells. By the end of 1977, 7 wells have been completed with a maximum production capability in excess of 4.5 x 10^5 kg (9.9x10^5lbs) of vapor and liquid per hour at a temperature greater than 260°C (500°F) (Parry, et al., in press). Lenzer, Crosby and Berge (1976) report the Roosevelt geothermal reservoir to be a fracture system with the depth to the top of the reservoir being less than 900 meters (2953 feet) over a significant portion of the anomaly. The fracture reservoir has high effective permeability yielding up to 113,000 kg/hr flashed steam. The reservoir temperature is in excess of 250°C, with pressures near hydrostatic and with less than 8000 ppm total dissolved solids in the reservoir fluid. At the time that this paper was being written Phillips Petroleum Company was conducting a field wide long term test on its wells. Thermal Power Company had just abandoned a dry hole in section 36, township 26 south, range 9 west.

(2) Hydrothermal Alteration

A chemical model for a hypothetical geothermal system was prepared by Dedolph and Parry (1976). The model was compared to the Roosevelt Hot Springs system which occurs in a reservoir rock of Tertiary granite intruded into Precambrian gneiss. The predominate minerals in the unaltered reservoir rocks are quartz and potassium feldspar. The alteration zones observed in the Roosevelt Hot Springs system proceeding outward from a fracture are (1) an alunite-quartz zone (2) an argillic zone containing potassium, mica-montmorillonite sub zone (3) a propylitic zone (Parry, Benson, and Miller, 1976 and Bryant and Parry, 1977). The alteration zones observed at Roosevelt Hot Springs are similar to the zones developed in the model of Dedolph and Parry (1976). This similarity allows the zoning to be explained as a function of temperature and pH.

In the Roosevelt thermal area surficial alteration consists of bleached or iron stained, acid-leached alluvium. The surficial alteration is usually closely associated with the opal and opal cemented alluvium discussed above. Surficial alteration varies in intensity from detritus with glassy feldspar to intensely altered rock consisting of opal alunite and traces of native sulfur in a rock containing only relic detrital textures (Parry, Benson and Miller, 1976).

Studies of the subsurface hydrothermal alteration were obtained from diamond drill cores in the Roosevelt thermal area (Parry, Benson and Miller, 1976 and Bryant and Parry, 1977). Figure 5 (figure 14, Parry, Benson and Miller, 1976) and figure 6 (figure 5, Bryant and Parry, 1977) graphically show the major alteration mineralogy in the two diamond drill holes #1A and #76-1 drilled by the University of Utah. Figure 7 is a comparison of the logs for diamond drill holes #1A and #76-1 adapted from data in Parry, Benson and Miller (1976) and Bryant and Parry (1977). The water table for both holes is located approximately at the bedrock-alluvium contact. Above the water table the alteration mineralogy consists primarily of opal + kaolinite + vermiculite with minor sulfur and realgar. Slightly above and below the water table the alteration mineralogy consists of montmorillonite +

pyrite + K-mica + chlorite with minor kaolinite in DDH #1A and illite in DDH #76-1. Bryant and Parry (1977) divided the alteration in DDH #76-1 into 3 alteration zones. A near surface advanced argillic zone passing at depth into an argillic zone down to the water table. Below the water table in the quartz monzonite (granitic rock) there is a propylitic zone. This propylitic alteration was recognized throughout the granitic portion of geothermal well 72-16 and may be the characteristic alteration assemblage throughout the Roosevelt reservoir although the author has no information concerning other wells drilled in the Roosevelt field. Parry, et~~.~~ al. (in press) suggests that the hypothesis developed for the Steamboat Springs, Nevada, thermal system is also applicable for the development of surface and near-surface alteration in the Roosevelt Hot Springs system. Thermal fluids in equilibrium with K-feldspar muscovite and quartz, containing H_2S and sulfate convectively rise along major fractures. H_2S oxidizes at the surface to sulfate decreasing the pH and causing the formation of alunite. Cooler temperatures in the ascending waters caused opal (SiO_2) to precipitate. Low pH, high sulfate surface water flows away from the fracture and percolates downward. As alunite forms, hydrogen ion and sulfate ion are consumed and kaolinite begins to precipitate.

(3) Faulting

The three fault sets described in the Roosevelt thermal area are a north-south set, an east-west set, and a northwest-southeast set (Ward and Sill, 1976a). Figure 8 (Ward et. al., in press and Ward and Sill, 1976a) is a fracture map of the Roosevelt thermal area. The methods used to define these fractures were geology, photogeology, resistivity and magnetics. The granitic rocks of the Mineral Range, which are the reservoir rocks for the Roosevelt steam field, have essentially no intergranular porosity or permeability. The fractures in the granite provide the permeabile passageways along which the geothermal fluids circulate. The interconnected network of fractures is the Roosevelt geothermal reservoir.

faulting cont.

The Dome fault, also known as the Opal Mound fault, is the most prominent fault in the Roosevelt Thermal Area. Mounds of opal and opal cemented alluvium have formed as a result of hot spring activity along the Dome fault. The mound-like deposits of opal mark the northerly strike of the Dome fault. Petersen (1975) described the displacement along the Dome fault as being up on the west side. Several of the gravity profiles of Crebs and Cook (1976) crossed the Dome fault. Line 2200 N crossed the Dome fault which is interpreted to be the east side of a small horst with a displacement of 50 meters (figure 9). Line 4000 N was also modeled to show a fault on the east side of a horst with 50 meters of displacement (figure 9). Fault #1, which is immediately to the east of and parallel to the Dome fault, is believed to bound a continuous zone of fracturing between the Dome fault and Fault #1. Figure 4, a contour map of apparent resistivity, shows a pronounced resistivity low lying along the zone of intense fracturing along the Dome fault, Fault #1, and the intervening ground. This highly permeable area marks the shallowest portion of the Roosevelt reservoir.

At the northern end of the Dome fault the hot spring deposits abruptly change strike and have a northwest trending alignment. This alignment of hot spring deposits is interpreted to occur along a northwest trending fault. The deflection in the resistivity contours (figure 4 and Plate II) also indicates the presence of northwest trending fractures. The northwest trending faults, which are determined by photo linears, appear to lie along a single northwest trending zone which crosses the Mineral Mountains immediately to the north of the point of intersection of the Dome fault and the Negro Mag fault. This suggests that the northwest trending fault set may lie along a northwest trending lineament.

The east-west fault set forms a series of topographic lows crossing the Mineral Mountains. The significance of the east-west fracture set upon the Roosevelt reservoir is at the present time unknown. Ward and Sill (1976a) suggest that the east-west fractures carry fresh water from the Mineral Range westward to mix with the hot saline waters of the convective hydrothermal system. Because their resistivity survey lines were oriented east-west, they were not located in a favorable fashion so as to detect east-west faulting. It was pointed out above that north-south and northwest-southeast trending faults are part of the fracture system along which the convective hydrothermal fluids of the Roosevelt reservoir circulate. The east-west trending faults probably contribute to the fracture reservoir where these faults intersect northwest trending and/or north-south trending faults. It remains to be demonstrated whether or not the east-west fault set is part of the fracture reservoir system away from such points of intersection. The east-west fault set may include both strike slip and normal faults. Crosby (1973), Crebs and Cook (1976), Brumbaugh and Cook (1977), and Lenzer, Crosby and Berge (1976) offer geologic and geophysical evidence which suggests east-west right lateral offset in the vicinity of the Mineral Mountains. Crebs and Cook (1976) show a north-south "baseline profile" through the center of the Roosevelt Thermal Area. This gravity baseline profile (figure 10) crosses two prominent east-west faults. The northernmost fault is known as the Negro Mag fault, sometimes referred to as the Hot Springs fault. Approximately 2.8 kilometers (1 3/4 miles) south of the Negro Mag fault there is a second parallel east-west fault. On the baseline gravity profile the southern unnamed east-west fault is upthrown on the south side and this fault defines the southern end of a graben structure. The Negro Mag fault, modeled as having 65 meters (213 feet) of displacement down on the south side, is the northern end of this graben structure. The baseline profile shows other east-west faults which bound horst and graben structures. Figure

11 (Crebs and Cook, 1976) illustrates faulting interpreted from gravity profiles in the Roosevelt Thermal Area. The map shows the Roosevelt Thermal Area to consist of a series of horst and graben structures bounded by north-south faults and east-west faults.

(4) Heat Source

The heat source for the Roosevelt steam field is unknown. The young rhyolitic volcanic rocks exposed immediately to the east of the Roosevelt Thermal Area suggest a possible common origin for the volcanism and thermal phenomena.

A gravity low centered in the immediate vicinity of the rhyolitic volcanism (Crebs and Cook, 1976) suggests the possibility that magma may still be present at depth beneath the central part of the Mineral Range. The magma chamber could be at a relatively shallow depth and provide the heat necessary for the existence of the Roosevelt geothermal reservoir. Lenzer, Crosby and Berge (1976) **suggest** that magma additions to the magma chamber feeding the young rhyolitic volcanic centers are the heat source beneath the Roosevelt Hot Springs.

Geochemistry

(1) Water Geochemistry

The geothermal waters of the Roosevelt geothermal reservoir are relatively dilute sodium chloride brines. Sampling of all existing springs, seeps and wells in and around the Roosevelt Hot Springs area by Phillips Petroleum Company (Lenzer, Crosby and Berge, 1976) identified several chemically distinct types of water (figure 12). The water types identified chemically are (a) sodium chloride, (b) sodium sulfate, (c) sodium bicarbonate, (d) calcium chloride, (e) calcium sulfate, and (f) calcium bicarbonate. The Roosevelt geothermal reservoir is leaking sodium chloride type water from the reservoir down into the valley in a northwesterly direction toward Black Rock.

Table 2 lists selected water analyses from the Roosevelt Hot Springs KGRA. The analyses are from seeps, Roosevelt Hot Spring and formation water from boreholes. The characteristic Roosevelt geothermal waters are sodium chloride brines containing between 6,000 to 8,000 mg/l total dissolved solids. The waters are characterized by high ammonia, boron and chloride. Sulfate concentrations vary between 45 and 200 mg/l. The principle chemical differences between surface springs and seeps and deep reservoir fluid consist of calcium and magnesium being in higher concentration and silica being in lower concentration in surface waters relative to deeper reservoir waters. Parry, et. al.(in press) report the surface waters to have a higher sulfate and lower sodium and potassium concentration than do the deeper reservoir waters. The analyses in Table 2 do not support these conclusions. They attribute the concentration differences to reflect a progressive leaching of magnesium and calcite by ascending thermal fluids, oxidation of H_2S or a mixture of oxidized, sulfate rich waters, and flashing and cooling with subsequent precipitation of opal.

A comparison of the chemical geothermometers in the Roosevelt Thermal Area showed a fairly close agreement with the actual reservoir temperatures. The Na-K-Ca geothermometer for surface waters ranged from 239 to 295°C and for deep reservoir waters from 262 to 294°C. The Na-K geothermometer for surface waters ranged from 248 to 307°C and for deep reservoir water from 290 to 294°C. The available SiO_2 analyses are limited to a single analysis for surface waters of 170°C and several analyses for deep reservoir waters ranging from 246 to 283°C. These temperatures are in reasonable agreement with the measured reservoir temperatures of 250 to 260°C. Parry, et/ al. (in press) attribute the lower surface water

SiO_2 temperatures to be the result of massive opal deposition at the surface. Decreases in temperature and potassium ion activity and increase in sulfate ion concentration between deep and shallow waters are consistent with the observed alteration assemblage zoning sequence. Oxygen isotope values suggest a relatively high thermal water to host rock mass ratio, which suggests relatively high permeabilities within the Roosevelt Hot Springs geothermal system.

The Roosevelt geothermal reservoir is classified as a sodium chloride hot-water dominated system (White, 1970 and White, Muffler and Truesdell, 1971).

(2) Heavy Metal Distributions

Parry, Benson, and Miller (1976) conducted geochemical surveys for heavy metal distributions in the Roosevelt KGRA. Mercury showed anomalous concentrations in the immediate vicinity of fractures whose presence had been determined by other geologic and geophysical methods. Arsenic and lead did not yield anomalous concentrations. A gamma ray spectrometer was used to determine U, Th, K and total gamma ray counting rate for various rock types. A traverse was made across the opal dome and showed a depletion of U, Th, and K near the center of the opal dome with values rising to normal counts as the traverse proceeded outward from the opal dome into the alluvium.

Geophysics

(1) Gravity

Figure 13 is the terrain corrected Bouguer gravity anomaly map for the central Mineral Mountains (Crebs and Cook, 1976 and Ward, et al., in press). The overall gravity pattern consists of elongate northward trending gravity highs centered to the

Gravity

west of the crest of the Mineral Mountains. A large gravity low occupies the Milford Valley graben to the west of the Mineral Mountains. Steep gravity gradients along the western flank of the Mineral Mountains reflect major displacement along northward trending Basin and Range faults. At location "A" in the vicinity of Ranch Canyon south of the Roosevelt Thermal Area a gravity saddle separates two gravity highs which have been offset approximately two kilometers in a right lateral sense. An elongate gravity spur extending southward from the northern gravity high at location "B" corresponds to the Roosevelt thermal anomaly as defined by temperature gradient contours (figure 3). The gravity contours are highly disturbed where they are intersected by the Negro Mag fault. The gravity data of Crebs and Cook (1976) and Ward et al (in press) showing a gravity high spur extending across the Roosevelt thermal anomaly is in disagreement with the reconnaissance gravity survey of Phillips Petroleum Company, (Lenzer, Crosby and Berge, 1976) which indicate an irregular low over the same area.

The gravity data suggests three locations where east-west trending faults cross the Mineral Mountains. At location "A" and location "C" the gravity data indicated approximately two kilometers of right lateral offset of the gravity contours. At the location of the Negro Mag fault the gravity contours are disturbed indicating faulting and possible right lateral displacement in this area also.

At map localities "D" and "E" two pronounced gravity lows correspond with a series of rhyolitic volcanic domes. These gravity lows may possibly indicate low density intrusive rocks or possible magma at shallow depths beneath the volcanic domes. highly unlikely

A series of gravity profiles across the Roosevelt thermal anomaly (figures 9 and 10) indicate that the bedrock beneath the alluvium has been disrupted into a series of small horsts and grabens. Figure 11 is a map showing the possible fault patterns with indicated directions of fault movement as determined from the gravity profiles. In the Roosevelt Thermal Area a northward trending horst is separated from the Mineral Mountains to the east by a series of northward trending normal faults. The east side of the horst block consists of the Dome fault in a series of closely spaced parallel faults each of which steps down to the east. The western side of the horst is marked by a similar series of northerly trending faults each of which steps downward to the west. The east-west trending Negro Mag fault cuts across the series of northerly trending normal faults and the displacement on the Negro Mag fault is down on the south.

(2) Magnetics

The total aeromagnetic intensity residual anomaly map (figure 14) of the Roosevelt Thermal Area shows a broad northward trending magnetic high, of about 250 gammas of average total relief, extending throughout the central part of the map and corresponding with the granitic pluton of the Mineral Mountains (Ward, et. al., in press). The small magnetic highs appear to represent areas of increased magnetic susceptibility of the granite and Precambrian gneiss. The constriction in the magnetic high, which occurs in the south-central part of the map, appears to represent the rhyolitic domes which intrude the granite in this area. This is also the location of a prominent east-west lineament which may terminate the southern end of the geothermal reservoir. The aeromagnetic map of figure 14 is in good agreement with a ground magnetic survey of Phillips Petroleum Company (figure 8 of Lenzer, Crosby and Berge, 1976). Lenzer, Crosby and Berge state that geothermal wells capable of commercial production, which have been drilled to date, are all situated within magnetic lows.

At the northern end of figure 14 at map locality "F" the magnetic contour lines show a striking east-west alignment with a large gradient. This location corresponds to the northern end of the granitic pluton and is the location of an east-west fault zone. Ward, et al. (in press) mention other magnetic features which correlate well with the geology and/or gravity features. These include: a magnetic low (G) which partly overlies the rhyolitic volcanic domes, an east-west magnetic lineament (H) along the Negro Mag fault with a 2 kilometer right lateral offset of the magnetic highs on opposite sides of the fault and a magnetic low (I) that corresponds almost exactly in aerial extent with the reversed magnetic polarity of the Bailey Ridge rhyolite flow.

A series of ground magnetic profiles (Crebs and Cook, 1976) were run along the same traverse lines as were the gravity profiles. The ground magnetic profiles and the gravity profiles are seen to be in good agreement.

(3) Passive Seismic

The Roosevelt Thermal Area is located where earthquake trends of the north-south trending Intermountain Seismic Belt begins to turn west-southwest across southwestern Utah (Smith and Sbar, 1974). The west-southwest seismic belt crosscuts the north-south trending late Cenozoic horsts and grabens of the eastern Great Basin. The west-southwest seismic zone may reflect a highly fractured crust that would be a favorable zone for magma intrusion into the upper crust (Ward, et al., in press). The distribution of Cenozoic volcanism and Tertiary intrusive rocks across southwestern Utah and southern Nevada indicates this area to have been a zone of intrusive and extrusive activity (Stewart, Moore and Zietz, 1977). Shallow bodies of magma located along the seismic belt would be potential heat sources for geothermal reservoirs.

Initial seismic monitoring by Phillips Petroleum Company (Lenzer, Crosby and Berge, 1976) and the University of Utah

(Olson and Smith, 1976) indicated that the Roosevelt Thermal Area was aeseismic.

The results of the University of Utah earthquake monitoring surveys of 1974 and 1975 are plotted on figure 15. Ward, et al. (in press) reports the earthquakes were grouped into three areas: (1) a north-south trend of moderate activity on the west side of Milford Valley, (2) a trend of minor activity along the west side of the Mineral Mountains, including the Roosevelt Hot Springs Thermal Area, and (3) an area of numerous swarm-like earthquake sequences around the Cove Fort area. The University of Utah survey indicated the Roosevelt Thermal Area to be only slightly seismically active. The ray paths for earthquake activity in the Cove Fort area that were monitored on the west side of the Mineral Range showed a low-velocity effect and shear wave attenuation possibly suggestive of a magma source beneath the southern part of the Mineral Mountains (Ward, et al., in press).

A passive seismic survey by Phillips Petroleum Company (Lenzer, Crosby and Berge, 1976) detected a microearthquake swarm activity in the Roosevelt Thermal Area which commenced in April, 1976 and continued until at least August, 1976. Figure 16 is a ground noise anomaly map of the Roosevelt Thermal Area. The map contoured for integrated power in the frequence window between 9 and 11 hertz showed an anomaly coincident with the extent of the Roosevelt geothermal reservoir as it is presently known.

(4) Electrical Surveys

Dipole-dipole resistivity surveys using 100 m (Plate II) and 300 m (figure 4) dipoles by the University of Utah show a pronounced resistivity low centered over the Roosevelt Thermal Area (Ward and Sill, 1976a and 1976b). The authors believe the resistivity anomaly is due to brine-soaked clays in the top 500 meters of the Roosevelt geothermal system. The clays are the alteration products of feldspars and occur predominately along faults and fractures in the reservoir rocks. Surface conduction in the kaolinite and montmorillonite secondary clay minerals is estimated to be three times as important as conduction in the high temperature brine of the reservoir (Ward, et al., in press). The alteration clays occur predominately along fractures and for this reason most of the fractures in the shallow portion of the reservoir were detected by the 100 meter dipole-dipole survey. These fractures are shown on figure 8 which depicts fractures determined from; (P) aerial photography, (G) observed geology, (R) dipole-dipole resistivity surveys and (M) aeromagnetic surveys (Ward, et al., in press). Geologic and geophysical evidence suggests that the faulting is greatest in a north-south trending zone through the center of the Roosevelt thermal anomaly and the north-south trending elongate resistivity low delineates this zone of intense faulting.

Phil to comment

A reconnaissance MT survey was conducted by Phillips Petroleum in the region aroung the Roosevelt Thermal Area. The four MT survey stations which fell on the Roosevelt thermal anomaly showed a low resistivity interval, the upper surface of which lay between 900 and 1370 meters (3000 to 4000 feet) below the surface. This model was confirmed by subsequent drilling (Lenzer, Crosby and Berge, 1976).

(5) Temperature Gradients

Thermal gradient measurements outline the same anomalous zone as the dipole-dipole resistivity survey. The thermal gradient survey detects anomalous subsurface temperatures which result from the circulation of high temperature geothermal reservoir water along fractures. The resistivity survey detects high temperature geothermal brines along fracture zones and the associated clay minerals, which are produced by the hydrothermal alteration of the wall rocks surrounding the fractures. Figure 3

(Ward et al., 1977)

~~(Sill and Bodell, 1977)~~ is a temperature gradient contour map of shallow (30 m to 60 m interval) thermal gradients. This map agrees with Sill and Bodell's thermal gradient contour map for gradients greater than 100 meters in depth and also with the temperature gradient map in Lenzer, Crosby and Berge (1976). The resistivity maps show lowest resistivity values to occur in a north-south trending zone between the Dome fault and Fault #1 with the low diverging to the northwest north of the Negro Mag fault. The temperature gradient maps show the same pattern of high temperature gradients in a north-south trending zone between the Dome fault and Fault #1 with a northwest bend north of the Negro Mag fault.

Sill and Bodell (1977) report temperature gradient holes west of the Dome fault show a decreasing gradient with depth in the hole. The temperature gradient data is interpreted to reflect high temperature geothermal fluids leaking to the surface along the Dome fault and the faults immediately to the east. The hot fluids move westward and gradually mix with the regional ground water flow in the Milford Valley. The Roosevelt thermal gradient anomaly is therefore asymmetric with the western portion of the anomaly being shallow subsurface leakage from faults located in the central portion of the anomaly. No successful geothermal wells have been drilled west of the Dome fault which tends to confirm the observation that the western portion of the thermal anomaly represents high temperature reservoir fluids leaking westward from the reservoir.

GEOLOGY OF SECTION 16

Geology

Section 16 lies at the south end of the Roosevelt steam field as it is presently defined by drilling. The temperature contour map (figure 3), the resistivity map (figure 4 and Plate II) and the surface distribution of opal deposition and hydrothermal alteration (Plate I) indicate that section 16 lies at the southern end of the shallow portion of the Roosevelt geothermal reservoir. The terrain-corrected Bouguer gravity anomaly map (figure 13) shows an elongate gravity high spur extending southward from the high at point "B". This spur closely corresponds in trend and aerial extent with the thermal anomaly (figure 3) suggesting a relationship between the gravity anomaly and the thermal anomaly which may be either a factor of the geothermal reservoir or both features possibly result from a common geologic controlling factor. The total aeromagnetic intensity residual anomaly map of the Roosevelt Thermal Area (figure 14) shows a magnetic high which closely corresponds with the geothermal reservoir. This anomaly is strikingly constricted at the southern end of the Roosevelt geothermal reservoir near the south edge of section 16. As is the case with the gravity anomaly map the magnetic map tends to indicate either a direct relationship between the magnetic anomaly and the geothermal reservoir or both features may have a common geologic controlling factor. Several lines of evidence therefore suggest that the shallow portion of the Roosevelt geothermal reservoir probably terminates immediately to the south of section 16. The temperature gradient data and the resistivity map indicate that the geothermal reservoir beneath section 16 lies at a very shallow depth. The drilling in section 16 of geothermal well 72-16 and the earlier Davie well have proven the very shallow occurence of high temperature geothermal fluids.

The most spectacular geologic feature in section 16 is the Dome fault (Opal Mound fault). Hot springs which once issued along the Dome fault in section 16 deposited a prominent ridge of banded opal. The relatively youthful age of the opal deposits (Brown, 1977a) proves the existence of active surface hydrothermal activity in section 16 in the recent geologic past. The location of the opal deposits along the trace of the Dome fault indicates that this fault served as a conduit along which the geothermal fluids were transported. A second fault, named Fault #1 (figures 8 and 11) is parallel to the Dome fault. Fault # 1 passes approximately through the northeast corner of section 16 and both the Dome fault and Fault #1 appear to be terminated in the southern part of section 16 by a series of closely parallel east-west trending faults. The intervening block of ground between the Dome fault and Fault #1 is probably intensely fractured and this fracture system is responsible for the high temperature gradients and low resistivity values (figures 3 and 4) encountered from section 16 and northward to the vicinity of the Roosevelt Hot Springs. Section 16 east of the Dome fault is intensely faulted by movement along the Dome fault and Fault #1 with probable additional faulting resulting from east-west trending faults. Excellent fracture reservoir potential apparently occurs in the east half of section 16. The faulting relationships west of the Dome fault in section 16 are somewhat obscure. Figures 8 and 11 indicate that north-south trending faults are present west of the Dome fault and therefore possible fracture reservoir conditions may exist in the western half of section 16. Plate II, a first separation resistivity dipole-dipole array (100m dipoles), shows low resistivity values west of the Dome fault in section 16 to be associated with north-south trending faults (Ward and Sill, 1976). These low resistivity values may be the result of high temperature geothermal fluids moving laterally in the subsurface to the west from east of the Dome fault or the north-south trending faults may be conduits for high temperature geothermal fluids and thus constitute part of the fracture reservoir. The presence or absence of a geothermal reservoir west of the Dome fault in section 16 can only be proven by drilling test wells. Phillips Petroleum Company well OH-1 (a stratigraphic test) drilled to the west of section 16 in section 17 was apparently a dry hole and offers evidence that the Roosevelt geothermal reservoir may not extend to the west of section 16. Phillips Petroleum well 9-1 drilled to the north of section 16 apparently did not encounter highly fractured ground. Phillips Petroleum disposal well 82-33 drilled approximately 3 miles to the north of section 16 and west of the Dome fault was reported to be dry (Ward and Sill, 1976a and Lenzer, Crosby and Berge, 1977). Evidently all holes west of the Dome fault that have been drilled to date are dry.

The subsurface geology beneath section 16 consists of a layer of arkose composed mainly of quartz, feldspar and mica (biotite, chlorite and phlogophite) with included fragments of granite and gneiss. The more indurated portions of the arkose are difficult to distinguish from the granite especially in well cuttings. The author (Plate III) placed the base of the arkose at approximately 430 feet (131 meters) in well 72-16. The mud logger's log shows the base of the arkose at 424 feet (129 meters) based on a change in drilling rate at this point. In well 72-16 the arkose contained hydrothermal alteration and geothermal fluids at a sufficiently high temperature so that the well flowed at 312 feet (95 meters). The arkose is cemented by a dense impervious clay which effectively renders the arkose impermeable.

Immediately below the arkose the bedrock consists of Tertiary granite intruding Precambrian gneiss. The granite is intensely faulted and the fractures contain high temperature geothermal fluids. The geothermal reservoir begins at the granite-arkose contact and in some places extends upward along faults into the arkose. The impermeable arkose acts as a cap over the Roosevelt reservoir beneath section 16. In the geologically recent past the reservoir was able to leak upward through the arkose along the Dome fault and create the opal ridge which extends through the center of section 16. The granite in the reservoir beneath section 16 shows an overall propylitic alteration characterized by biotite to chlorite. Where the granite with propylitic alteration is cut by hydrothermal fluid bearing fault zones the alteration changes to a potassium silicate assemblage. The potassium silicate assemblage consists of K-feldspar, secondary quartz, pyrite, transformation of the biotite and dark green chloritoid of the propylitic zone to a silvery blue green chlorite, increase in magnetite, possible siderite and possible minor sericite. During drilling the change in the alteration assemblage to a potassium silicate mineralogy indicates hydrothermal fluid bearing or former hydrothermal fluid bearing fracture is about to be encountered.

The subsurface geology beneath section 16 can be described as a highly fractured altered granite containing a high temperature and high pressure geothermal reservoir which is contained by a relatively impervious arkose cap. In places faulting with associated hydrothermal fluids and alteration extends into the overlying arkose.

Geothermal Well 72-16

Geothermal well 72-16 was spudded on October 22, 1976 and completed December 31, 1976 at a total depth of 382.2 meters (1254 feet). The well location is 990 feet south and 990 feet west from the northeast corner of section 16, T 27 S, R 9 W. The ground elevation at the well site is 5580 feet above sea level and all depth measurements were taken from the top of the Kelly Bushing which was 21 feet above ground level. The casing in well 72-16 consists of: 20 inch conductor pipe to a depth of 85 feet, 13 3/8 inch to a depth of 580 feet and 9 5/8 inch to a depth of 1098 feet. A cement bond log for the 13 3/8 inch casing was measured to a depth of 522 feet. The cement bond log showed fair bonding to a depth of 270 feet and excellent cement bonding from 270 to 522 feet. A remedial grouting program upon completion of the well was undertaken for the upper portion of the 13 3/8 inch casing. A seep of boiling water which was leaking up around the outside of the 13 3/8 inch casing from a hot water zone in the alluvium was successfully sealed off by this grouting program. A cement bond log for the 9 5/8 inch casing was run from a depth of 20 feet to 999 feet and showed perfect bonding down to the 999 foot interval.

Lithologic sampling of well 72-16 began at a depth of 85 feet. The interval from 85 to 150 feet consists of arkose composed mainly of granitic mineral and rock fragments with rare gneissic schist fragments. The mineral fragments are predominately quartz, K-feldspar and biotite cemented by an iron stained clay. At 120 feet doubly terminated quartz crystals appear which were probably derived from the rhyolitic tuffs to the east of the Roosevelt Thermal Area.

The interval from 150 feet to 210 feet showed the first evidence of weak hydrothermal alteration. The alteration consisted of a sharp increase in chlorite after biotite, euhedral pyrite crystals and weak alteration of K-feldspar to clay. The interval 210 to 360 feet showed evidence of more intense alteration associated with faulting. Above 210 feet the arkose exhibited a weathered appearance which suddenly was no longer present at 210 feet. The hydrothermal alteration effectively gave the cuttings a "fresher" appearance which was originally mistaken for the granite-alluvium contact. This interval consisted of a faulted zone from 270 to 320 feet showing strong evidence of faulting with abundant pyrite, vein quartz, slickensided fragments and chloritic gouge. Above and below the faulted interval the arkose consisted of quartz and K-feldspar fragments with minor biotite, chlorite and pyrite. At 312 feet the well flowed hot water and this interval was cemented off.

From 360 to 430 feet the lithology again became arkose consisting of weathered granitic mineral and rock fragments with rare gneiss fragments.

At approximately 425 to 430 feet occurs the arkose-granite contact. Below 430 feet the hole remained in granite until the bottom at 1254 feet. The granite mineralogy is quartz, K-feldspar, biotite altering to chlorite, magnetite, sphene, and secondary pyrite, specular hematite or rutile, and sericite. The chlorite is very abundant and this would be described as a chloritic granite. Fracture zones show veinlets of secondary quartz.

In the interval from 460 feet until approximately 850 feet a series of closely spaced faults were penetrated. The fracture zones contain chlorite gouge, breccia and slickensided fragments. Quartz and jasperoid veins with pyrite are common. Secondary silicification of the faulted granite occurs at 610 to 640 feet and 670 to 680 feet. In addition to the quartz, K-feldspar and chlorite the mineralogy includes sphene to leucoxene, epidote, drusy quartz, specular hematite or rutile, and possible hornblende. At 514 feet and 628 feet the well flowed discharging boiling water and steam. Throughout the interval from approximately 500 to 800 feet the well was constantly making steam and drilling proceeded with great difficulty. At 580 feet the 13 3/8 inch casing was set as a safety measure.

From about 800 to 850 feet the evidence of faulting and hydrothermal effects gradually decreased. There was no sharp lower limit to the zone of faulting and alteration. With increasing depth the percentage of dark colored minerals increased so the rock gradually no longer resembled a granite but rather a granodiorite. The mineralogy of the granodiorite is quartz, K-feldspar, plagioclase, chlorite after biotite, hornblende, hematite after magnetite, epidote and sphene. A faulted zone occured from 940 feet to 960 feet with a drilling break at 944 to 945 feet showing CO_2 gas increase, trace of pyrite and altered feldspar.

At 1030 feet a sharp lithology change occured which consisted of an increase in the percentage of quartz and feldspar and a decrease in the percentage of mafic minerals. As the percentage of mafic minerals decreased the percent of chlorite relative to other mafic minerals increased. At 1110 feet the lithology again changed to the same granodiorite as above 1030 feet.

The interval from 1130 to 1190 feet is a faulted zone showing evidence of disseminated pyrite, feldspar altered to sericite, hematite, jasperoid and fault gouge.

Below 1190 feet minor faulting continues, but secondary hydrothermal mineralogy is very minor. At 1230 feet the top of the alteration associated with the main reservoir zone in well 72-16 is encountered. The mineralogy shows a sharp increase in the percentage of chlorite and mafic minerals just above 1230 feet and then a sharp decrease immediately below 1230 feet. Below 1230 feet there is an increase in quartz and feldspar, the biotite is altered to books of silvery blue green chlorite, a reddish brown crystalline mineral is possibly siderite, magnetite increases in percentage and becomes more coarsely crystalline, hematite and fault gouge are also present. At 1244 to 1254 feet the main reservoir zone occurs. A one foot drilling break at 1245 feet resulted in the complete loss of drilling returns and is believed to be the fractured reservoir for well 72-16.

Because of the large quantity of steam from the 1245 foot interval it was necessary to install Otis snubbing equipment before the well could be completed. A Halliburton EZSV Retainer-bridge plug was set at 1144 feet and a cement plug was set on top of the retainer-bridge plug. 9 5/8 inch casing was cemented to a depth of 1098 feet. The cement plug and the retainer-bridge plug were drilled out and the hole was deepened from 1245 to 1254 feet. Sample collection below the 1245 foot depth was extremely difficult because of the temperatures and pressures which existed in the hole. Two samples were collected from 1245 to 1250 feet and 1250 to 1253 feet. Both samples consisted mainly of cement, baked drilling mud and metal fragments from the drilled out retainers. The samples were washed with dilute hydrochloric acid and then water. By this procedure any carbonate minerals and most of the clay were lost. The metal fragments and baked drilling mud were separated from the sample and the remaining material was then logged. The sample was 90% quartz and K-feldspar in a ratio of 1:4 estimated by using refractive index oils. The remaining 10% of the sample consisted of euhedral pyrite and magnetite, biotite, chlorite and amphibole. Many of the quartz fragments are pieces of quartz crystals and the interval probably contains vuggy quartz veins.

The granitic rock in well 72-16 shows overall propylitic alteration characterized by chlorite after biotite, minor pyrite, and clay minerals. Potassium silicate alteration is characteristic of the high temperature hydrothermal fluid bearing fracture zones. The alteration mineral assemblage consists of K-feldspar, secondary quartz, pyrite, transformation of the biotite and dark green chloritoid of the propylitic zones to a silvery blue-green chlorite, an increase in magnetite and the possible presence of siderite and minor sericite.

The well was tested on April 4-5, 1977 and the data obtained in this test was reviewed by DeGolyer & McNaughton (1977). Based on the observations obtained during the test, DeGolyer and McNaughton reached the following conclusions:

"(a) At a depth of 1240 feet the reservoir is overpressured at 720 psig. Because of this excessive pressure over the normal hydrostatic gradient and because of the nature of the reservoir fluid, it is estimated that the communicating fluid of comparable temperature exists at least 2275 feet below the total depth of this well. This indicates a significant enough reservoir that the production of 30 million pounds of mass flow during the test would not be expected to reduce the reservoir pressure.

(b) The failure of the build-up test to obtain the shut-in initial pressure after 25 hours is cause for concern. At this time, the build-up to 95.1% of the initial pressure should not be viewed as an effective reservoir pressure loss. In addition to a subsequent build-up test, static pressure and temperature runs will be necessary to verify or deny the indicated results.

(c) The 3% decrease in flow rate should not be viewed as an expression of depletion. In our opinion, the 24 hour flow period was of insufficient duration to establish a stabilized flow rate. The production characteristics of the Utah State 72-16 indicate a large, significant geothermal discovery of undetermined magnitude."

DeGolyer & McNaughton go on to state "the flow tests for April 4-5, 1977, that provided a mass flow rate of 1.3 million pounds per hour, exhibited a calculated flash potential of approximately 18% to 19% vapor at 100 psig. flowing pressure and 394°F flowing temperature. Hence the calculated vapor-state flow rates would range from 234,000 to 247,000 pounds per hour. Using a conservative conversion factor of 20 pounds of vapor per kilowatt of generated electricity, the demonstrated flow rate would provide an instantaneous potential of 12 to 13 megawatts per hour. No estimate can be made at this time to support sustained mass-flow rates on the order of 1.3 million pounds per hour. Prolonged flow testing will be necessary to establish the future production capabilities of this and future wells on your property.

Summary

The Roosevelt geothermal reservoir is a sodium chloride hot-water dominated system. The system is contained in a fracture reservoir in Tertiary granite which intruded Precambrian gneiss. The top of the Roosevelt geothermal reservoir is relatively shallow lying at a depth of from 100 to 900 meters. The reservoir fluid contains 6000 to 8000 ppm total dissolved solids and the reservoir temperature is in the range of 250 to 260°C. The reservoir shows an overall propylitic alteration with a potassium silicate alteration assemblage noted along the geothermal fluid bearing fractures. The distribution of hydrothermal alteration and shallow thermal phenomena are located within a north-south trending oval shaped area which several geophysical methods show to be anomalous.

There are three fault sets observed in the Roosevelt Thermal

Area. A north-south set and a northwest-southeast set are normal faults which can be demonstrated to contain geothermal reservoir fluids. As presently defined by drilling the proven portion of the Roosevelt geothermal field lies along a series of closely spaced north-south trending faults. The third east-west fault set shows evidence of both vertical and right lateral displacement. There is evidence to suggest that the east-west fault set may bound the southern limit of the Roosevelt geothermal reservoir.

The heat source for the Roosevelt geothermal reservoir is unknown. The presence of relatively young rhyolitic volcanic rock immediately to the east of the Roosevelt geothermal reservoir suggests that the magma chamber responsible for this volcanism may be the heat source for the Roosevelt geothermal field. Limited gravity and passive seismic data suggest that magma may still underlie a portion of the area of young rhyolitic volcanism.

Several geophysical methods have been used to explore the Roosevelt geothermal reservoir. Gravity profiles indicate that the Roosevelt thermal anomaly consists of a series of horst and graben structures delineated by the three previously mentioned fault sets. The University of Utah gravity survey showed an overall gravity high across the Roosevelt Thermal anomaly. An independent gravity survey by Phillips Petroleum Company indicates an irregular low over the same area. Aeromagnetic data indicates that the successful geothermal wells have all been drilled within a magnetic low. This low is situated to the east of the Dome fault and extends eastward beneath the area of Recent rhyolitic volcanism.

A passive seismic survey by Phillips Petroleum Company showed an anomaly coincident with the Roosevelt geothermal reservoir. An electrical survey by the University of Utah delineated a resistivity low across the Roosevelt geothermal reservoir which is essentially the same area outlined as a thermal anomaly by temperature gradient measurements.

Evidence, both geologic and geophysical, has defined a shallow geothermal reservoir in the Roosevelt KGRA. The shallow portion of this reservoir appears to be approximately 8 miles long in a north-south direction and 2 miles wide. A larger reservoir may extend farther to the east at depths in excess of 1500 to 2000 meters.

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TABLE I. STRATIGRAPHY MINERAL MOUNTAINS

	Earll [1] 1957	Liese [1] 1957	Hintze [1] 1963	Baetcke [1] 1969 (Star Range)	Petersen [1] 1973	Evans 1977	
Quaternary						alluvium	
						volcanic talus	
						opal & opal cemented alluvium	
						Cove Fort basalt	
						Black Rock basalt	
						rhyolite domes 0.73 ± 0.10 to 0.49 ± 0.07 m.y.	
						ash-flow & ash-fall tuffs 0.68 ± 0.04 m.y.	
						Bailey Ridge & Wildhorse Canyon rhyolite	
Quat. or Tert.						lake limestone	
Tertiary						Corral Canyon rhyolite 8.02 ± 0.3 m.y.	
						rhyolite porphyry	
			T(?)congl.		T(?)congl.	aplite & mafic dikes	
						granite	
						granodiorite	
						dark granite	

STRATIGRAPHY (continued)

	Earll [1] 1957	Liese [1] 1957	Hintze [1] 1963	Baetcke [1] 1969 (Star Range)	Petersen [1] 1973	Evans 1977
Cretaceous	Claron(?)	Indianola(?)				Indianola(?) congl.
Jurassic	Carmel 574 ft. Navajo 1,538 ft.		Carmel Navajo	Navajo(?) ss.	Navajo	
Triassic	Moenkopi Group upper red middle ls. lower red 1,282-1,708 ft.		Moenkopi Group	"Moenkopi Fm."	Moenkopi Fm. lower red unit(?) Timpoweap Mem.	Moenkopi Fm siltstone
Permian	Kaibab 698 ft. Coconino 1,181 ft.		Kaibab Fm.	Kaibab-Plympton undiff. Talisman Qzite Pakoon undiff.	Talisman Qzite Pakoon undiff.	Kaibab ls. Coconono ss.
Pennsylvanian			Talisman Qzite Callville	Callville	Callville	
Mississippian	Topache ls. 917 ft.		Redwall ls.	Monte Cristo	Redwall ls.	
Devonian				Mowitza Fm. Guilmette Fm.		

STRATIGRAPHY (continued)

	Earll [1] 1957	Liese [1] 1957	Hintze [1] 1963	Baetcke [1] 1969 (Star Range)	Petersen 1973	Evans 1977
Cambrian		Cambrian(?) ls. Undiff. € ls. Pioche Sh. Prospect Mt.	Undiff. € ls. Pioche Sh. Prospect Mt.		Undiff.€ ls. Pioche Sh. Prospect Mt.	Pioche Fm. 8-30 meters Prospect Mt. Qzite 30 mtrs. Undiff.€ ls. 370 meters exposed
Precambrian	P€ undiff.	P€ undiff.	P€ undiff.		P€ undiff.	P€(?) undiff.

1. from Petersen, 1973

TABLE 2

SELECTED ROOSEVELT KGRA WATER ANALYSES

	(1)	(2)	(3)	(4)	(5)
Na (ppm)	2080	2500	2400	1800	1840
Ca (ppm)	19	22	113	107	122
K (ppm)	472	488	378	280	274
SiO_2 (ppm)	405	313	76	107	173
Mg (ppm)	3.3	0	17	23.6	25
Cl (ppm)	3810	4240	3800	3200	3210
SO_4 (ppm)	65	73	142	70	120
HCO_3 (ppm)	158	158	536	300	298
S (ppm)					1.03
Al (ppm)					
Fe (ppm)					
Sr (ppm)					
Li (ppm)		0.27		17	
F (ppm)	7.1	7.5	5.2	3.3	
B (ppm)		38	37	29	
Total Dissolved Solids	7040	7800	7506	5948	
Temperature (°C)	85	55	17	28	25
pH		7.9	8.2	6.43	6.5
Na-K-Ca (T°C)	295	285	247	239	241
Na-K (T°C)	307	282	250	248	
SiO_2 (T°C)					170
Hardness as $CaCO_3$					
N.C. Hardness as $CaCO_3$					
Date	11-4-50	9-11-57	5-9-73	8-15-75	6-25-75

(1) Roosevelt Hot Springs, Mundorff (1970) 11-4-50

(2) Roosevelt Hot Springs, Mundorff (1970) 9-11-57

(3) Roosevelt seep, Phillips Petroleum Company 5-9-73

(4) Roosevelt seep, Phillips Petroleum Company 8-15-75

(5) Roosevelt seep, University of Utah 6-25-75

TABLE 2 Continued

	(6)	(7)	(8)	(9)	(10)
Na (ppm)	2437	2210	2150	2072	2400
Ca (ppm)	8.0	83	9	31	9
K (ppm)	448	425	390	403	565
SiO_2 (ppm)	560?	364	640	639	775
Mg (ppm)	0.01		0.6	0.26	19
Cl (ppm)	4090	3800	3650	3532	4800
SO_4 (ppm)	59	122	78	48	200
HCO_3 (ppm)	180			25	
S (ppm)					
Al (ppm)				1.86	
Fe (ppm)				0.016	
Sr (ppm)					
Li (ppm)	20.0				18
F (ppm)	5.0				
B (ppm)	25				45
Total Dissolved Solids	7067				
Temperature (°C)	205		260	92	
pH	6.3		5.9	5.0	
Na-K-Ca (T°C)	273	262	286	274	
Na-K (T°C)	294				
SiO_2 (T°C)			283	283	
Hardness as $CaCO_3$					
N.C. Hardness as $CaCO_3$					
Date	5-25-75		March 1977	1-29-77	

(6) Phillips Petroleum Company, well 3-1, Phillips Petroleum Company 5-25-75

(7) Phillips Petroleum Company, well 9-1, University of Utah

(8) Thermal Power Company, well 14-2, water condensate, U.S. Geological Survey

(9) Thermal Power Company, well 72-16, University of Utah 1-29-77

(10) Phillips Petroleum Company, well 54-3, Chemical and Mineralogical Services analyst

TABLE 2 Continued

	(11)	(12)	(13)	(14)	(15)
Na (ppm)	2000	2000	2000	1800	1350
Ca (ppm)	10.1	12.20	12.2	12.4	8.2
K (ppm)	410	400	400	380	290
SiO_2 (ppm)	560+	521	104+	510	104+
Mg (ppm)	0.24	0.29	0.28	0.29	0.23
Cl (ppm)	3400	3260	3180	2110	2330
SO_4 (ppm)	54	32	34	33	25
HCO_3 (ppm)	200	181	193	181	134
S (ppm)					
Al (ppm)					
Fe (ppm)					
Sr (ppm)		1.20	1.20	1.36	1.06
Li (ppm)	19.0	16.0	15.0	15.0	12.2
F (ppm)	5.0	5.3	5.3	5.2	3.8
B (ppm)	29	27.2	27.2	26.4	183
Total Dissolved Solids	6442	6444	5868	6074	4277
Temperature (°C)	>260	*	*	*	*
pH	6.5	7.53	7.72	7.83	7.91
Na-K-Ca (T°C)	294				
Na-K (T°C)	290				
SiO_2 (T°C)	263+	246	246	246	246
Hardness as $CaCO_3$		297	316	297	220
N.C. Hardness $CaCO_3$		64.6	64.6	71.5	48.4
Date	8-26-75	April 1977	April 1977	April 1977	April 1977

* measured bottom hole temperature 242°C

(11) Phillips Petroleum Company, well 54-3, Phillips Petroleum Company 8-26-75

(12) Thermal Power Company, well 72-16, U.S.G.S., April 1977, sample #1 water

(13) Thermal Power Company, well 72-16, U.S.G.S., April 1977, sample #1 condensate

(14) Thermal Power Company, well 72-16, U.S.G.S., April 1977, sample #2 water

(15) Thermal Power Company, well 72-16, U.S.G.S., April 1977, sample #2 condensate

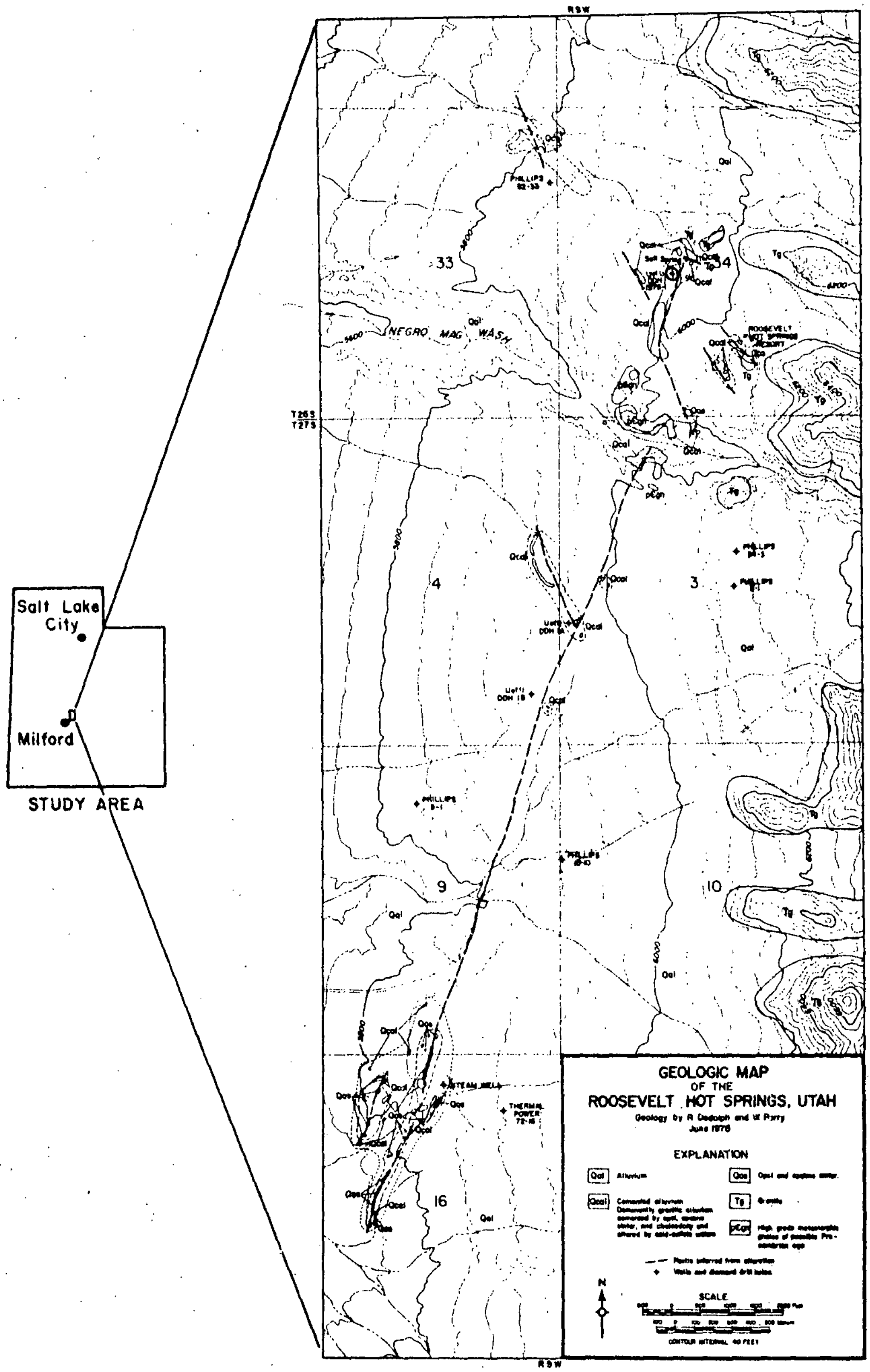


Figure 1. Map of Utah showing location of Roosevelt KGRA and geologic map of the Roosevelt Hot Springs KGRA.

(Bryant and Parry, 1977)

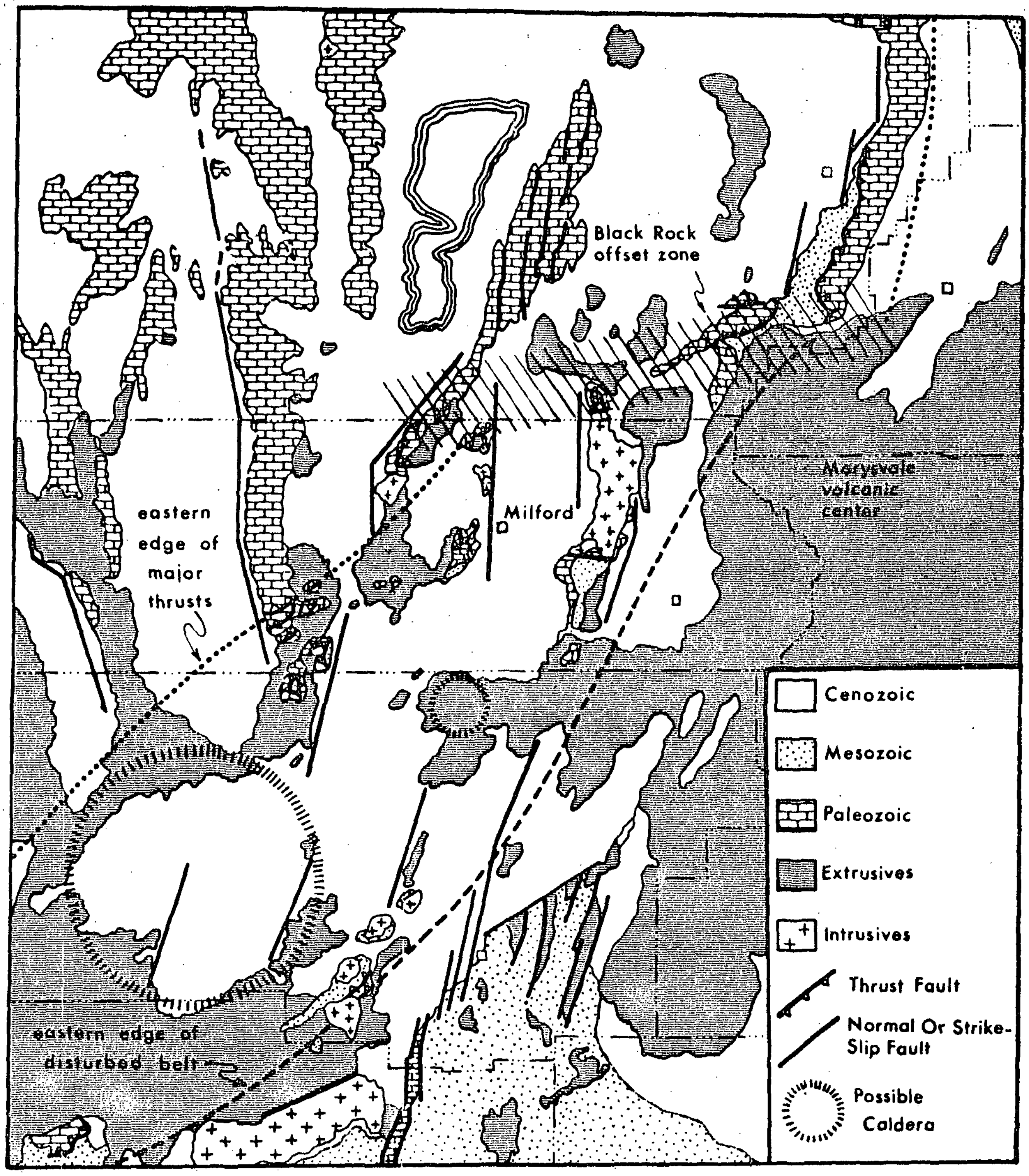


Figure 2 –Tectonic map of southwestern Utah.

(Crosby, 1973)

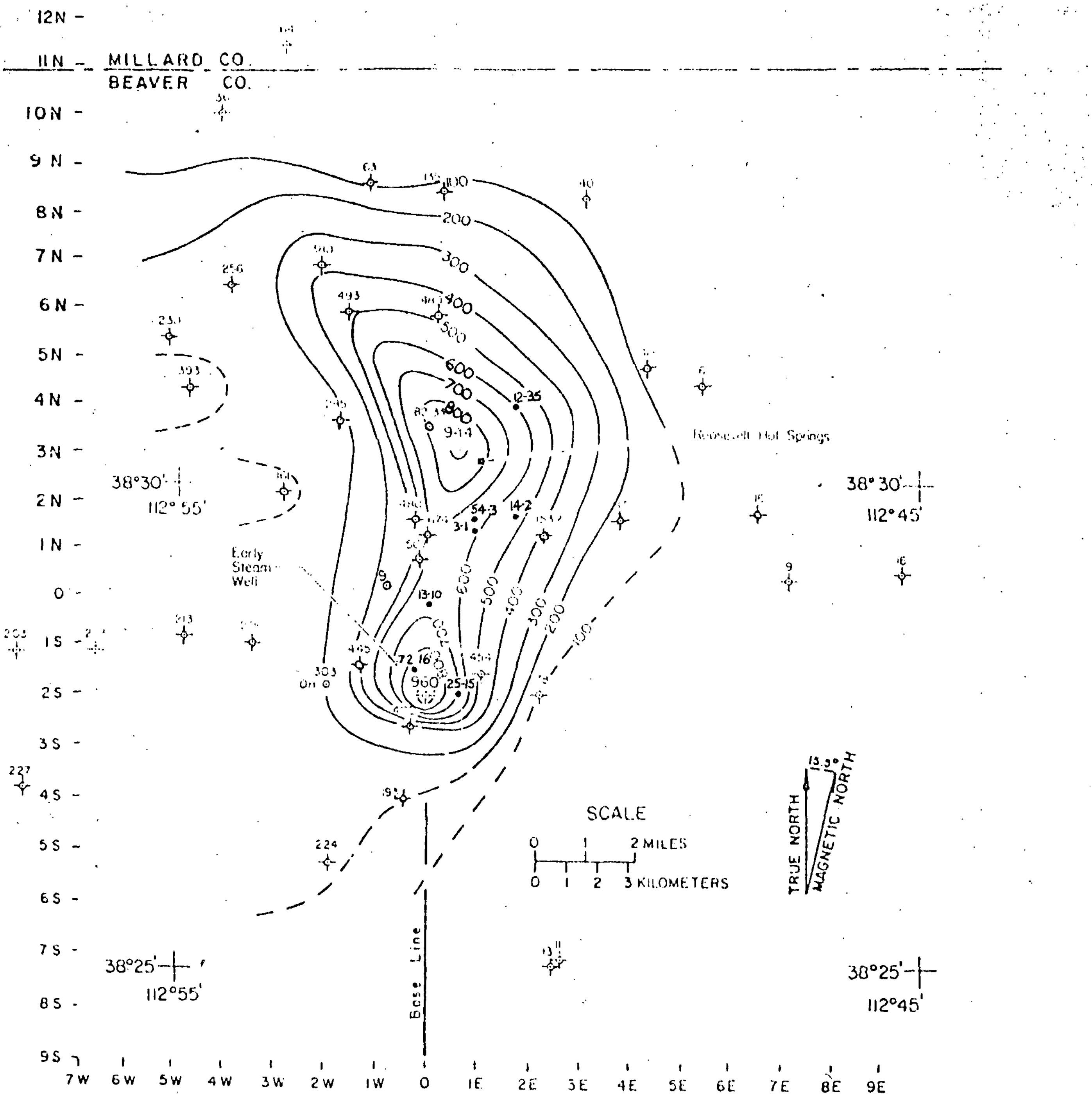
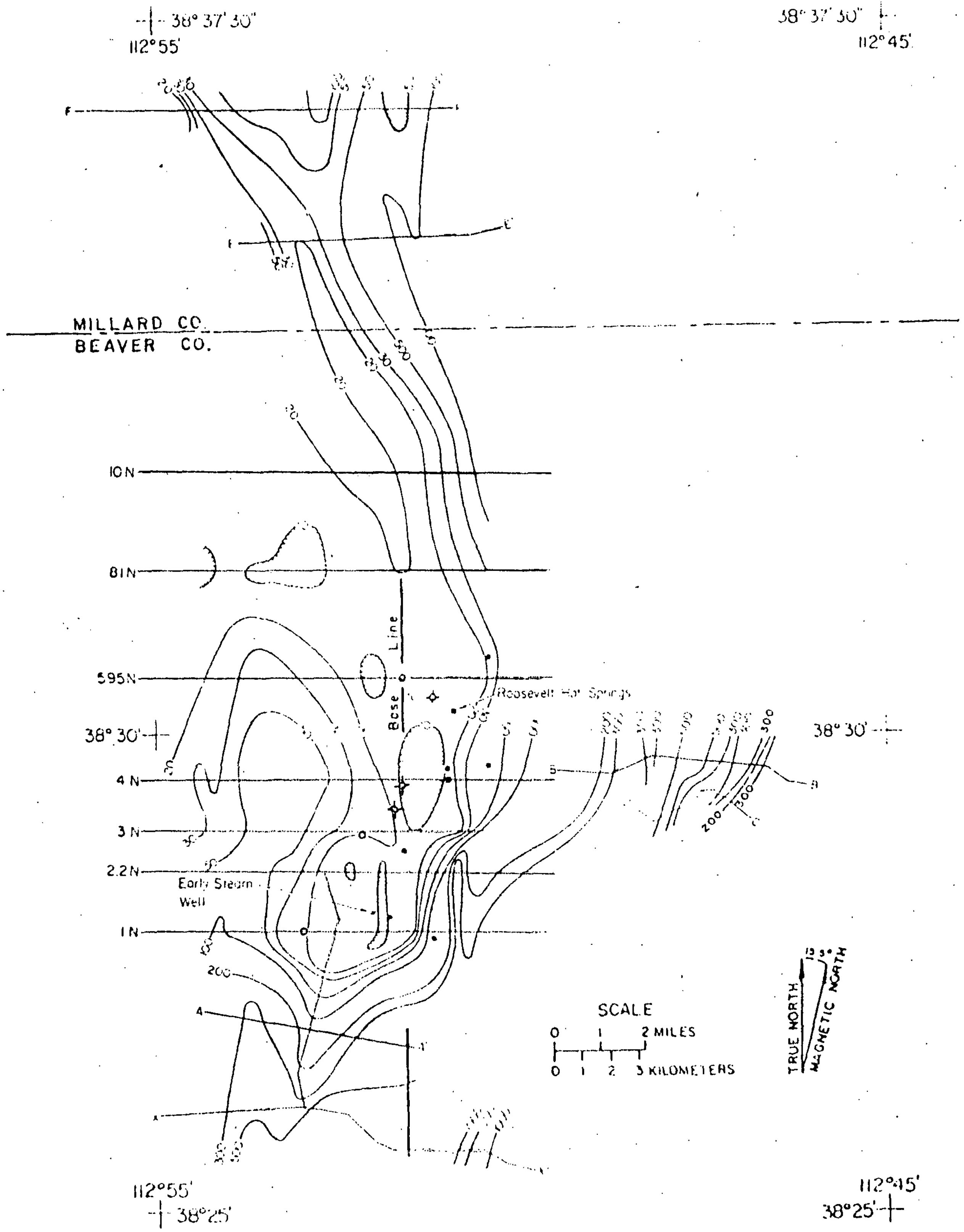


Figure 3. Plan map of contoured thermal gradients in the depth interval from 30 to 60 m. Contour interval 100°C/km. Also shown are the locations of thermal gradient holes (circle with cross), as well as productive (solid dots) and nonproductive (open circle) wells

(Ward, et. al., in press)

Figure 4. Contours of apparent resistivity obtained with dipole-dipole array, first separation. Contours at 10,20,30, 50,100 ohm-m and multiples of ten times these figures. Productive wells shown by solid dots, "dry wells" by open circles, shallow alteration holes by circles with crosses. Traverse lines are shown. (Ward, et. al., in press)



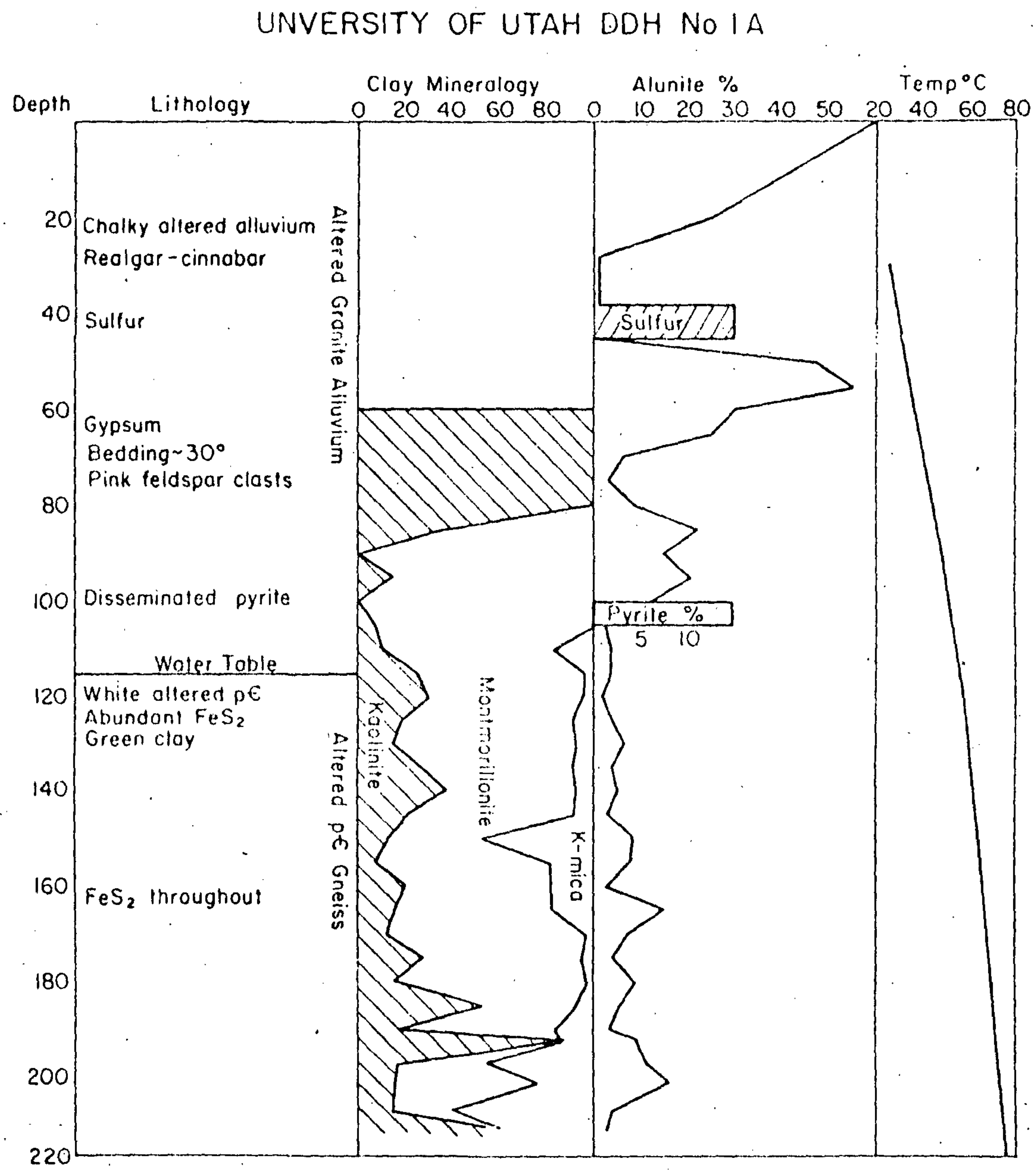


Figure 5 Diamond Drill Hole 1A logs. Clay mineralogy shown is relative percent of total clay.

(Parry, Benson and Miller, 1976)

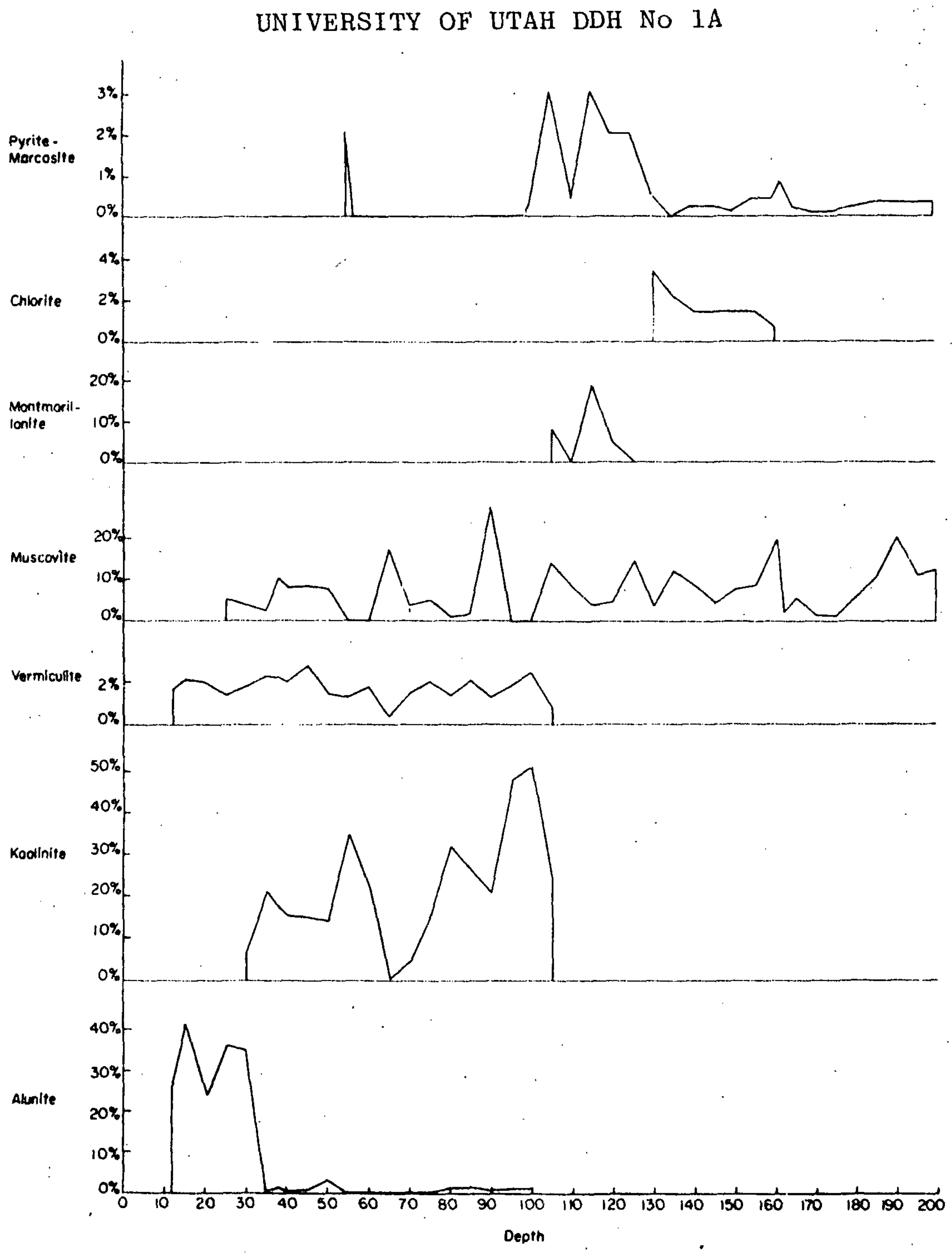


Figure 6 Major Alteration Mineralogy (MODECALC) versus Depth.

(Bryant and Parry, 1977)

LOGS UNIVERSITY OF UTAH DIAMOND DRILL HOLES

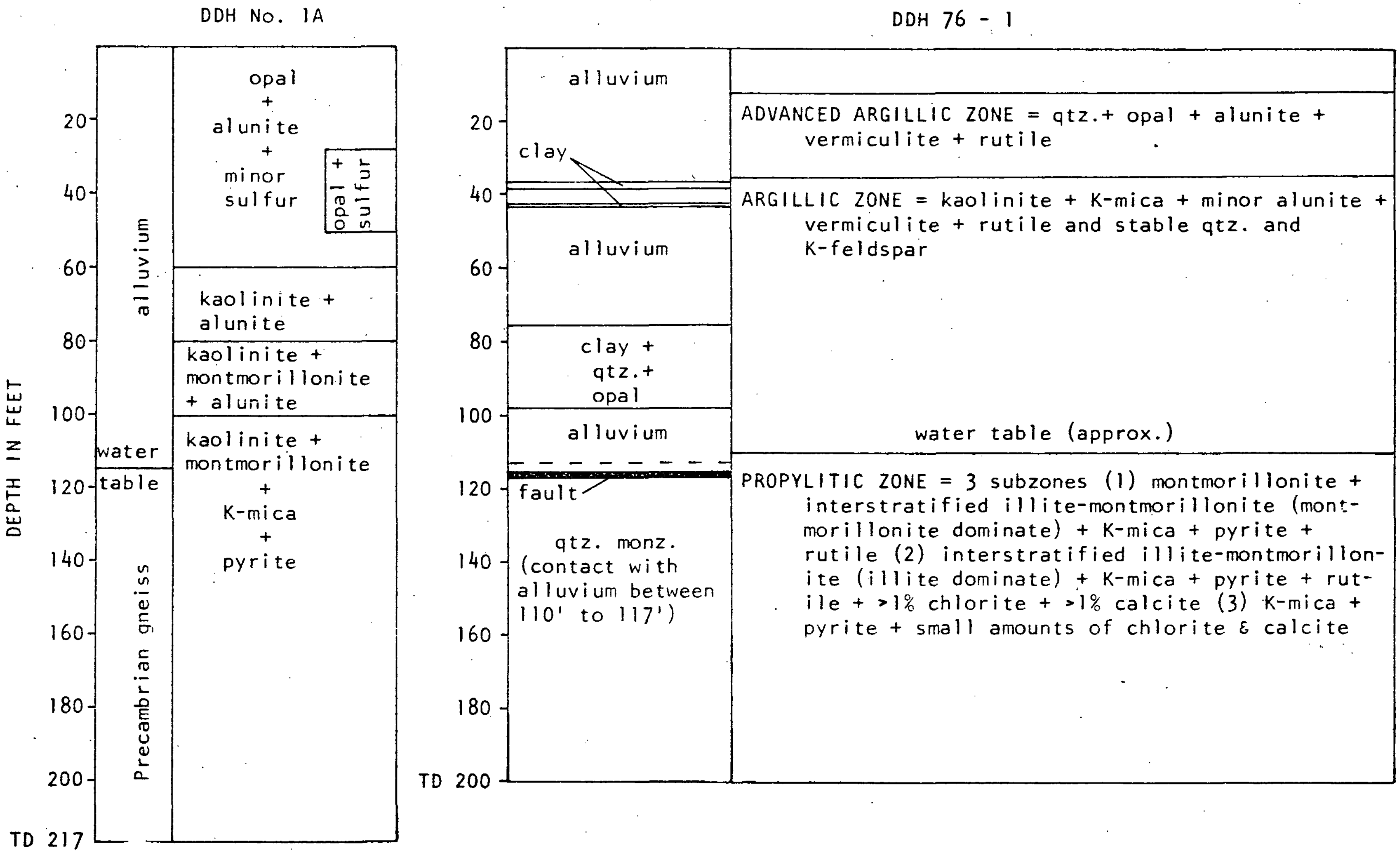
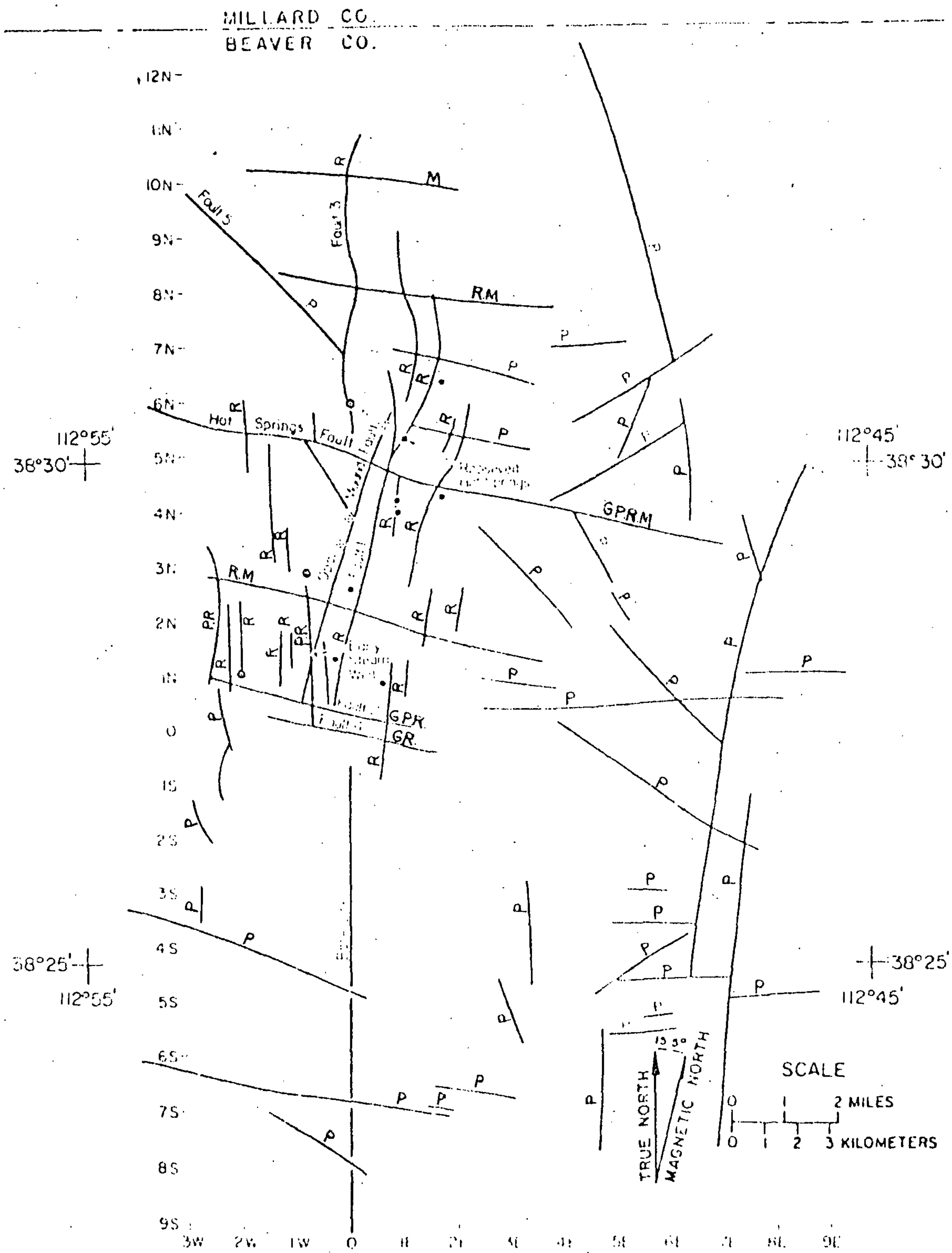


Figure 7 compiled from data in Parry, Benson and Miller, 1976 and Bryant and Parry, 1977

Figure 8. Lineaments, interpreted as fractures and faults, mapped by photos (P), geologic observation (G), resistivity survey (R), and aeromagnetic survey (M). Producing wells shown by solid dots, "dry wells" by open circles, shallow alteration holes by circles with crosses.

(Ward, et al., in press)



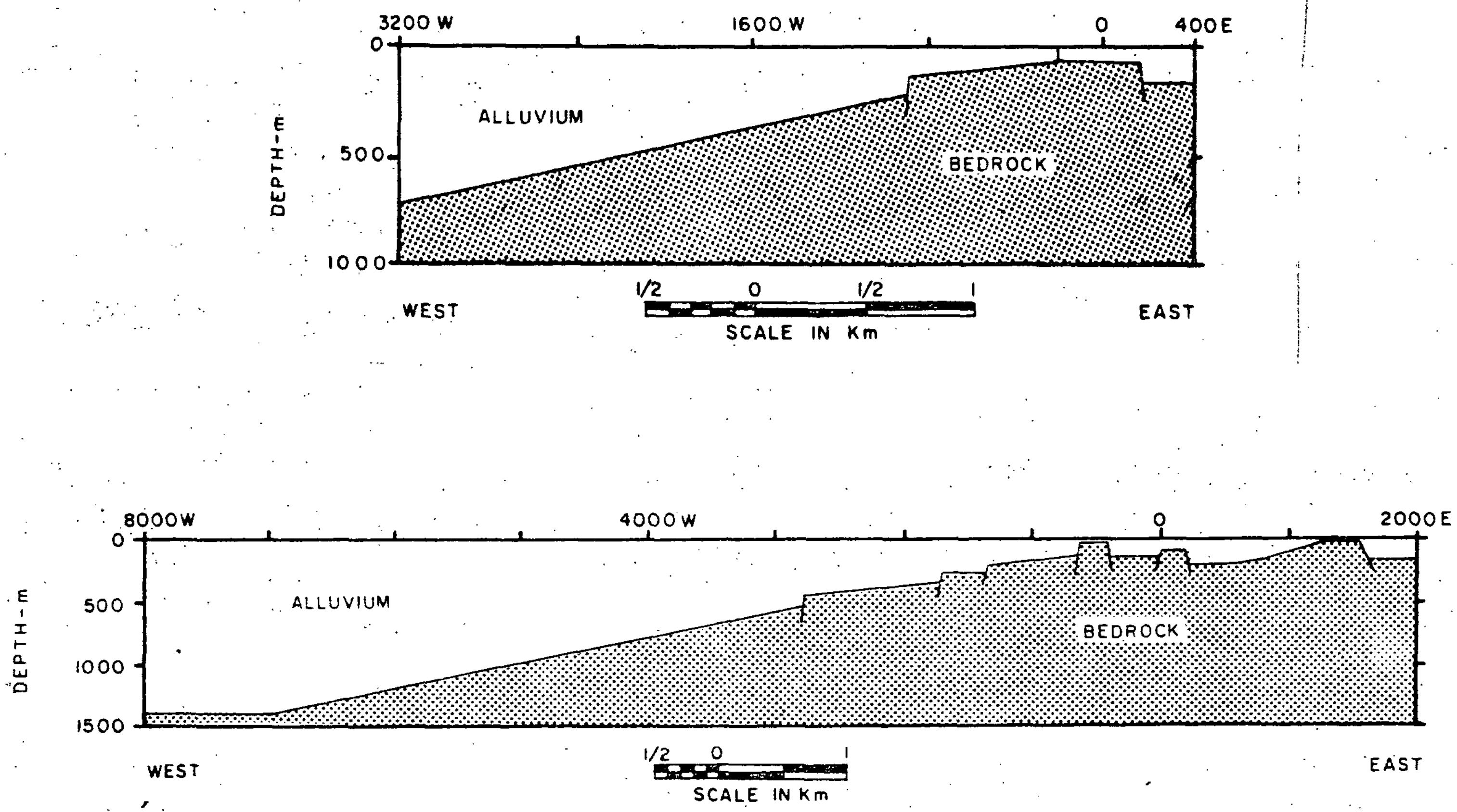


Figure 9. Interpretive two-dimensional model for gravity profiles 4000 N (top) and 2200 N (bottom). Assumed density contrast is 0.5gm/cc.

(Crebs and Cook, 1976)

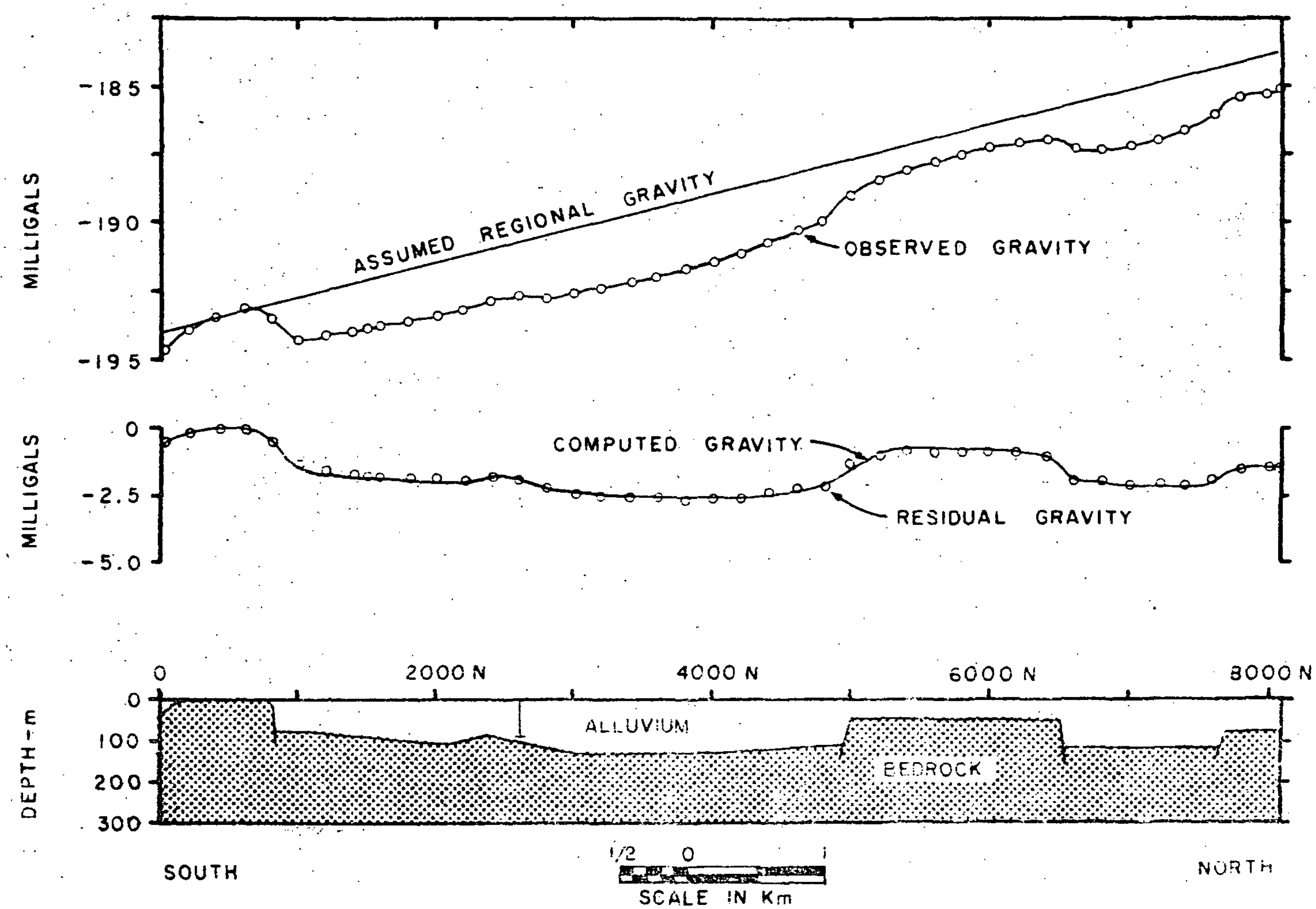


Figure 10 Interpretive two-dimensional model for gravity profile Baseline. Assumed density contrast is 0.5 gm/cc. Drillhole at 2600 N entered bedrock at a depth of 83 m.

(Crebs and Cook, 1976)

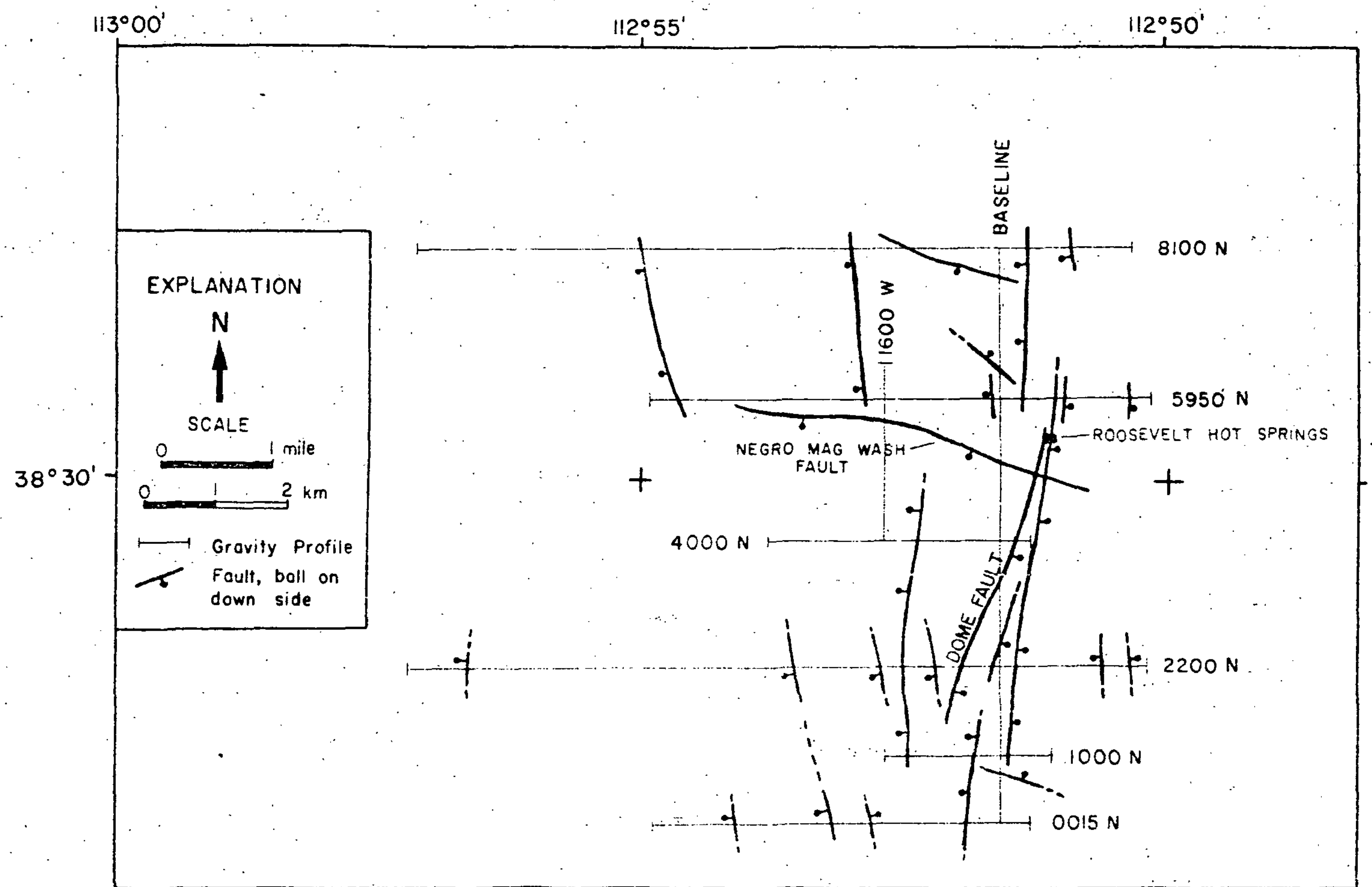


Figure 11 Faulting interpreted from the gravity profiles near the Roosevelt Hot Springs. Continuation of many faults from Ward and Sill (1976).

(Crebs and Cook, 1976)

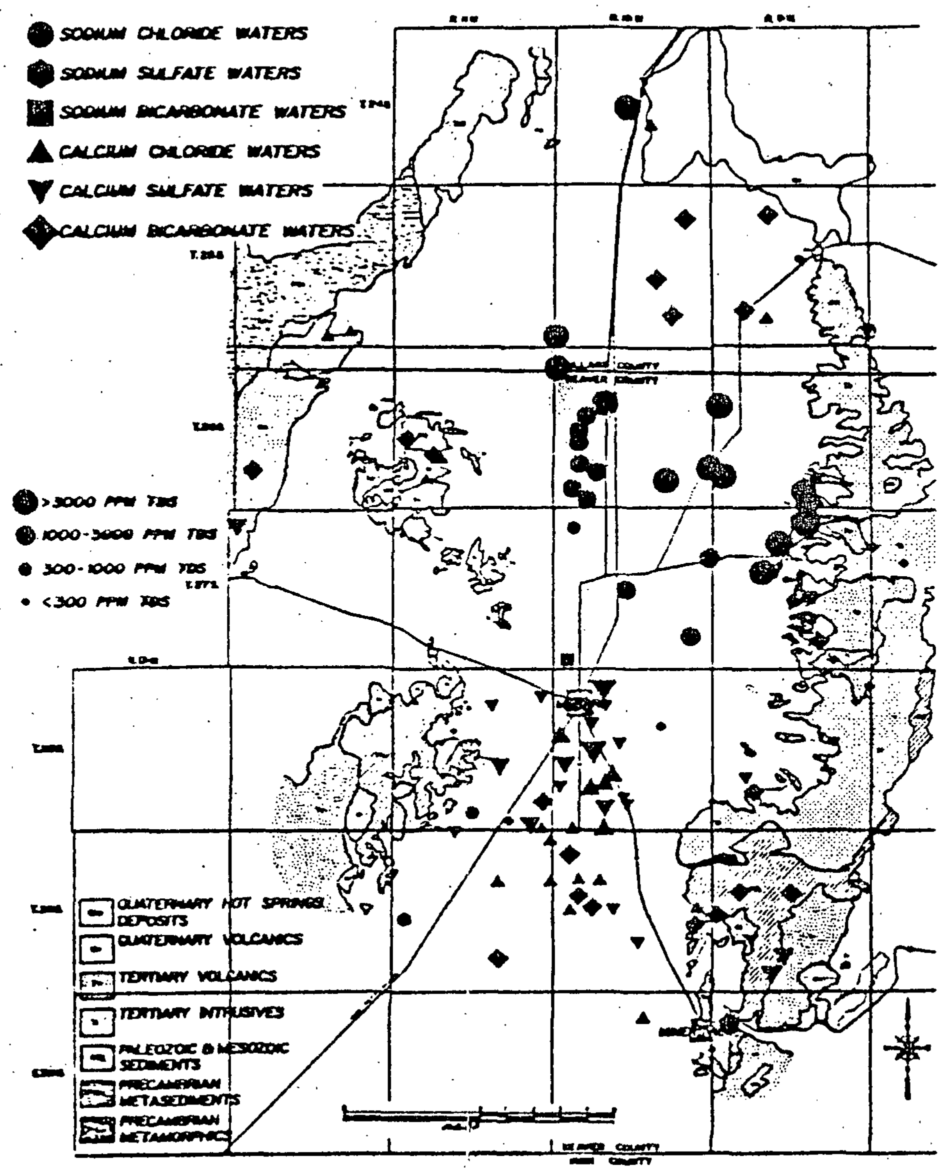
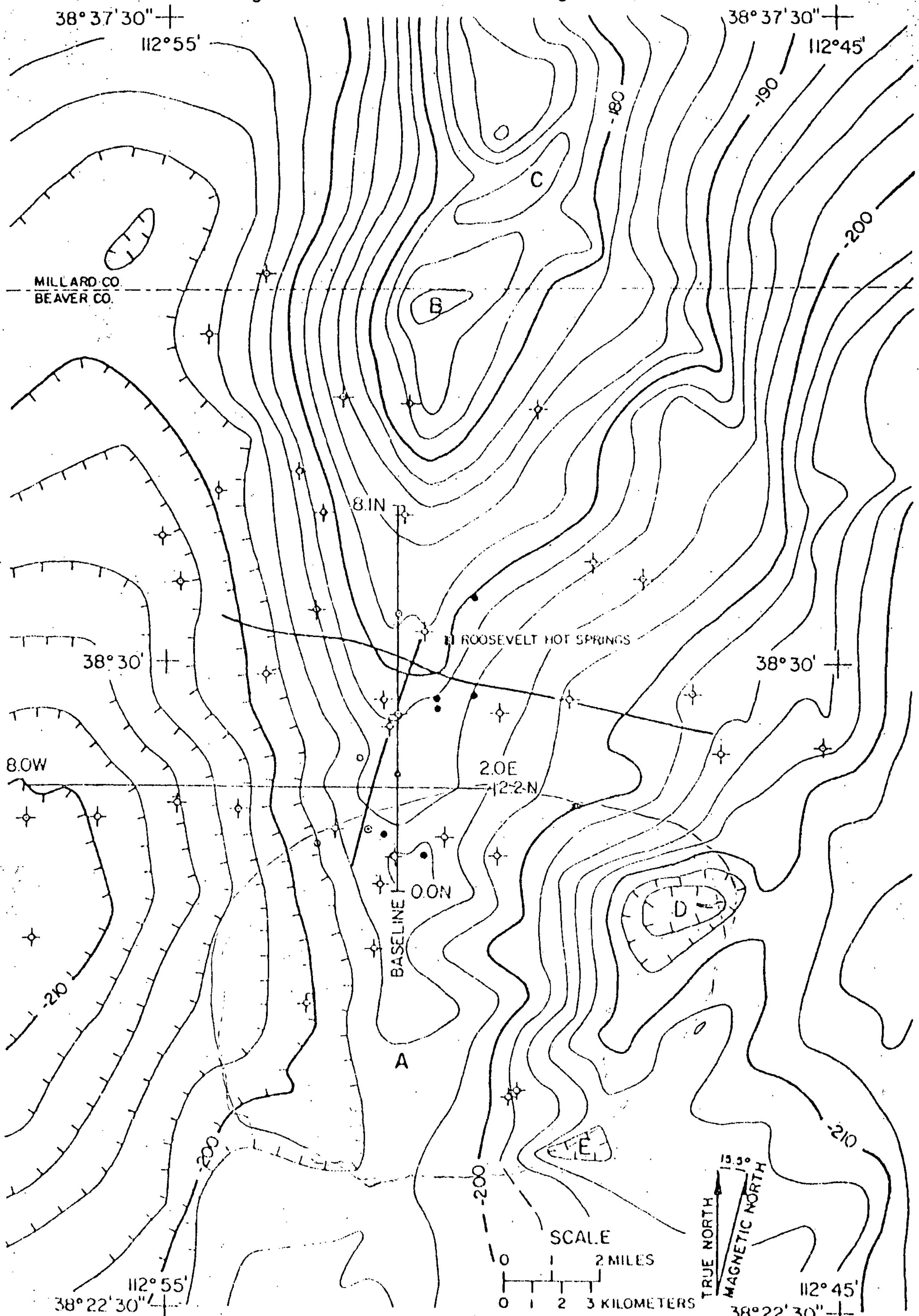


Figure 12. Water characterization map showing sample locations, water types, and water quality.

(Lenzer, Crosby and Berge, 1976)

Figure 13. Terrain-corrected (out to 20km) Bouguer gravity anomaly map of the Roosevelt Thermal Area. Contour interval=2mgal. Well designations: • productive well, ∘ nonproductive well, ⌖ thermal gradient well. Letter designations described in text.



(Ward, et. al., in press)

Figure 14. Total aeromagnetic intensity residual anomaly map of the Roosevelt Thermal Area. Contour interval=50 gammas. Letter designations are described in text. (Ward, et.al., in press)

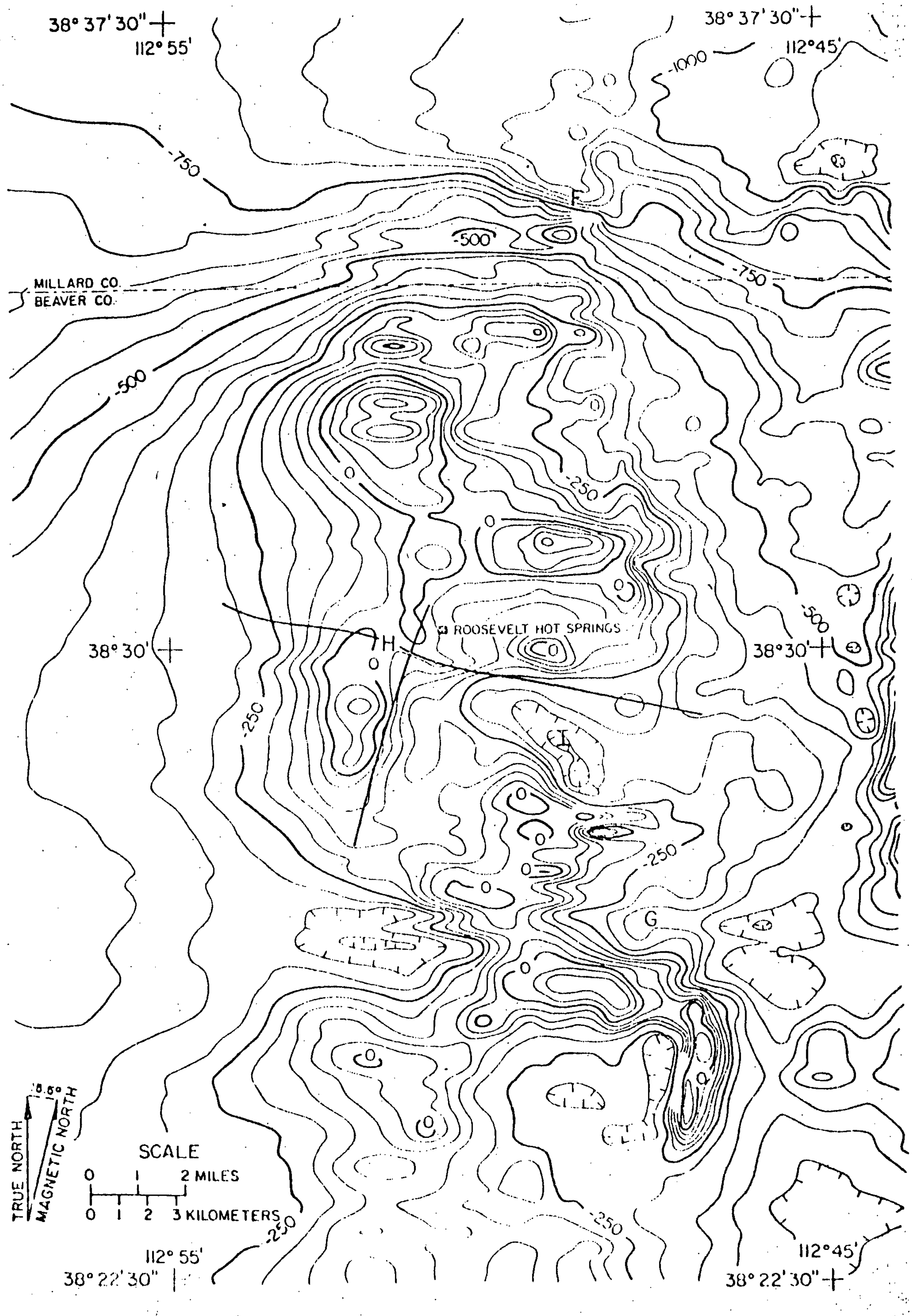
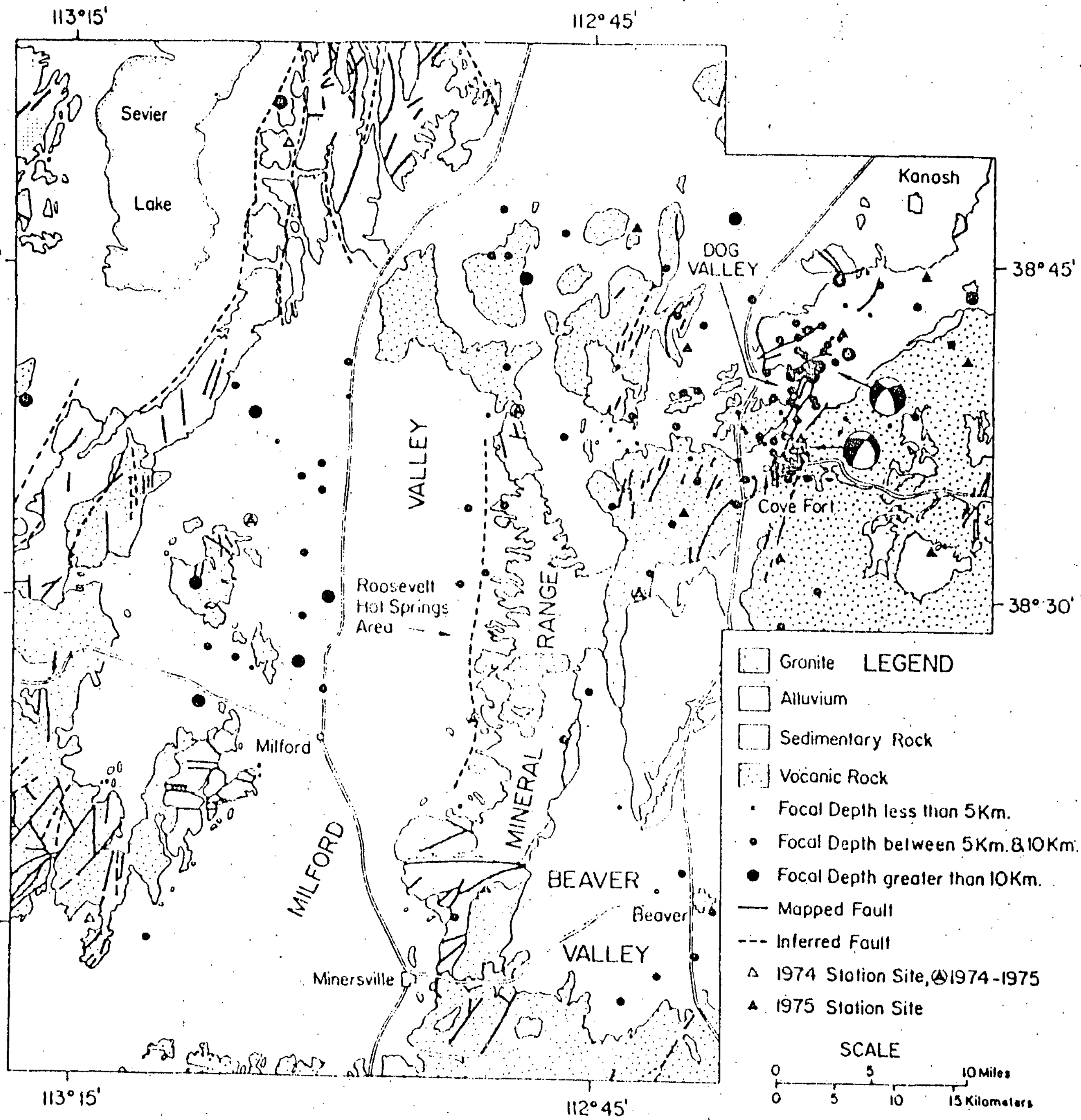


Figure 15. Epicenter map of the Roosevelt-Cove Fort, Utah area from earthquake surveys in 1974 and 1975. Fault plane solutions are lower-hemisphere, equal area stereographic nets. Dark areas are quadrants of compression; light areas are quadrants of dilation. Arrows indicate directions of minimum compressive stress (T-axes).

(Ward, et al., in press)



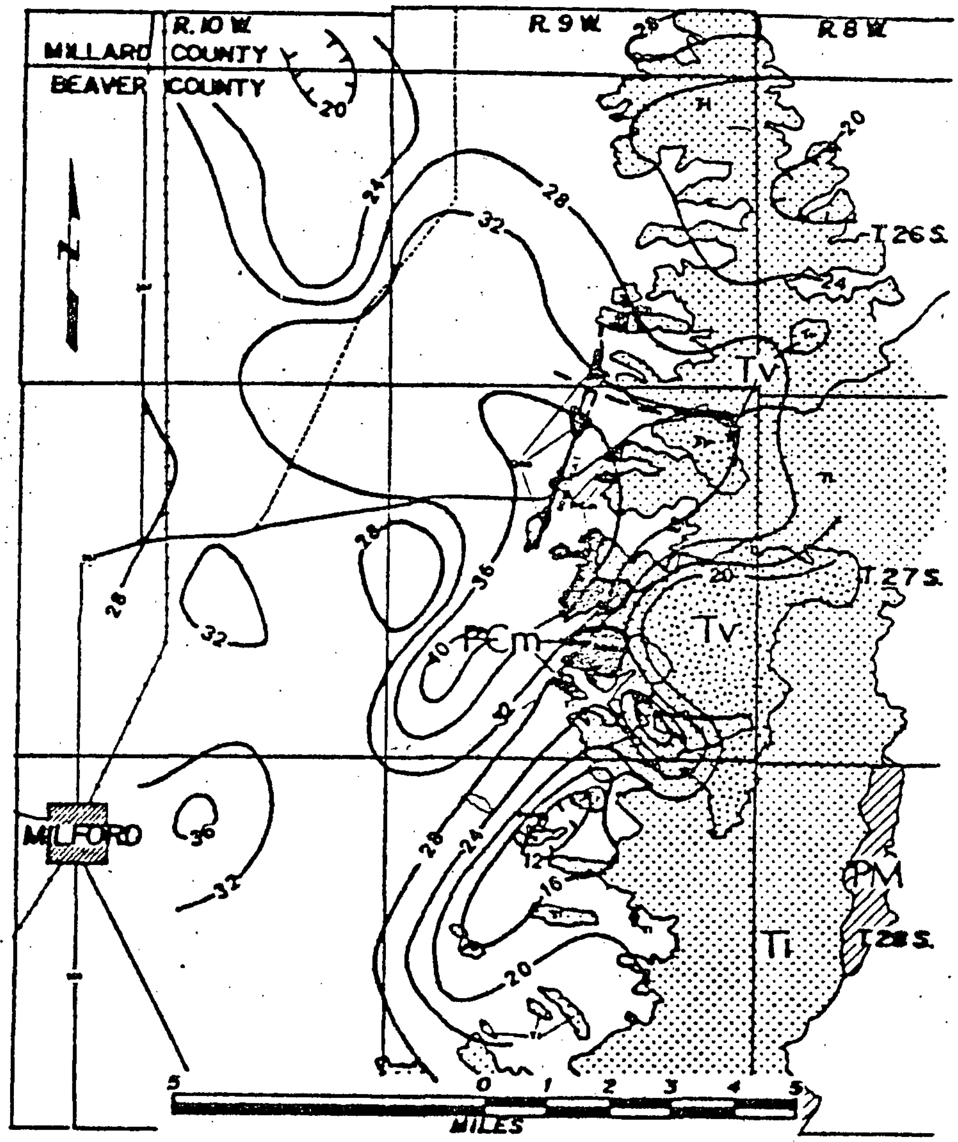
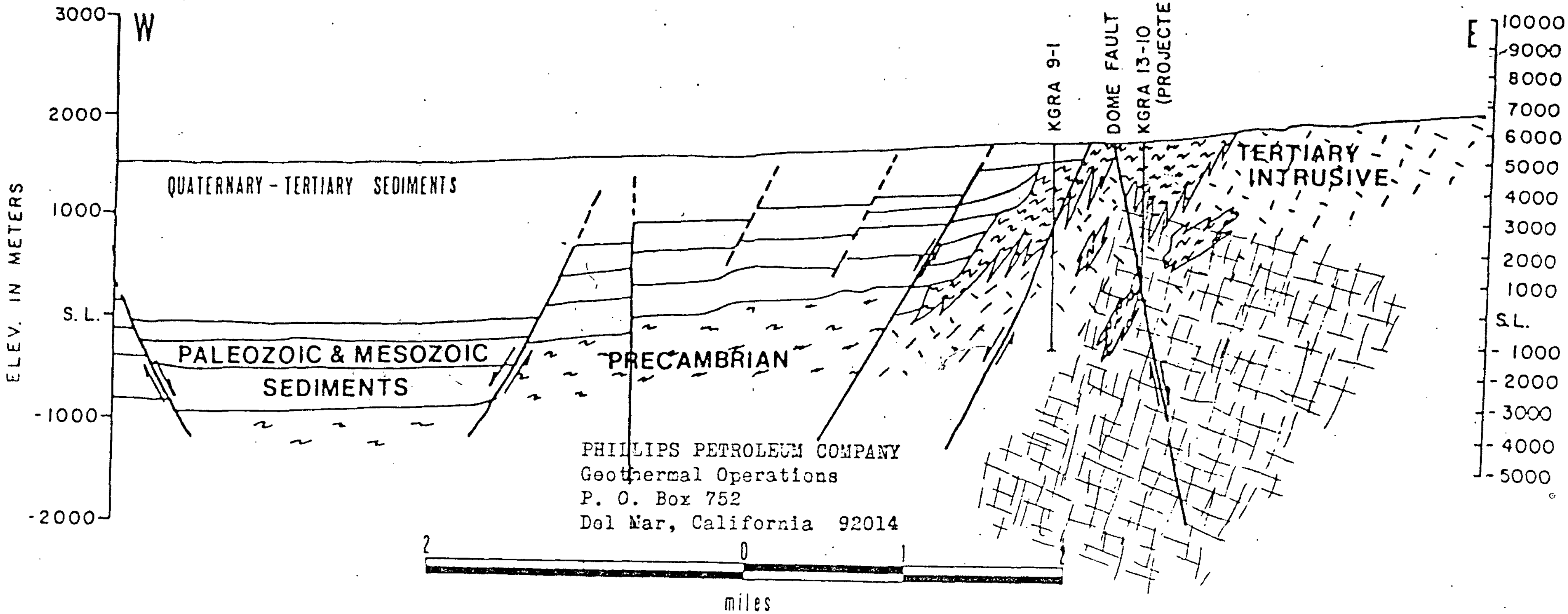


Figure 16. Ground noise anomaly map showing contoured integrated power in frequency window 9 to 11 hz.

(Lenzer, Crosby and Berge, 1976)

L-9



ELEV. IN METERS
3000
2000
1000
S.L.
-1000
-2000
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E
10000
9000
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7000
6000
5000
4000
3000
2000
1000
S.L.
-1000
-2000
-3000
-4000
-5000
QUATERNARY - TERTIARY SEDIMENTS
PALEOZOIC & MESOZOIC SEDIMENTS
PRECAMBRIAN
TERTIARY INTRUSIVE
KGRA 9-1
DOME FAULT
KGRA 13-10 (PROJECTED)
PHILLIPS PETROLEUM COMPANY
Geothermal Operations
P. O. Box 752
Del Mar, California 92014
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The American Association of Petroleum Geologists Bulletin
V. 66, No. 7 (July 1982), P. 879-902, 11 Figs., 6 Tables

Roosevelt Hot Springs Geothermal System, Utah—Case Study[1]

HOWARD P. ROSS, DENNIS L. NIELSON, and JOSEPH N. MOORE[2]

ABSTRACT

The Roosevelt Hot Springs geothermal system has been undergoing intensive exploration since 1974 and has been used as a natural laboratory for the development and testing of geothermal exploration methods by research organizations. This paper summarizes the geological, geophysical, and geochemical data which have been collected since 1974, and presents a retrospective strategy describing the most effective means of exploration for the Roosevelt Hot Springs hydrothermal resource.

The bedrock geology of the area is dominated by metamorphic rocks of Precambrian age and felsic plutonic phases of the Tertiary Mineral Mountains intrusive complex. Rhyolite flows, domes, and pyroclastic rocks reflect igneous activity between 0.8 and 0.5 m.y. ago. The structural setting includes older low-angle normal faulting and east-west faulting produced by deep-seated regional zones of weakness. North to north-northeast-trending faults are the youngest structures in the area, and they control present fumarolic activity. The geothermal reservoir is controlled by intersections of the principal zones of faulting.

The geothermal fluids that discharge from the deep wells are dilute sodium chloride brines containing approximately 7,000 ppm total dissolved solids and anomalous concentrations of F, As, Li, B, and Hg. Geothermometers calculated from the predicted cation contents of the deep reservoir brine range from 520 to 531°F (271 to 277°C). Hydrothermal alteration by these fluids has produced assemblages of clays, alunite, muscovite, chlorite, pyrite, calcite, quartz, and hematite. Geochemical analyses of rocks and soils of the Roosevelt Hot Springs thermal area demonstrate that Hg, As, Mn, Cu, Sb, W, Li, Pb, Zn, Ba, and Be have been transported and redeposited by the thermal fluids.

The geothermal system is well expressed in electrical resistivity and thermal-gradient data and these methods, coupled with geologic mapping, are adequate to delineate the fluids and alteration associated with the geothermal reservoir. The dipole-dipole array seems best suited to acquire and interpret the resistivity data, although controlled source AMT (CSAMT) may be competitive for near-surface mapping. Representations of the thermal data as temperature gradients, heat flow, and temperature are all useful in exploration of the geothermal system, because the thermal fluids themselves rise close to the surface. Self-potential, gravity, magnetic, seismic, and magnetotelluric survey data all contribute to our understanding of the system, but are not considered essential to its exploration.

INTRODUCTION

The Roosevelt Hot Springs geothermal system is located along the western side of the Mineral Mountains, approximately 12 mi (19 km) northeast of Milford, Utah (Fig. 1). The geothermal system is a high-temperature water-dominated resource, and is structurally controlled with permeability localized by faults and fractures cutting plutonic and metamorphic rocks.

The earliest exploration in the area was carried out in 1967 and 1968 by Eugene Davies of Milford who encountered boiling water in two holes on and adjacent to siliceous sinter deposits. Phillips Petroleum Co. initiated geothermal exploration at Roosevelt in 1972 and successfully bid for 18,871 acres (7,637 ha.) of the Roosevelt KGRA (Known Geothermal Resources Area) at the July 1974 competitive lease sale (Forrest, 1980). Exploratory drilling began in 1975 and the second well drilled, Roosevelt KGRA 3-1, encountered producible quantities of geothermal fluid. Additional drilling by Phillips Petroleum Co. and Thermal Power Co. has brought the total number of successful wells in the Roosevelt Hot Springs geothermal field to seven. A mobile 1.6 MWe generator has been installed in the field. Present development plans call for increasing capacity to 7 MWe in 1983 and 20 MWe by 1984. A discussion of the history of the Roosevelt Hot Springs KGRA and the exploration methods employed by Phillips Petroleum Co. is presented by Forrest (1980).

A large amount of geological, geophysical, geochemical, drilling, and reservoir data for the Roosevelt Hot Springs area has been made available in 1977-79 through the Department of Energy's Industry Coupled Program. A review of these data and earlier work helped to identify inconsistencies and weak areas in an already large geoscience data base. Additional work has helped to complete this data base. The integration of these data constitutes the present case study.

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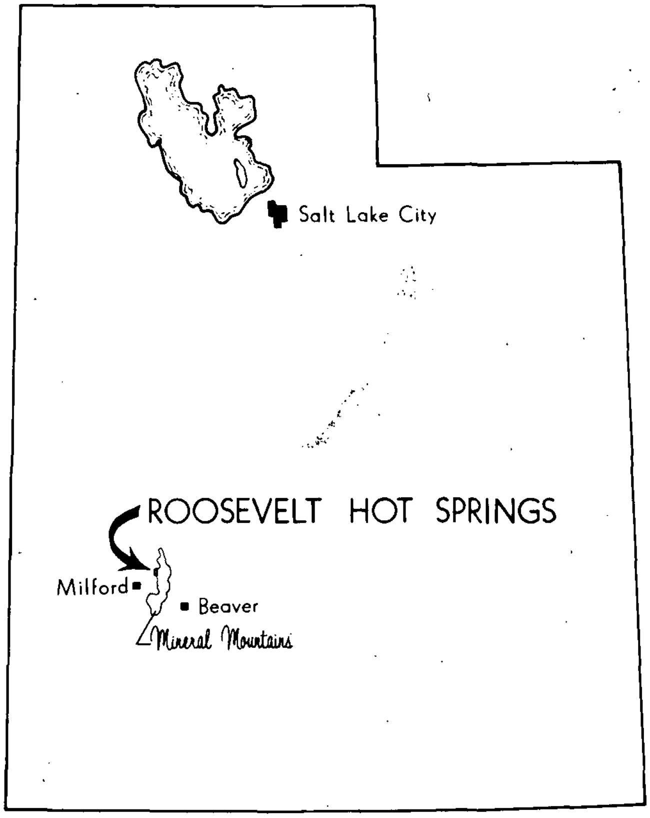

FIG. 1—Index map showing location of Roosevelt Hot Springs KGRA, Utah.

ROOSEVELT HOT SPRINGS GEOTHERMAL SYSTEM

Regional Setting

The Roosevelt Hot Springs KGRA is located on the western flank of the Mineral Mountains (Fig. 2). The Mineral Mountains are a north-trending horst bounded by Basin and Range normal faults and lie at the western edge of the transition zone between the Colorado Plateau and the Basin and Range physiographic provinces. The area is located on the western edge of the Intermountain seismic belt as defined by Smith and Sbar (1974). In addition, the Roosevelt Hot Springs KGRA lies along east-west-trending magnetic anomalies which follow the trend of the Wah Wah–Tushar mineral belt (Mabey et al, 1978). This belt has been the site of intrusive activity through the Tertiary and into Quaternary time. Associated with this igneous activity are deposits of uranium and base and precious metals.

The central part of the Mineral Mountains is a structural high relative both to adjacent ranges and also to the northern and southern parts of the range. In these northern and southern areas, sedimentary rocks of Cambrian through Cretaceous age are exposed. The southern area also contains volcanic and plutonic rocks of Tertiary age (Earll, 1957). In contrast, the central part of the Mineral Mountains contains Precambrian metasedimentary rocks and Tertiary plutonic rocks of the Mineral Mountains intrusive complex (Sibbett and Nielson, 1980). These Tertiary rocks possibly represent plutonic equivalents of the Marysvalle volcanic province which is exposed to the east, south, and southwest of the Mineral Mountains.

Local Setting

The geology in the vicinity of the Roosevelt Hot Springs geothermal system has been described in detail by Nielson et al (1978). The central part of the Mineral Mountains has been mapped by Sibbett and Nielson (1980). A simplified geologic map of the Roosevelt Hot Springs area is shown in Figure 2.

The oldest unit exposed in the area of the geothermal system is a banded gneiss which was formed from regionally metamorphosed quartzo-feldspathic sediments. The rock was metamorphosed to the upper amphibolite facies during middle Proterozoic time. The banded gneiss is strongly layered with adjacent layers distinguished principally on the content of mafic minerals. The rock is compositionally heterogeneous and contains thick sequences of quartzo-feldspathic rocks. The unit also contains metaquartzite and sillimanite schist layers which have been differentiated in the more detailed geologic study (Nielson et al, 1978).

The Mineral Mountains intrusive complex is the largest intrusive body exposed in Utah. K-Ar dating and regional relationships suggest that the intrusive sequence is middle to late Tertiary in age. In the vicinity of the geothermal system, the lithologies range from diorite and granodiorite through granite and syenite in composition (Fig. 2).

Rhyolite flows, pyroclastics, and domes were extruded along the spine of the Mineral Mountains 800,000 to 500,000 years ago (Lipman et al, 1978). The flows and domes are glassy, phenocryst-poor rhyolites. The pyroclastic rocks are represented by air-fall tuff and nonwelded ash-flow tuffs. Smith and Shaw (1975) hypothesized that young rhyolites such as these are indicative of an upper-level magma chamber which could serve as a heat source for geothermal systems.

Hot spring deposits in the vicinity of the geothermal system have been mapped as siliceous sinter, silica-cemented alluvium, hematite-cemented alluvium, and manganese-cemented alluvium. The principal areas of hot-spring deposition are along the Opal Mound fault and at the old Roosevelt Hot Springs. In both of these areas, the deposits consist of both opaline and chalcedonic sinter.

The geothermal reservoir at Roosevelt Hot Springs is structurally controlled. The controls are thought to be produced by the intersection of the faults mapped in the KGRA (Nielson et al, 1978, 1979). The structural evolution of the Roosevelt Hot Springs geothermal system is envisioned as follows. During rapid uplift of the Mineral Mountains structural block, westward-dipping low-angle normal faults formed. The most important of these is the Wild Horse Canyon fault (Fig. 2). This tectonic event produced intense zones of cataclasis both along low-angle fault planes and within the hanging wall of the principal fault block. The cataclastic zones within the hanging wall are steeply dipping zones up to 10 ft (3 m) wide which strike generally to the northwest. These zones in the hanging wall were produced by internal brecciation and interaction between rigid blocks during the low-angle faulting. East-west-trending high-angle normal faulting, represented by the Negro Mag fault, cuts the low-angle faults. The trend is parallel with the Wah Wah–Tushar mineral belt and is probably related to movement along this deep-seated structural trend. The Opal Mound fault and parallel structures are north-northeast-trending faults which are the youngest structures in the area. They localize siliceous hot-spring deposits and are often marked by zones of alteration and silicification of alluvium.

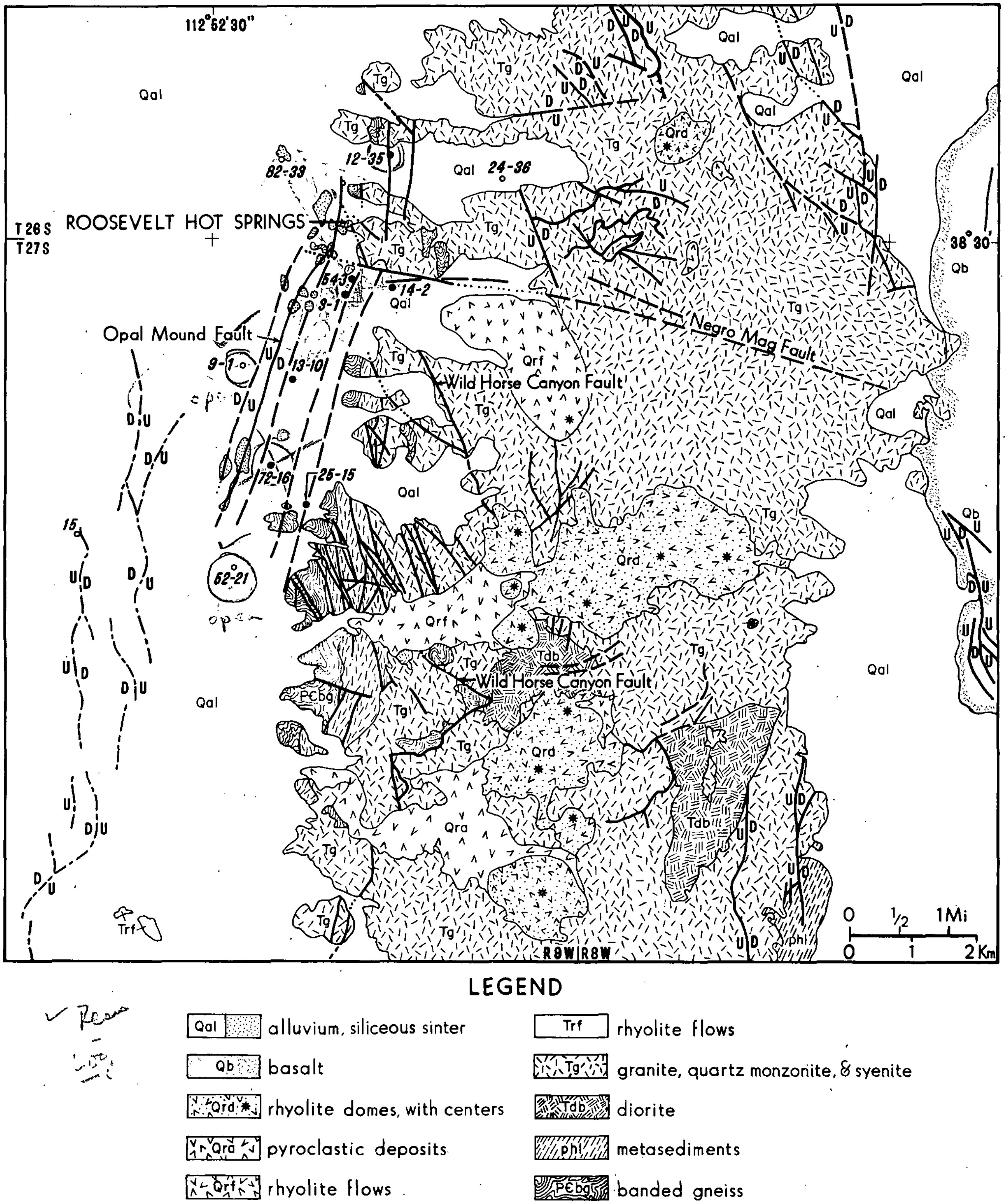

FIG. 2—Geologic map of Roosevelt Hot Springs KGRA and vicinity (after Sibbett and Nielson, 1980a).

Hydrothermal System

The limits of the hydrothermal system have been partially defined by deep drilling within the Roosevelt Hot Springs KGRA. The principal wells are shown on Figure 2, and important facts about the wells are presented in Table 1. The area of geothermal production is bounded on the east by the range front of the Mineral Mountains. On the west, the system is bounded by the Opal Mound fault, and on the south it terminates between Utah State 72-16, which is a producer, and Utah State 52-21, which is a hot but dry hole. The northern boundary of the system has not been determined.

The deep wells confirm that the host rocks for the hydrothermal system are the Tertiary plutonic rocks and Precambrian metamorphic rocks which have been mapped in the adjacent Mineral Mountains (Fig. 2). Thus, the rocks show little primary permeability and the system is controlled by faults and fractures (Lenzer et al, 1976; Nielson et al, 1978, 1979).

Fluid chemistry.—The hydrothermal fluids are relatively dilute sodium chloride brines which contain approximately 7,000 ppm total dissolved solids. Capuano and Cole (1982) provide the most comprehensive review of the chemistry of these fluids. Table 2 lists the composition of fluids collected from the wells and springs. These fluids are compositionally similar throughout the field, and differ mainly in their concentrations of calcium, magnesium, and bicarbonate which may be the result of mixing with calcium-rich nonthermal ground water.

Reservoir temperatures estimated from the Na-K-Ca contents of the brines range up to 554°F (290°C), and exceed the measured maximum temperatures by as much as 36 to 54°F (20 to 30°C). Capuano and Cole (1982) suggest that temperatures calculated by geothermometers may in fact be too high, reflecting changes in the fluid chemistry produced by flashing of the brines in the well bore. Their arguments are supported by geothermometer calculations based on the composition of the deep reservoir fluid determined from the combined analyses of brine, steam, and gas. These calculations suggest a reservoir temperature of 520°F (271°C) based on the Na-K-Ca geothermometer (Fournier and Truesdell, 1973) and a temperature of 531°F (277°C) based on the SiO_2 content of the fluid and the conductively cooled quartz saturation model of Fournier (1973).

The isotopic composition of the geothermal fluids indicates that they are of meteoric origin (Rohrs, 1980). A comparison of the fluid isotopic compositions with that of water from the mountain ranges to the east suggests that the thermal fluids could be derived either from the Mineral Mountains or ranges to the east.

Hydrothermal alteration.—Hydrothermal alteration in the geothermal system and the adjacent Mineral Mountains is localized along faults and fractures. There have been several periods of alteration and it is often

Table 1. Well Summary from Roosevelt Hot Springs, Utah

Well	Location	Depth	Status	T Max	Operator	Reference
O.H.-2	SW NW Sec. 10, T27S, R9W	2,250′	Deep T-gradient	NA	Phillips	Lenzer et al, 1977
O.H.-1	SE NE Sec. 17, T27S, R9W	2,321′	Deep T-gradient	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 9-1	NE NW Sec. 9, T27S, R9W	6,885′	dry	227°C	Phillips	Lenzer et al, 1977 Glenn et al, 1980
Roosevelt KGRA 3-1	SW NE Sec. 3, T27S, R9W	2,724′	producer	NA	Phillips	Lenzer et al, 1977
Roosevelt KGRA 54-3	SW NE Sec. 3, T27S, R9W	2,882′	producer	NA	Phillips	Lenzer et al, 1977
Roosevelt KGRA 12-35	NW NW Sec. 35, T26S, R9W	7,324′	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 13-10	SW NW Sec. 10, T27S, R9W	5,351′	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 82-33	NE NE Sec. 33, T26S, R9W	6,028′	dry	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Utah State 14-2	SW NW Sec. 2, T27S, R9W	6,108′	producer	254°C	Thermal Power	Lenzer et al, 1977 Glenn and Hulen, 1979
Roosevelt HSU 25-15	NW SW Sec. 15, T27S, R9W	7,500′	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Utah State 72-16	SE NE Sec. 16, T27S, R9W	1,254′	producer	243°C	Thermal Power O'Brien	Lenzer et al, 1977 Glenn and Hulen, 1979
Utah State 24-36	SW NW Sec. 36, T27S, R9W	6,119′	dry	NA	Thermal Power	
Utah State 52-21	NW NE Sec. 21, T27S, R9W	7,500′	dry	206°C	Getty Oil Co.	Glenn and Hulen, 1979
GPC-15	SE SE Sec. 18, T27S, R9W	1,900′	Deep T-gradient	72°C	Geothermal Power Corp.	Glenn and Hulen, 1979

not possible to separate these different periods on the basis of their mineralogy and chemistry. Older periods of hydrothermal alteration are associated with the intrusion of various phases of the Mineral Mountains intrusive complex. These episodes have produced minor copper mineralization which is generally associated with xenoliths of Paleozoic and Precambrian rocks. In addition, zones of sodium metasomatism have been identified with the contact zones between some of the felsic plutonic rocks (Sibbett and Nielson, 1980). The intrusion of the various phases of the pluton have also superimposed contact metamorphic assemblages on the older rocks.

A hydrothermal event which altered the fault zones and deposited pyrite and chalcopyrite occurred with the low-angle faulting. The hydrothermal alteration produced assemblages of quartz + chlorite + epidote + hematite. Hematite is commonly found as specularite veinlets and, where genetic relationships can be observed, hematite mineralization follows sulfide mineralization.

The hydrothermal alteration assemblages associated with the present geothermal system are crudely zoned with depth. The uppermost assemblage, occurring around the hot-spring deposits and fumeroles, is characterized by quartz, alunite, kaolinite, montmorillonite, hematite, and muscovite. Parry et al (1980) have studied the near-surface alteration and suggest that these minerals have formed above the water table by downward-percolating acid sulfate waters. Upward-convecting geothermal brines have produced, with increasing depth, alteration assemblages characterized by montmorillonite + mixed layer clays + sericite + quartz + hematite and chlorite + sericite + calcite + pyrite + quartz + anhydrite (Ballantyne, 1978). Thermochemical calculations and petrologic observations suggest that the brines are in equilibrium with the alteration assemblages produced by the upward-migrating fluids (Capuano and Cole, 1982).

Flow tests.—Most of the flow-test, production-logging, and reservoir-engineering data for the Roosevelt Hot Springs field is proprietary. However, data from Utah State 14-2 and Utah State 72-16 (Fig. 2, Table 1) are in the public domain and are summarized here.

A short flow test conducted in Utah State 72-16 on

Table 2. Chemical Analyses of Selected Thermal Waters[1]

		Wells				
		14-2	54-3	72-16	52-21	Hot Spring
Na	ppm	2,150	2,320	2,000	1,900	2,500
K	ppm	390	461	400	218	488
Ca	ppm	9.2	8	12.20	114	22
Mg	ppm	0.6	<2	0.29	3.9	0.
Fe	ppm		0.03		6.9	
Al	ppm		<0.5		<0.1	0.04
Si	ppm	229	263	244	67	146
Sr	ppm		1.2	1.20		
Ba	ppm		<0.5			
Mn	ppm		<0.2			
Cu	ppm		<0.1			
Pb	ppm		<0.2			
Zn	ppm		<0.1			
As	ppm	3.0	4.3			
Li	ppm		25.3	16.0		0.27
Be	ppm		0.005			
B	ppm	29	29.9	27.2	27.0	38
Ce	ppm		0.27			
F	ppm	5.2	6.8	5.3	3.4	7.5
Cl	ppm	3,650	3,860	3,260	2,885.1	4,240
HCO_3	ppm		232	181	550.0	156
SO_4	ppm	78	72	32	86	73
NO_3	ppm				1.3	11
Total Dissolved Solids	ppm	>6,614	7,504	6,444	5,727	7,800
pH (at collection T)		5.9		7.53	7.3	7.9
T (at surface)	°C	14				55
T (bottom hole)	°C	268	>260	243	204	284
Geothermometer						
T (Na-K-Ca)	°C	286	297	288	210[2]	284
T (quartz cond.)	°C	276	263	256	158	212
T (quartz adiab.)	°C	244	234	229	150	194
T (chalcedony)	°C	274	259	250	134	197
Total Depth	m	1,862	878	382	2,289	

[1]From Capuano and Cole (1982). A blank indicates data not determined or information not available.
[2]Corrected for the Mg content of the fluid.

December 30, 1976, produced 454,546 kg/hr steam and hot water at a flowing wellhead pressure of 25 kg/cm^2 and a temperature of 432°F (222°C). A 24-hour test was conducted April 4 to 6, 1977, in 72-16. The total mass-flow rate was determined by the James method (James, 1966) as 595,000 kg/hr at a wellhead pressure of 20.7 kg/cm^2 psia and a temperature of 415°F (213°C). However, the mass-flow rate and wellhead pressure dropped throughout the test, indicating a longer flow test was needed to determine steady-state well production. Thermal Power Co. also concluded that the James method of determining the flow rate of a two-phase flow was unsatisfactory. On the basis of the longer test, it was calculated that Utah State 72-16 could yield 119,546 kg/hr of steam if flashed at 5.6 kg/cm^2. The well's electrical generating capacity was determined to be 12.5 Mw at a heat rate of 9,546 kg steam/hr/Mw.

Thermal Power Co. conducted a 48-hour flow test in geothermal well Utah State 14-2 between November 16 and 18, 1976. The last 35 hours of the test stabilized at 225,000 kg/hr and an enthalpy of 444.5 Btu/hr. Thermal Power Co. calculates that at a wellhead pressure of 4.9 kg/cm^2 and 17.8% flash, Utah State 14-2 has a generating capacity of 4.5 Mw. This capacity was calculated at a heat rate of 8,900 kg steam/hr/Mw. Well surges were observed during the test and their cause was attributed to temporary obstructions in the system that induced in-pipe flashing.

The Denver Research Institute (DRI) conducted several production logging tests in 1978 and 1979 (Butz and Plooster, 1979). Pressure and temperature data were logged versus depth during several constant flow-rate tests from 146,818 kg/hr to 263,636 kg/hr as measured by the James method. The flow rates are probably accurate to ±15%. Butz and Plooster (1979) were able to match the measured borehole temperature and pressure profiles by using a program that modeled two-phase flow in the well bore and determined the flash depth in the borehole and the productivity index of the well. The best models were those that included a component of compressible gas.

EXPLORATION METHODS

The Roosevelt Hot Springs geothermal area has served as a laboratory for the development and testing of geothermal exploration methods for some time (Ward et al, 1978). Since 1978 more exploration data have become available, including detailed geologic mapping, additional deep drill holes, additional electrical resistivity surveys, a reflection seismic profile, passive seismic data, and extensive application and testing of surface and subsurface geochemical methods. These new data allow us to expand upon earlier studies. With the results of the individual exploration methods and the supreme advantage of 20/20 hindsight, the most ef-

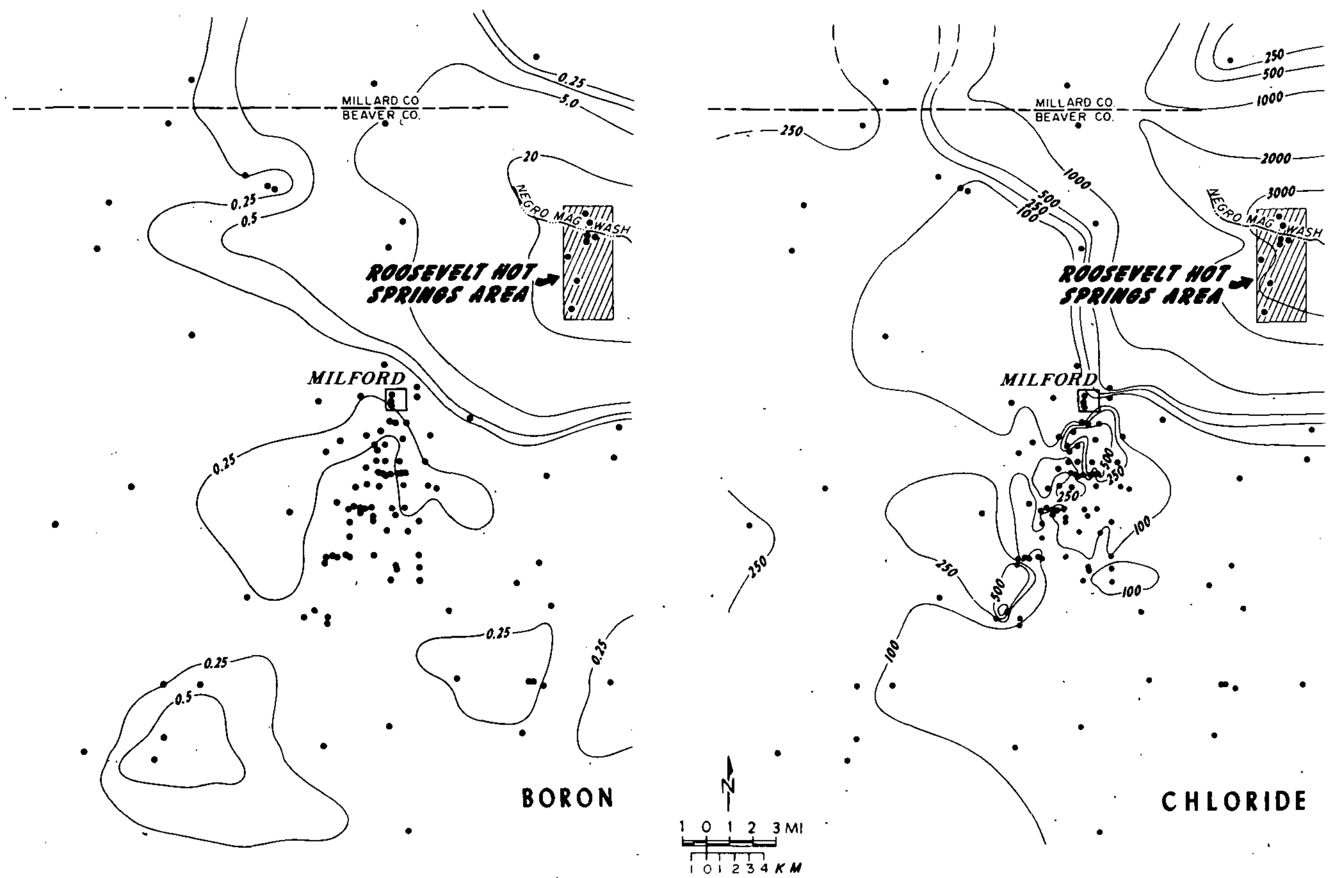

FIG. 3—Concentrations of boron and chloride in ground water in Roosevelt Hot Springs KGRA and adjacent Milford Valley. Data are from Mower and Cordova (1974) and Mower (1978) and was compiled by D. R. Cole.

ficient strategy for the exploration of the Roosevelt Hot Springs geothermal system will be presented. The results of some of the geological and geochemical exploration methods have been presented in the previous section.

Ground Water—Flow Systems and Chemistry

In contrast to the geothermal reservoir, the ground-water aquifer in the vicinity of the Roosevelt Hot Springs system is nonconfined. The study of this ground-water system is important from two aspects. First, it is probable that the present ground-water system provides the recharge to the thermal reservoir. This information bears on estimates of the longevity of the system and also whether the system will remain water-dominated during production or become vapor dominated (White et al, 1971) or possibly depleted. Second, studies of ground-water flow and chemistry are important exploration procedures which can be used to detect leakage of thermal waters into the regional ground-water systems. Hydrologic data can be inexpensively collected during thermal-gradient and heat-flow studies and for many areas are available in the ground-water literature.

Reconnaissance ground-water studies of the Milford and Beaver Valleys (Mower and Cordova, 1974; Mower, 1978) do not discuss the hydrology in the geothermal area but they do reveal the generalized ground-water configuration. The probable direction of ground-water movement within the KGRA is west-northwest toward Beaver Bottoms. The much higher elevation of the potentiometric surface in Beaver Valley than in Milford Valley—more than 984 ft (300 m) near the KGRA—prompts the speculation that if there were sufficiently transmissive fractures within the crystalline units of the Mineral Mountains, the difference in head would force ground water to flow through the range. The chances for this interbasin ground-water flow are enhanced by the cross-cutting structural zones like the Negro Mag fault zone. If ground water originating in the Beaver Valley drainage is recharging the geothermal reservoir, vast quantities of recharge may be available to the geothermal wells.

The ground-water papers by Mower and Cordova (1974) and Mower (1978) contain additional information which can be of value in exploration for the geothermal resource. Figure 3 is a map showing concentrations of chloride and boron in the vicinity of the Roosevelt Hot Springs. Lenzer et al (1976) show plots of ground water total dissolved solids which are similar to the patterns developed in Figure 3. These plots indicate that the chemical patterns produced by leakage of geothermal fluids into the ground-water system can be detected at significant distances from the geothermal system. These data are commonly available in the published literature, and its compilation and analysis can be a very efficient exploration method.

Solids Geochemistry

The geochemistry of hydrothermally altered rocks is routinely used by the minerals exploration industry as a guide to buried ore deposits. Yet despite the similarities between hydrothermal mineral deposits and active geothermal systems, the development of geochemical zoning models by the geothermal industry has been largely neglected. Recent studies (Browne, 1971; Ewers and Keayes, 1977) have suggested that trace-element zoning may be developed around the high-temperature centers of active geothermal systems. The studies at Broadlands have demonstrated that volatile and base metals are crudely zoned with depth as a result of decreasing temperature and boiling in permeable zones.

The trace- and major-element contents of surface and drill-hole samples from the Roosevelt Hot Springs thermal field were documented to determine if geochemical zoning models could be developed into an effective exploration tool (Bamford, 1978). Thirty-four elements were analyzed using an inductively coupled argon plasma spectrometer. Arsenic was determined by colorimetric methods, and mercury by a gold film detector. The analytical techniques are described by Christensen et al (1980a) and Capuano and Bamford (1978).

Geochemical analyses demonstrate that Hg, As, Sb, Mn, Cu, Co, W, Li, Pb, Zn, Be, Sr, and Ba have been transported and redistributed by the thermal fluids. These elements are strongly concentrated in hot-spring deposits (Table 3) with surface concentrations being typically more than an order of magnitude higher than

Table 3. Analyses of Hot Spring Deposits*

	Chalcedonic Sinter, Opal Mound	Mn-Cemented Alluvium	Altered Alluvium Over Fumarole
Na (%)	0.15	1.79	0.07
K (%)	0.14	3.12	0.26
Ca (%)	0.14	0.39	0.06
Mg (%)	0.01	0.10	0.03
Fe (%)	0.02	0.74	0.16
Al (%)	0.09	5.18	4.01
Ti (ppm)	19	560	2,040
P	—	651	336
Sr	33	386	266
Ba	—	4.9%	326
Cr	—	9	24
Mn	388	18.8%	173
Co	—	28	—
Ni	—	—	—
Cu	—	231	3
Mo	—	5	—
Pb	—	68	15
Zn	1	23	7
Cd	—	4	1
As	145	858	5
Sb	243	291	21
W	—	2,940	29
Li	11	17	8
Be	99.8	18.6	3.4
Zr	—	17	42
La	—	37	34
Ce	—	42	56
Hg (ppb)	352	2,210	49,300

*From Bamford et al (1980). Dash indicates that element was not detected.

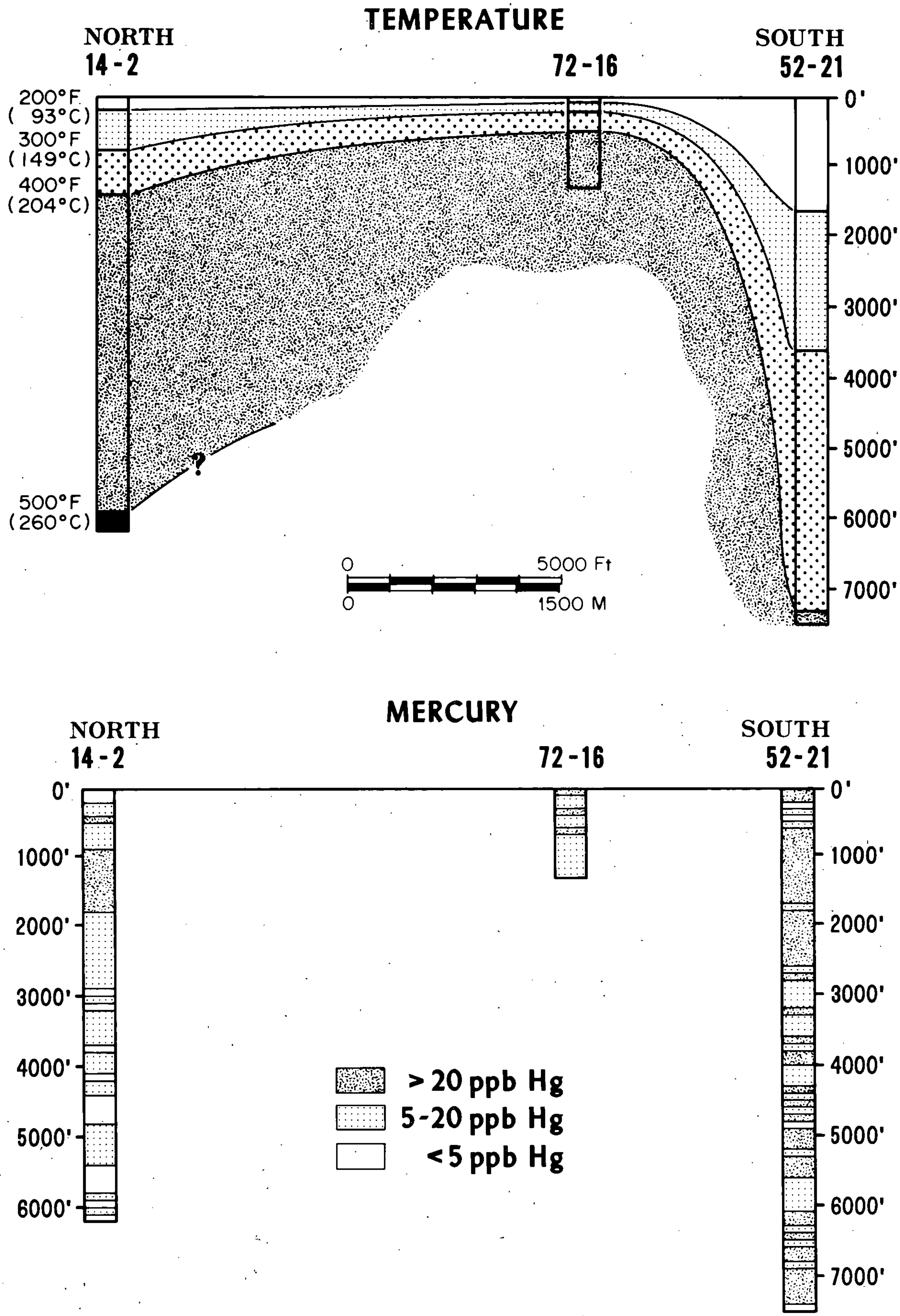

FIG. 4—Distribution of mercury and temperatures in wells 14-2, 72-16, and 52-21.

the subsurface abundances. However because of contamination during drilling, Hg, As, Sb, and Li are the only trace elements whose abundances in drill cuttings can be unambiguously related to chemical redistribution by geothermal activity.

Trace-element distributions, like the distribution of hydrothermal alteration minerals, are not pervasive but are localized along fractures and permeable alluvial horizons that have served as fluid channels. Detailed lithologic and geochemical logging has shown that at depth, arsenic is diagnostic of the hydrothermally altered rocks in the central part of the thermal field. Electron microprobe analyses and selective chemical leaching of the drill cuttings indicate that arsenic occurs primarily within pyrite and crystalline iron oxides formed from pyrite. Concentrations of arsenic as high as 4% have been detected in some pyrite crystals. In general, concentrations of arsenic are irregularly distributed within individual pyrite grains suggesting that the composition of the thermal brines has varied during the life of the system.

Mercury, in contrast to arsenic, is concentrated in the cooler parts of the thermal field, occurring in both weakly and highly altered rocks. Concentrations of mercury greater than 20 ppb define a broad envelope in the outer part of the system to depths approximately marked by the 420°F (215°C) isotherm (Fig. 4). This distribution reflects the extreme mobility of mercury at high temperatures within the thermal system.

Christensen et al (1980b) have experimentally investigated the mobility of mercury by measuring its progressive loss from drill cuttings and surface samples heated in air. These studies demonstrated that by 392°F (200°C) mercury loss had become significant, and had reached a maximum by 482°F (250°C). They concluded that mercury is present within the geothermal reservoir mainly as a native metal and suggested that its distribution reflects the present thermal configuration of the geothermal system.

At depths between 98.4 and 196.9 ft (30 and 60 m), pronounced enrichments in both arsenic and mercury occur in cuttings from thermal gradient holes located within the productive part of the thermal field. Despite the relatively small number of samples and the wide spacing between drill holes, geochemical data from shallow thermal-gradient holes nevertheless appear to be a useful means of prioritizing drilling targets that is independent of temperature measurements.

Concentrations of mercury and arsenic in soils over the thermal system are closely associated with hot-springs deposits and faults that are connected to the geothermal reservoir. The distribution of the anomalies and their shapes confirm that northeast- and west-northwest-trending (Negro Mag) structures control the near-surface hydrology of the thermal fluids.

Geophysics

Gravity Methods

Gravity methods are widely used in geothermal exploration to map structure and, in some cases, to detect directly the depositional products of hydrothermal systems. The latter use has been demonstrated in the Imperial Valley of California (Biehler, 1971) where the precipitation of silica and carbonates in sediments above the hydrothermal systems results in an increase in density of the sediments and thus a positive gravity anomaly.

The Bouguer gravity map, for the Roosevelt Hot Springs area (Fig. 5), is dominated by two major features: a broad, closed minimum centered over Milford Valley, and a north-trending elongate high over the Mineral Mountains. The data are fully terrain corrected. The gravity relief of approximately 36 mgal reflects a density contrast of 0.3 to 0.7 between valley fill and the bedrock (Table 4) within the range. A generally planar regional gradient of approximately 1 mgal/km decreasing eastward may be estimated for the Mineral Mountains area based on regional data (Crebs and Cook, 1976).

Bouguer gravity decreases gradually from the central part of the range to the valley low. The gravity high along the northern and western parts of the Mineral Mountains can be correlated with the Paleozoic limestones, Precambrian metasediments, and Tertiary diorites. In the vicinity of the geothermal field, denser lithologies such as the Precambrian banded gneiss and Tertiary diorites have been intruded by less dense felsic phases of the Mineral Mountains intrusive complex. Local perturbations on the smooth gradient (residuals), generally less than 2 mgal, can be readily explained by density variations of 0.1 to 0.3 g/cc within the crystalline rocks, as documented in Table 4. The interpretation of structural details beneath the alluvium within and west of the geothermal field is within the range of ambiguity due to uncertainties in the regional gradient and the variable density within the bedrock. A principal result of model studies supported by well control and density data is the absence of a large (>650 ft; 200 m) displacement in the bedrock surface along any single normal fault. Instead we see a gradual dip to the west, and possibly several minor faults near the Opal Mound fault and the outcropping range front. Ward et al (1978) present a similar interpretation, and these interpretations are in agreement with reflection and refraction seismic data presented later.

Magnetic Methods

The aeromagnetic map shown in Figure 6 is part, about 100 mi^2 (270 km^2), of a much larger 300 mi^2 (780 km^2) survey flown in 1975 (Ward et al, 1978). Data were obtained along east-west flight lines with an average line spacing of 1,380 ft (420 m). The flight path was smoothly draped at an average 1,000 ft (305 m) above ground level.

An extensive program of magnetic susceptibility measurements was undertaken in the summer of 1979 to provide support for the magnetic interpretation. The susceptibility data (Table 5) are in-situ susceptibility measurements at more than 60 locations on smooth, unweathered outcrop surfaces. A Bison magnetic susceptibility meter Model 3100 with in-situ coil accessory was used for the measurements. The data shown in Table 5 are fully corrected for outcrop surface

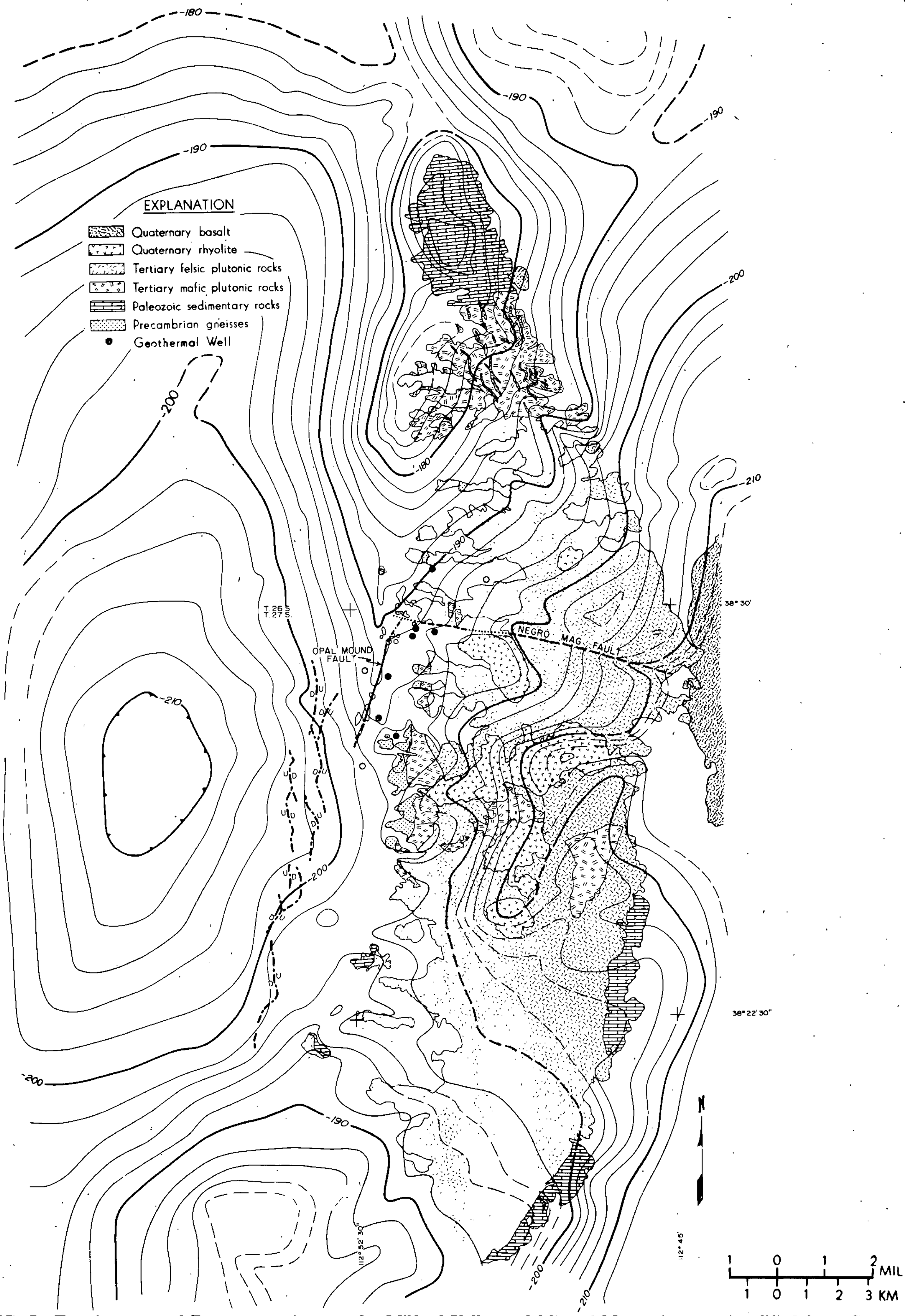

FIG. 5—Terrain-corrected Bouguer gravity map for Milford Valley and Mineral Mountains area (modified from Carter and Cook, 1978).

roughness. Additional susceptibility data determined from drill cuttings of Roosevelt KGRA 9-1 are reported by Glenn et al (1980).

A detailed inspection of the map reveals about 30 closed highs and lows over outcrop of the Mineral Mountains. The apparent complexity of the aeromagnetic map is not surprising in view of the complex igneous geology (Fig. 2) and the unavoidable variations in terrain clearance over rocks of varying magnetization. A study of the analog altimeter profiles shows actual terrain clearance values as little as 480 ft (150 m) over sharp topographic highs and as much as 1,200 ft (365 m) over canyons and between hills. This terrain clearance variation contributes to considerable east-west elongation and irregularities in the contoured map.

An in-depth discussion of the entire magnetic map is beyond the scope of this more general paper. An interpretation has been completed with integrated spatial correlation of magnetic and geologic maps, simple depth estimates, model comparisons, and extensive magnetic susceptibility data. On this basis, most of the magnetic features over the Mineral Mountain range can be attributed to mapped rock-type and altitude variations. Table 6 summarizes the main characteristics and interpreted sources for major anomalies identified in Figure 6. Most important to the present study is the interpretation of new information that relates to the struc-

Table 4. Densities of Lithologies from Mineral Mountains

Rock Type/Unit	g/cc ρ range	g/cc ρ (x)	No. of Samples	Reference
Tqm	2.43 - 2.80	2.64 ± .04	56	Glenn et al (1980)
"granitic"	2.45 - 2.71	2.59 ± .07	25	Carter and Cook (1978)
"granitic"	2.53 - 2.60	2.57 ± .02	11	Crebs and Cook (1976)
Td	2.54 - 2.90	2.76 ± .08	48	Glenn et al (1980)
Ts	2.43 - 2.63	2.55 ± .06	23	Glenn et al (1980)
Qrf/Qrd	2.16 - 2.24	2.22 ± .05	8	Crebs and Cook (1976)
Qrf/Qrd	2.22 - 2.38	2.31 ± .07	8	Carter and Cook (1978)
Obsidian	2.15 - 2.35	2.30 ± .07	8	Carter and Cook (1978)
P€bg (upper plate)	2.11 - 2.95	2.73 ± .28	136	Glenn et al (1980)
Gneiss and schist	2.63 - 2.74	2.69 ± .04	8	Crebs and Cook (1976)
Gneiss	2.63 - 2.74	2.69 ± .04	5	Carter and Cook (1978)
Limestone	2.55 - 2.97	2.71 ± .13	9	Carter and Cook (1978)
Quartzite[1]	2.50 - 2.74	2.62 ± .09	5	Carter and Cook (1978)
Alluvium		2.0 ± .1[2]		Carter and Cook (1978)
		2.05[3]		Glenn and Hulen (1979)

[1]Samples from Star Range, Rocky Range, Beaver Lake Mountains or Beaver Dam Mountains.
[2]Nettleton's method.
[3]Well log determination above water table.

Table 5. Magnetic Susceptibilities of Lithologies, Mineral Mountains, Utah[1]

Rock Type	Symbol#	No. Sites	No. Observations	Mean	Std. Dev.*	Range
				← Units of micro cgs	→	
Alluvium	Qal	2	8	500	86	388-640
Rhyolite flows	Qrf	3	56	179	69	41-347
Rhyolite domes	Qrd	4	43	131	84	0-372
Rhyolite dikes	Trd	1	10	406	29	363-463
Diabase dikes	Tds	1	3	474	—	466-483
Granite dikes	Tgr-d	3	14	411	129	166-650
Granite	Tg	6; tr.	63	1,353	333	623-2,053
Seyenite	Ts	5; tr.	36	1,998	272	1,546-2,056
Porphyritic granite	Tpg	6; tr.	86	1,626	282	10-2,467
Quartz monzonite	Tqm	7; tr.	58	1,918	166	1,433-2,244
Hornblende gneiss	Tgn	4	24	2,310	830	1,520-3,440
Biotite gneiss	Tn	2	6	2,644	405	2,118-3,170
Sillmanite schist	P€s	tr.	10	22	32	0-89
Banded gneiss	P€bg	6; tr.	35	1,252	1,698	38-5,421

[1]Corrected in-situ susceptibilities, 6" diam. coil, Bison system.
Map symbol after Nielson et al, 1978.
*Standard deviation not always statistically significant, but indicative of susceptibility variability.
tr. = indicates observations along traverses, 500 to 2,500 ft long.

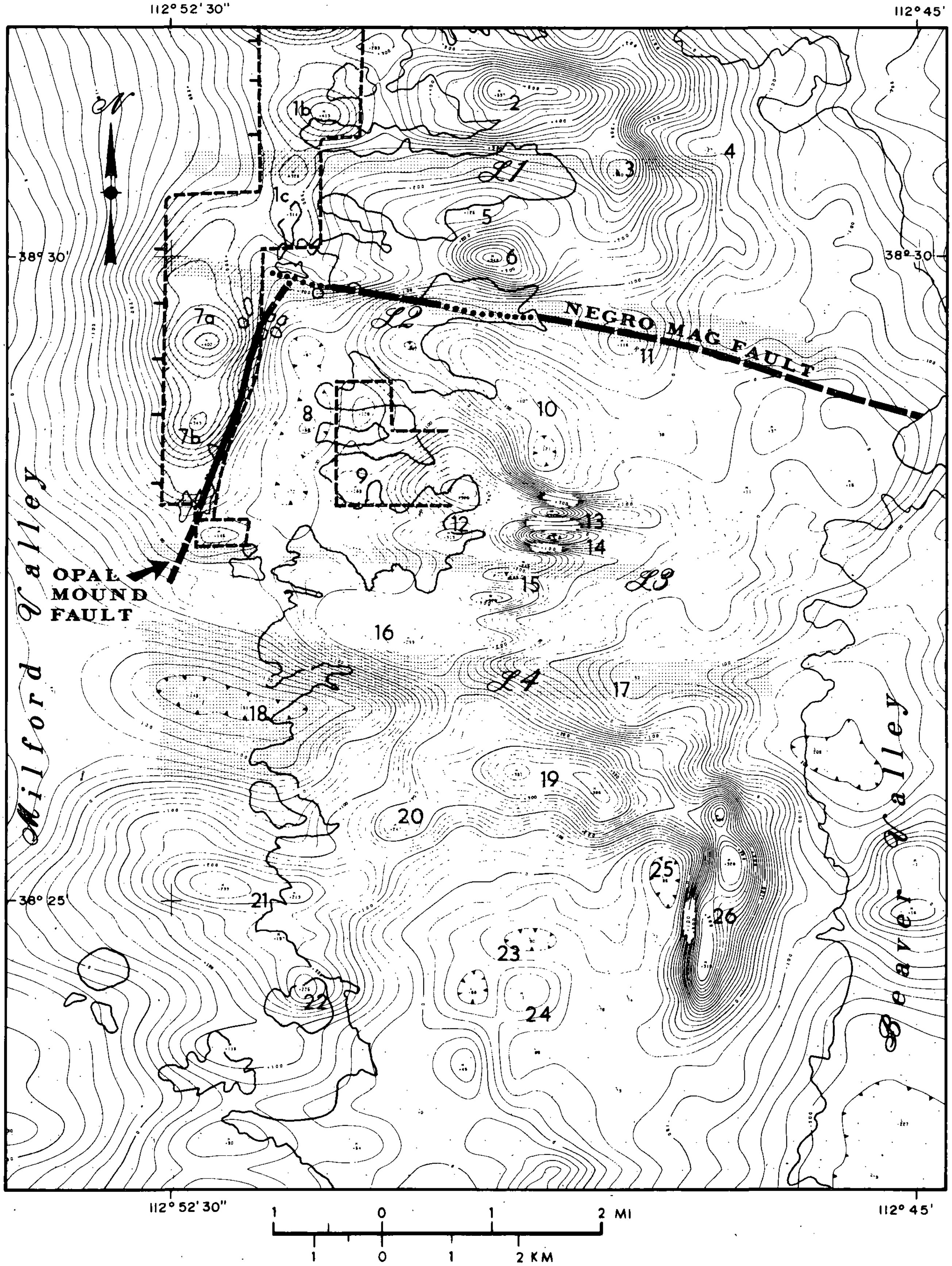

FIG. 6—Total magnetic intensity map of Roosevelt Hot Springs-Mineral Mountains area. Suballuvial magnetic sources and principal structural zones are indicated. Major anomalies are numbered for reference to text.

tural setting of the geothermal reservoir.

Several east-trending zones of low magnetic intensity cut across the Mineral Mountains (L1, L2, L3, L4). These magnetic trends are explained in part by the outcropping rock types and topographic features. The topography, geology, and magnetics are all expressions of east-west structural features. Feature L2 corresponds closely to the Negro Mag fault zone and its eastward continuation as mapped faults. It also appears to displace to the west a north-trending magnetic anomaly (7) closely associated with the Opal Mound fault. L3 terminates anomaly 7 between productive reservoir well Utah State 72-16 and the hot but dry hole, Utah State 52-21. The magnetic bodies appropriate to this magnetic field inclination are indicated on Figure 6.

The magnetic data over bedrock areas reflect the mapped geology and magnetic susceptibility. The identification of four east-trending structural zones, two of which (L1 and L3) may limit the north-south extent of the reservoir, is new information. The delineation of the alluvium-covered Opal Mound horst, bounded on the east by the reservoir-bounding Opal Mound fault, is the most important contribution to understanding the geothermal reservoir provided by the magnetic data. In view of known susceptibility values for the igneous and metamorphic rocks and the alluvium, it is not necessary to postulate an alteration low resulting from magnetite destruction to explain magnetic lows east of the Opal Mound fault (anomaly 8) and south of well Utah State 52-21 (anomaly 18).

Seismic Methods

A broad spectrum of seismic data is available at Roosevelt Hot Springs. Passive seismic data include long-term historical records of major earthquake activity, microearthquake surveys and, at a lower magnitude of naturally occurring seismic disturbance, seismic emissions or "noise" surveys. Single profiles of both refraction and CDP Vibroseis reflection data are also available for study.

In 1974 and 1975, an array of up to 12 portable high-gain seismographs was established within the Roosevelt Hot Springs and Cove Fort–Sulphurdale areas (Olson and Smith, 1976). In two survey periods totaling 49 days, 163 earthquakes of magnitude $-0.5 < M < 2.8$ were recorded. However, only four events could be associated with a 12-mi (20-km) length of the western flank of the Mineral Mountains which includes the Roosevelt Hot Springs area (Ward et al, 1978). Olson and Smith (1976) determined P-wave delays of up to 0.1 sec and detectable S-wave attenuation of ray paths across the Mineral Mountains. Ward et al (1978) interpreted the low-velocity effect and shear wave attenuation for these ray paths as possibly indicative of partial melting or major intense fracturing of crustal rocks beneath the southern part of the Mineral Mountains.

Robinson and Iyer (1981) observed P-wave residuals of up to 0.3 sec in a well-defined pattern corresponding to a region of anomalously low velocity (5 to 7%) centered under the geothermal area and extending from

Table 6. Magnetic Anomaly and Source Characteristics, Roosevelt Hot Springs Area, Utah

Anomaly	Amplitude (gammas)	Correlation[1] ρ topo/mag	Geologic Setting[2]	Probable Source[3]	Susc. Contrast (micro cgs)
1 b, c	120; 100	X	Qal, Qcal, Tqm	Tqm	2,000
2	280	H	Tqm, Tgr	Tqm, Ts; r.t.c.	~ 2,000
3	120+	M	Tqm	Tqm; r.t.c.	
4	− 80	R	Qrd, Qal	Qrd, Qal; r.t.c.	~ −1,400
5	−100	M-X	Qal, Tgr, Ts, Tqm	Qal, Tgr; i.t.c.	
6	190	H	Ts, Tqm	Ts, Tqm; r.t.c.	~ 2,000
7 a, b	300; 240	X	Qal, Qcal		2,000-5,000
8	− 40	X	Qal	Qal; i.t.c.	
9	50	H	Tgn, Ts, Tpg	Tgn, Ts; r.t.c.	
10	−200	L-R	Qrt, Qal	Qrt (reversed)	~ −1,500
11	50	H	Tqm	Tqm; r.t.c.	
12	190; 200	L-X	Ts, Qal	Ts;	~ 2,000
13*	250	H	Tqm	Tqm; * r.t.c.	
14*	−180	R?	Tqm	Tqm; * i.t.c.	
15*	− 80	R	Qrd	Qrd; * r.t.c.	
16	200	H-M	Tpg, Tn, Tgn, Ts, Tg	Ts, Tpg, Tgn; r.t.c.	~ 2,000
17	−150	X-R	Qrd, Tg	Qrd; i.t.c.	
18	− 90	L-X	Qal, Pєbg, Tpg	Qal; Pєbg, Tg	
19	350	H	Tg, Tn, Tpg, Qra, Qrd	Tn, Tg, Tpg; r.t.c.	~ 2,500
20	250	H	Tn, Tpg, Tgn	Tn, Tpg, Tgn; r.t.c.	~ 2,000
21	250	X	Qal, Tn, Tgn, Tpg	Tpg, Tgn, Tn	~ 2,000
22	150	L	Qal, Tbg, Tgr	Tbg, Tgr	
23	− 30	L-R	Qra, Qrd	Qra; i.t.c.	
24	20	H	Qrd	Qrd; r.t.c.	
25	− 50	H-R	Tq	— r.t.c.	
26	350	H	Tdb, Tg, Tbg	Tdb; r.t.c.	~ 3,500

[1]H = high, M = moderate, L = low, X = insignificant, R = reversed.
[2]Map symbol after Nielson et al (1978).
[3]r.t.c. = reduced terrain clearance; i.t.c. = increased terrain clearance; * indicates poor contour representation.

about 3 mi (5 km) depth down into the uppermost mantle. They preferred an explanation for these delays in terms of abnormally high temperatures and a small fraction of partial melt. Wechsler and Smith (1979) noted that these delays could arise from the fractured and possibly fluid-filled porosity of the western part of the Mineral Mountains pluton. Wechsler and Smith further noted the problems of accurate epicenter location and the limitations of P-wave studies in areas of complex near-surface lateral-velocity variations, such as exist at Roosevelt Hot Springs.

The relatively few earthquake locations determined for the western Mineral Mountains from this 49-day recording period and the complex near-surface velocity structure almost preclude the use of microearthquakes in delineating the Roosevelt Hot Springs geothermal system. P-wave delay studies may ultimately improve our understanding of the deep heat source at Roosevelt, but to this time have not contributed to delineation of the system.

Schaff (1981) has reported on seismic activity detected with the present nine-station seismograph array for the period October 1979 through January 1981. His results to date substantiate the earlier conclusion that seismicity in the immediate vicinity of Roosevelt Hot Springs is generally of a low level and somewhat episodic in nature. In the first 12-month period, no earthquakes were located within the anticipated production zone or along the Opal Mound fault. However, swarmlike activity was recorded east of the reservoir area in December 1980–January 1981. This trend of earthquake epicenters across the Mineral Mountains and east of the geothermal reservoir appears to lie along the eastern projection of the Negro Mag fault. More than 1,000 earthquakes of magnitude less than 1.5 have occurred in three major swarms between July and December 1981. The epicenters for these most recent events define a trend extending from the surface to 4 mi (6 km) depth parallel with the trend of the Negro Mag fault (L. McPherson and G. Zandt, personal commun.).

Seismic emissions survey.—Seismic emissions surveys have been promoted by several geophysical contractors as a geothermal exploration method in which the seismic emissions or "noise" would hopefully delineate active fault and fracture zones possibly associated with geothermal activity. The method employs an array of geophones (four or five) spaced approximately 2,000 ft (610 m) apart. In surveys at Roosevelt Hot Springs (Katz, 1977a, b) data were recorded at the array for 1 to 3 days, then moved to another station. Five such stations within a 14 to 16 mi^2 (36 to 41 km^2) area constitute a survey. The data were edited and processed with algorithms which determine the noise source locations based on delay times computed for a half-space velocity model and the correlation of these delays with the observed data.

A reevaluation of these data has been completed by Ross et al (1982). As employed at Roosevelt Hot Springs, the seismic emissions survey may indicate areas of geothermally induced seismic noise, but it is dominated by other noise sources and is imprecise in defining geothermal conduits. The correlation procedure is severely limited by model simplicity and velocity assumptions, and generally recognizes source direction more accurately than distance to the seismic noise source. A more refined velocity model could perhaps improve the resolution of the noise source areas through a higher correlation of source-to-geophone array delay times. It is unlikely that the velocity model could be refined enough to justify inclusion of the method in geothermal exploration in complex geologic environments.

Seismic refraction.—In April 1977, a 19 mi (30-km) long seismic refraction profile was recorded across the Roosevelt Hot Springs geothermal area (Gertson and Smith, 1979). Multiple shots at seven different shot locations were used to provide multiple subsurface coverage. Although the large station spacings of nearly 820 ft (250 m) were not adequate for locating narrow structural features, Gertson and Smith were able to define a somewhat generalized velocity model for the area and also determined that the first large displacement in range front faulting occurs at least 0.6 mi (1 km) west of the Opal Mound fault. P-wave attenuation across the geothermal reservoir was much less than attenuation in other parts of the profile, and Gertson and Smith (1979) concluded that the record sections did not appear to contain any evidence for seismic waves that had penetrated and returned from "hot rock" or magma chambers even at great depth. One complicating aspect of the refraction study is that the eastern part of the refraction profile, of necessity, followed the Negro Mag fault zone from the reservoir area eastward across the Mineral Mountains.

Seismic reflection.—One profile totaling 27 mi (43 line km) of detailed reflection seismic data was available for study of the Roosevelt Hot Springs area. Line 5 and 5 OPTW of a GSI speculation survey, recorded in March and April 1978, crosses the Mineral Mountains along the same path as the refraction survey. The data cannot be reproduced, but an interpretation of the data is presented here.

The GSI profiles are high-quality 24-fold CDP Vibroseis data with a 200 ft (60 m) group interval and a 400 ft (120 m) vibrator point interval. The source consisted of 16 sweeps per VP, using a 12-sec sweep and a 12 to 60 Hz sweep band. The sample rate was 2 msec. Processing included deconvolution, velocity analysis, CDP stack, and migration.

Our interpretation of the reflection profile is presented in Figure 7. The time-to-depth conversion is supported by 14 velocity analyses along the profile. These velocities are generalized in Figure 7. The refraction survey and sonic logs of three deep wells in the geothermal system—Utah State 52-21, 72-16, and 14-2—also provide velocity control. Although similar velocities are noted for alluvium, the refraction survey and well logs indicate much higher velocities at depth than velocity analyses of CDP reflection data. To those depths where good reflection quality persists and the coherence of the velocity analysis is well supported, the reflection survey velocities are considered more valid. At greater depths (times), the velocity analyses appear to be less valid and may be as much as 40% too low.

Figure 7 shows gently dipping layers in unconsolidated sediments to depths of 4,000 ft (1,220 m) in

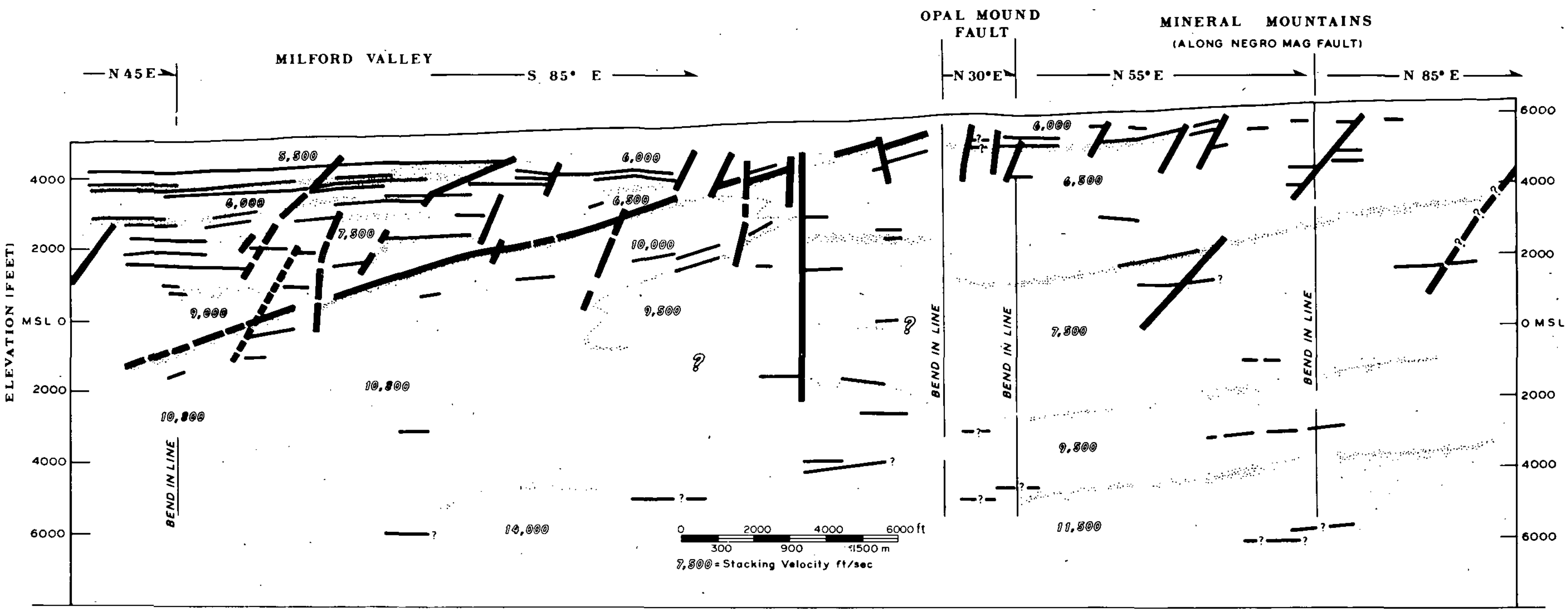

FIG. 7—Structural interpretation of Vibroseis profile GSI-5. Units as determined from velocity analysis and reflection continuity.

the Milford Valley. This sequence thins abruptly to the east as the Opal Mound fault is approached. Numerous faults cut the sedimentary sequence. A very prominent reflector which dips westward at approximately 20° from the Opal Mound fault indicates the base of the unconsolidated sediments. Few coherent reflectors are noted beneath this interface, presumably a sediment-igneous or sediment-metamorphic contrast.

The complexity of Basin and Range faulting is indicated by the reflection data. Unfortunately, the trend of the profile is along the southern edge of the horst block indicated by magnetic data, then N30°E along, and at a small angle to, the Opal Mound fault. The faulting is indicated, but less clearly than would be the case for an east-west line. East of the Opal Mound fault, the higher velocities and lack of coherent reflections indicate igneous rocks extending to depth. The interpretation of the vertical displacement along Basin and Range faults within the first 0.25 seconds is difficult because of high noise levels and probable lateral energy returns associated with the Opal Mound fault area. No single, major vertical displacement is indicated but a complex series of small displacements, of the order of 100 to 320 ft (30 to 100 m), is suggested by the data.

Electrical Studies

The recognition of Roosevelt Hot Springs as a relatively shallow geothermal resource of commercial potential has resulted in numerous electrical surveys designed to characterize the electrical resistivity distribution, and to test the effectiveness of various methods. To a large degree, Roosevelt Hot Springs has become a "test" area because of the type of resource and amount of publicly available supporting data. Included in the electrical surveys already completed are the following: magnetotelluric (MT), controlled source audio magnetotelluric (CSAMT), natural source AMT, controlled source EM, spontaneous polarization or self-potential (SP), induced polarization (IP), and electrical resistivity.

Electrical resistivity surveys.—Electrical resistivity surveys include both the dipole-dipole and bipole-dipole arrays. The bipole-dipole survey (Frangos and Ward, 1980) was undertaken with the knowledge of the resistivity distribution primarily to characterize the effectiveness of the method. The dipole-dipole survey includes at least 50 lines of various lengths, electrode separations, and depth penetrations, and constitutes one of the most complete resistivity data bases assembled. In view of this, and because all the other electrical methods are influenced by the resistivity distribution mapped by these data, a major effort was devoted to interpreting these data using two-dimensional numerical models (Ross et al, 1982).

Figure 8 shows the location of the resistivity lines which were numerically modeled and presents a summary of the intrinsic resistivity distribution for the depth interval of 330 to 500 ft (100 to 150 m). The modeled resistivity data are supplemented with a qualitative interpretation of more than 30 additional lines of data. The modeled resistivities at these depths have been transferred from the interpreted resistivity sections shown in Figure 9.

The resistivity distribution, even as generalized by numerical modeling, is quite complex. Generally high resistivities (100 to 500 ohm-m) are observed in the range over the Precambrian gneiss and Tertiary intrusive rocks. Very low (<10 ohm-m) to low (10 to 20 ohm-m) resistivities, often modeled as thin vertical conduits, occur along the trend of the Opal Mound fault. West of the Opal Mound fault moderate to high-resistivity (30 to 400 ohm-m) layers overlie layers of moderate to very low resistivity. This appears to represent dry alluvium (above the water table) overlying areas where fresh ground waters mix with warmer, more conductive geothermal fluids migrating downdip from the Opal Mound fault and other conduits. Very low resistivities (2.5 ohm-m) observed along UU line 8100N may represent geothermal outflow, dissolved salts in Lake Bonneville sediments, or some mixture of both. South of the survey base line and Utah State 52-21, high resistivities near the surface decrease with depth but do not indicate the presence of geothermal fluids noted to the north. Figure 8 provides a better comparison with key geologic features and drill-hole locations than does the section representation (Fig. 9), and shows that the low-resistivity area of the geothermal system is bounded by high resistivities southwest, south, and east of the geothermal system at these depths. Low and moderate resistivities extend well into the range along the Negro Mag fault zone. Most of the low-resistivity areas straddle or lie east of the Opal Mound fault at these depths. Resistivities of 3 to 10 ohm-m along the fault are bounded by 10 to 20 ohm-m on all sides as a crude zoning pattern. The resistivity of the alluvium is generally 20 to 50 ohm-m to the west, and much higher (200 to 400 ohm-m) to the south. The probable continuity of resistivity zones is indicated by heavy dashed lines (Fig. 8).

Ross et al (1982) present a similar map for the depth interval 1,500 to 2,000 ft (450 to 600 m) which is less complex because only 1,000 ft (300 m) dipole data could be used for this depth and because of the resistivity averaging inherent in modeling large separation data. The high resistivities east of the Opal Mound fault extend to the west but are substantially reduced to the south (GOC lines 1, 2, 4, 5). In fact, 50 to 400 ohm-m alluvium becomes quite conductive (5 to 20 ohm-m) with the increased depth. The Negro Mag Wash area is still a moderate resistivity reentrant into the range. The alluvium west of the Opal Mound fault has become quite low in resistivity (5 to 15 ohm-m), probably due to the downdip migration of conductive thermal fluids leaking from the geothermal system. The region of the Opal Mound fault itself is modeled as 30 to 50 ohm-m except at the southernmost part of the Opal Mound itself (8 ohm-m). This is in marked contrast to the low resistivity expression at 330 to 500 ft (100 to 150 m) depth. A high resistivity (450 ohm-m) body is indicated less than 1,250 ft (380 m) south of successful well Utah State 72-16, which reached total depth at 1,256 ft (383 m). The hot but dry Getty Oil Co. well, Utah State 52-21, is located in a resistive (100 to 400 ohm-m) area. Several productive wells (Roosevelt KGRA 12-35, Roosevelt KGRA 54-3, Roosevelt KGRA 3-1, Utah State 14-2, Roosevelt KGRA 13-10, and Roosevelt HSU

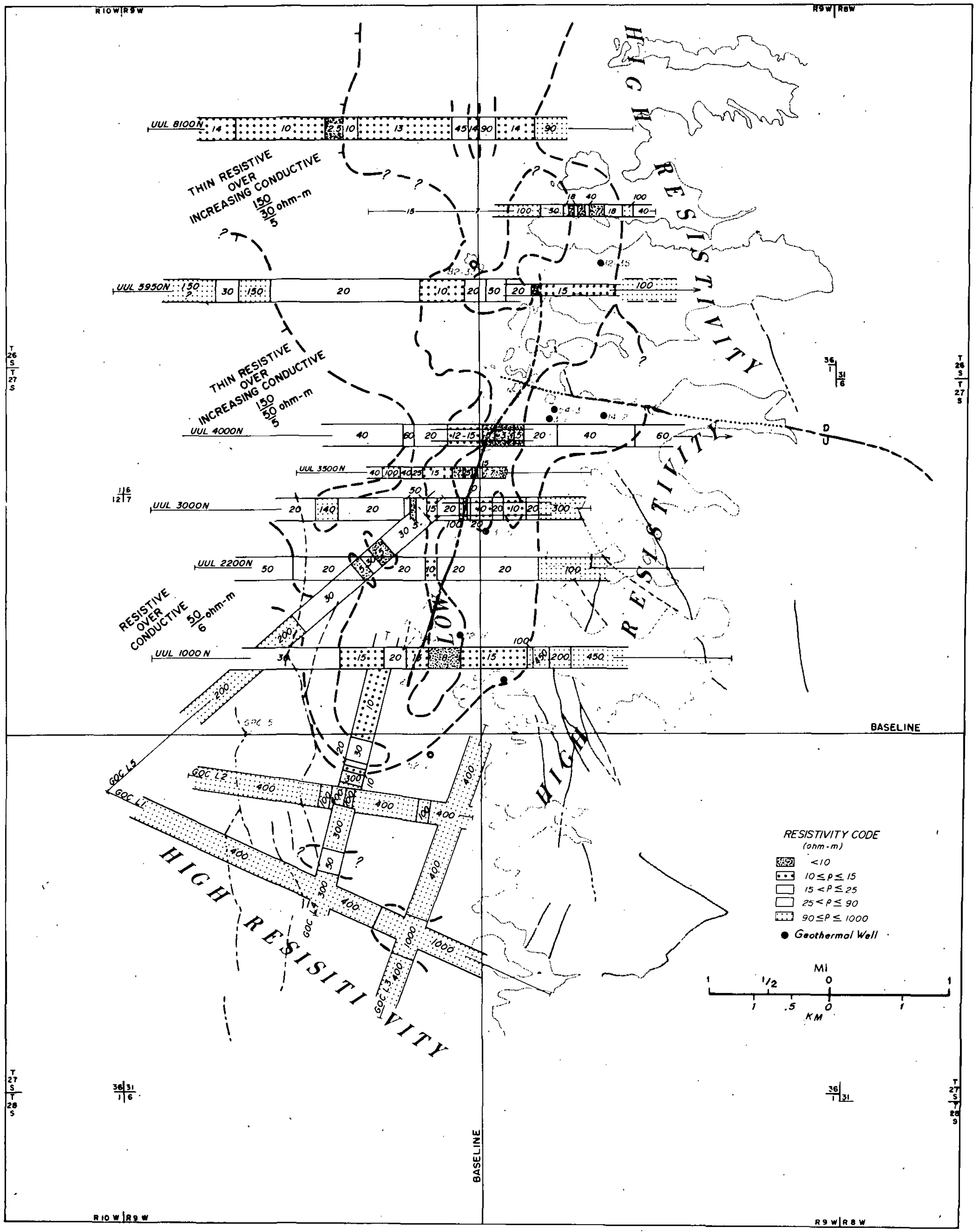

FIG. 8—Interpreted electrical resistivity distribution for depth interval 100 to 150 m, Roosevelt Hot Springs KGRA.

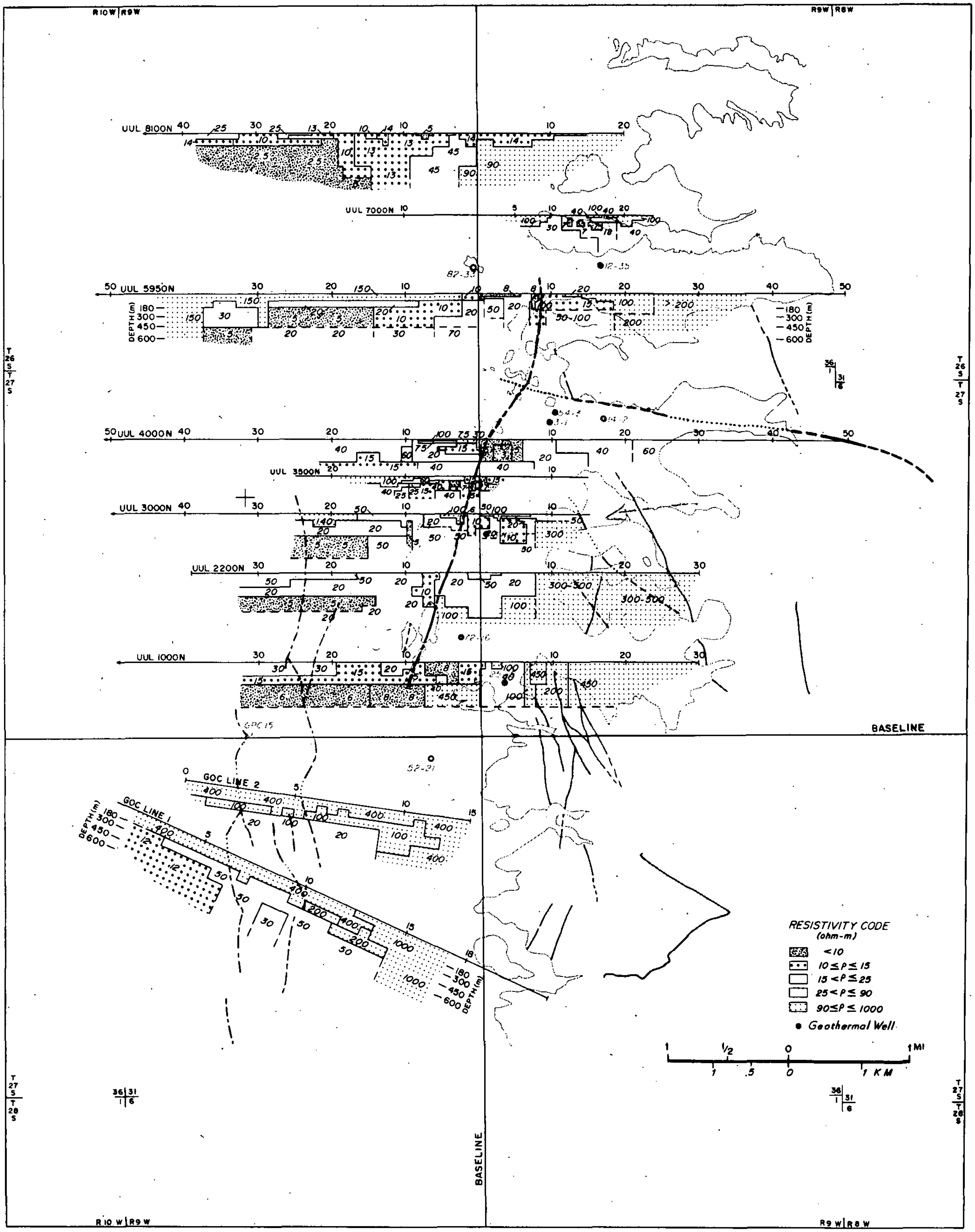

FIG. 9—Interpreted electrical resistivity sections, Roosevelt Hot Springs KGRA.

25-15) are sited in areas of moderate (15 to 40 ohm-m) resistivity which are relative lows.

The interpretations presented in Figure 9 are non-unique and are limited by the grid size which becomes coarser with depth, the validity of the two-dimensional model, the goodness of fit of computed to observed data values, and the choice of body size, position, and resistivity. Nonetheless, careful modeling of dipole-dipole resistivity data has resulted in a more accurate representation of earth resistivity distribution for a given cost than any other electrical method. The non-uniqueness is reduced by using a network of profiles, several of which intersect, and the integration of geologic data, such as the detailed (1:24,000) map of Nielson et al (1978).

Near-surface electrical methods.—Sandberg and Hohmann (1980) described a controlled source audio-magnetotelluric (CSAMT) survey at Roosevelt Hot Springs which compares well with dipole-dipole resistivity data for shallow depths ($\leq$ 330 ft; 100 m). Minor disagreements in the interpretative models from these methods serve to indicate the noise and ambiguity levels for each method. Chu et al (1980) described induced polarization surveys which were unable to map clay alteration or pyritic zones related to the geothermal system.

Corwin and Hoover (1979) discussed thermoelectric coupling and electrokinetic coupling as possible mechanisms for generating self-potential anomalies observed over geothermal systems. They reported a dipolar anomaly located directly over the Opal Mound fault and concluded that it results from geothermally generated electrical activity along the fault. More detailed self-potential studies have been completed by Sill and Johng (1979). A complex "quadrapolar" anomaly of two lows (-100 mv) and two highs (+25 and +50 mv) was found to be associated with the southern part of the geothermal system. The 100-mv low which occurs over the Opal Mound fault is the most unambiguous expression of the geothermal system. Other self-potential features could arise from large resistivity contrasts, the movement of ground water from higher to lower elevation, or fluid movement along structures. Additional numerical modeling is required to obtain a more complete understanding of the SP response at Roosevelt Hot Springs, but the observed data document a complex expression of the geothermal system.

Magnetotelluric (MT) studies.—The magnetotelluric (MT) method is often used in both the reconnaissance and detailed stages of geothermal exploration. The earth's electric and magnetic fields vary as a function of frequency in response to natural electrical (telluric) currents flowing within the earth's crust. Through precise measurements of the electric and magnetic field components made at the surface, one may obtain information relating to the impedance distribution (i.e., electrical resistivity) to depths as great as 25 mi (40 km) within the earth's crust. The reader is referred to an excellent paper by Vozoff (1972) for a detailed description of the method.

In keeping with the detailed geophysical definition of the Roosevelt Hot Springs area an extensive MT network of 93 stations was established in the Milford Valley–Mineral Mountains area (Wannamaker et al, 1978, 1980).

The reduced data for a central 14 mi (22 km) east-west profile are presented as observed apparent resistivity versus frequency for the TM (transverse magnetic) mode in Figure 10a. The best-fit model results computed for a two-dimensional model geometry TM mode (Fig. 10b) are seen to be very similar. The comparison between observed and modeled impedance phase (not shown here) is also good but has a poorer fit in the central part of the profile. The two-dimensional model that produced this best fit is shown as Figure 10c. The rather fanciful 50,000 ohm-m prism in Figure 10c was inserted in an unsuccessful attempt to match the mid-frequency (10 to 0.1 Hz) data. A frequency dependent, three-dimensional current gathering effect involving the Milford Valley to the west is now believed responsible for these particular modeling difficulties (Wannamaker et al, 1980).

Stations 78-22 through 76-2 are located over low-resistivity alluvium and valley fill. Figure 10c indicates resistivities decreasing to near 1 ohm-m at depths of 1,300 ft (400 m). Dipole-dipole resistivity data and Schlumberger soundings also indicate a thin (330 ft; 100 m) resistive (100 ohm-m) near-surface layer with resistivity decreasing to 5 ohm-m or less at depths of 1,000 ft (300 m). Station 76-3 near the center of the cross section occurs over a buried horst inferred from gravity and magnetic data and exhibits high resistivities at moderate depths. Station 76-4 was sited along the Opal Mound fault (Wannamaker et al, 1978), which has been identified earlier as a narrow, vertical conductive zone associated with ascending geothermal fluids. Silica cementation has locally decreased the porosity but fracture permeability remains.

Wannamaker et al (1978, 1980) noted great difficulty in obtaining good representations of both the TE and TM mode resistivities for any single one-dimensional or two-dimensional model. Extensive three-dimensional model studies indicate the limitations of one- and two-dimensional modeling at Roosevelt Hot Springs, and probably for most Basin and Range-type geothermal reservoir areas. A few of their more general conclusions are restated here.

1. Current gathering in the valley results in a regional distortion of the electric field affecting all stations at Roosevelt Hot Springs for lower frequencies.
2. The TM mode is most appropriate for two-dimensional interpretation, and has yielded good results for geometrically regular three-dimensional prisms.
3. Clays (ϱ = 1 to 2 ohm-m) may exist to depths of several hundred meters in the Milford Valley. These overlie more than 0.6 mi (1 km) of semiconsolidated and unconsolidated sediments and volcanics of moderate ($\varrho \simeq$25 ohm-m) resistivity.
4. A geometrically regular reservoir of conductive brine beneath the thermal anomaly seems improbable, so the search for any economic hydrothermal reservoir at Roosevelt Hot Springs using MT must be considered unsuccessful at this time. If present, it is not resolved by the two-dimensional TM algorithm. The brine-saturated reservoir zone is clearly three-dimensional and difficult to model satisfactorily with present interpretation

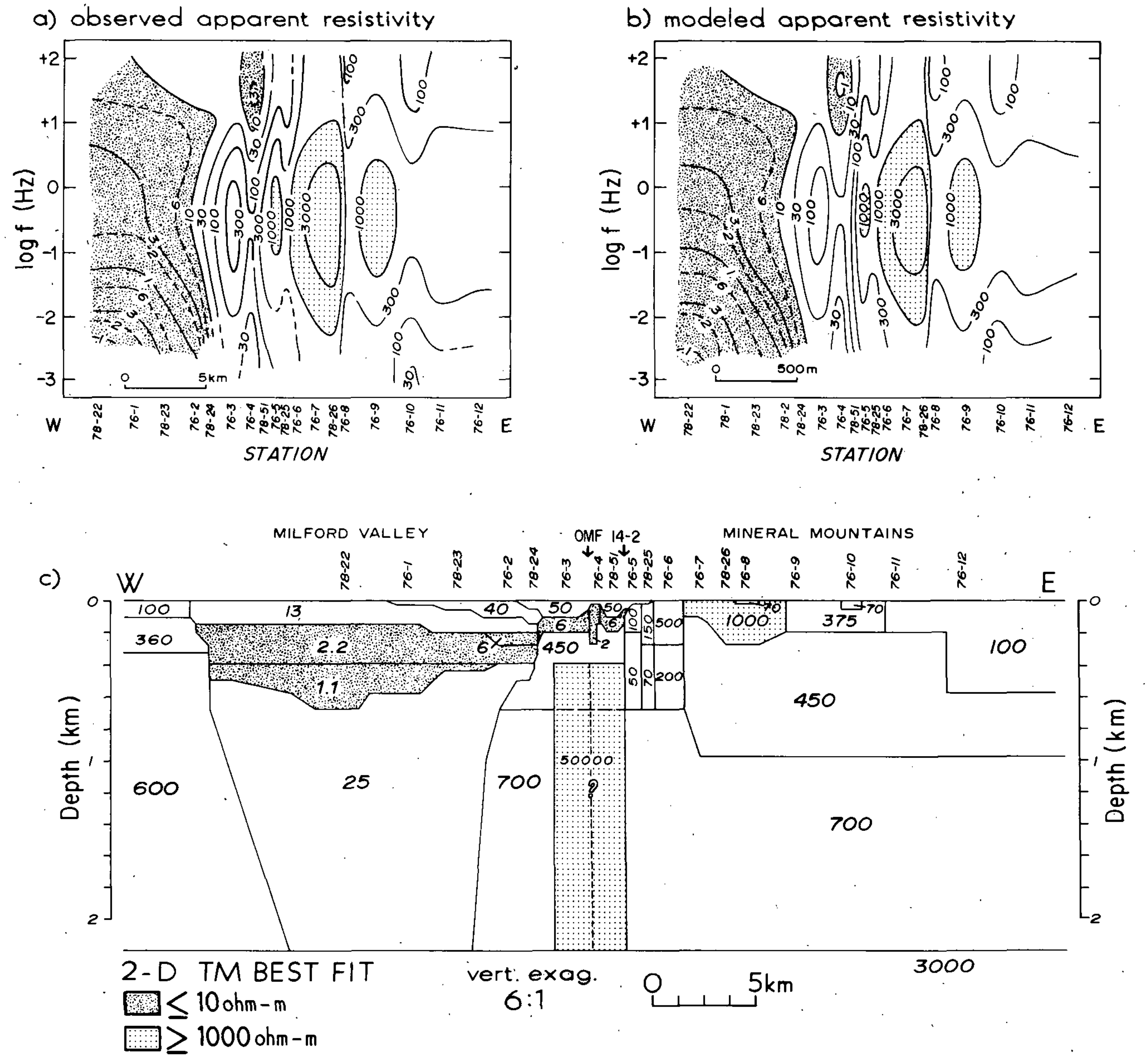

FIG. 10—Magnetotelluric profile CC′: (a) observed apparent resistivity, TM mode; (b) modeled apparent resistivity, TM mode; (c) finite element model for profile CC′ giving best fit between computed and observed TM mode resistivities. Intrinsic resistivity values in ohm-m; OMF = Opal Mound fault. (After Wannamaker et al, 1980.)

capabilities.

5. A deep heat source for the geothermal system also has not been discerned by MT interpretation to this time.

Thermal Studies

The thermal characteristics of the Roosevelt Hot Springs area have been defined by measurements in 53 thermal gradient, water well, and exploration drill holes (Wilson and Chapman, 1980). Most of the 53 gradient holes bottomed at depths of 200 to 360 ft (60-110 m). The observed near-surface (30 to 230 ft; 10 to 70 m depth) gradients range from 6°C/km to 3,330°C/km compared to a Great Basin average of 35 to 40°C/km.

Thermal conductivity measurements are listed in Wilson and Chapman (1980) and Glenn et al (1980). Wilson and Chapman (1980) assign the following average conductivities (W/m/K) to principal rock types: quartz monzonite, 2.54; opaline sinter, 2.00; biotite gneiss, 2.00; alluvium, 1.64. Using the appropriate thermal conductivities and gradients for each hole, Wilson and Chapman have produced a detailed map of the near-surface heat flow associated with the geothermal system. The 400 mW/m^2 and 1,000 mW/m^2 areas determined from their study are indicated in Figure 11. The 400 mW/m^2 contour, approximately four times background, encloses an area of 22 mi^2 (57 km^2), whereas the 1,000 mW/m^2 contour encloses an area of 6 mi^2 (16 km^2) including the Opal Mound fault and most of the successful production drill holes. The anomalous surface heat loss for the system was calculated at 64 MW.

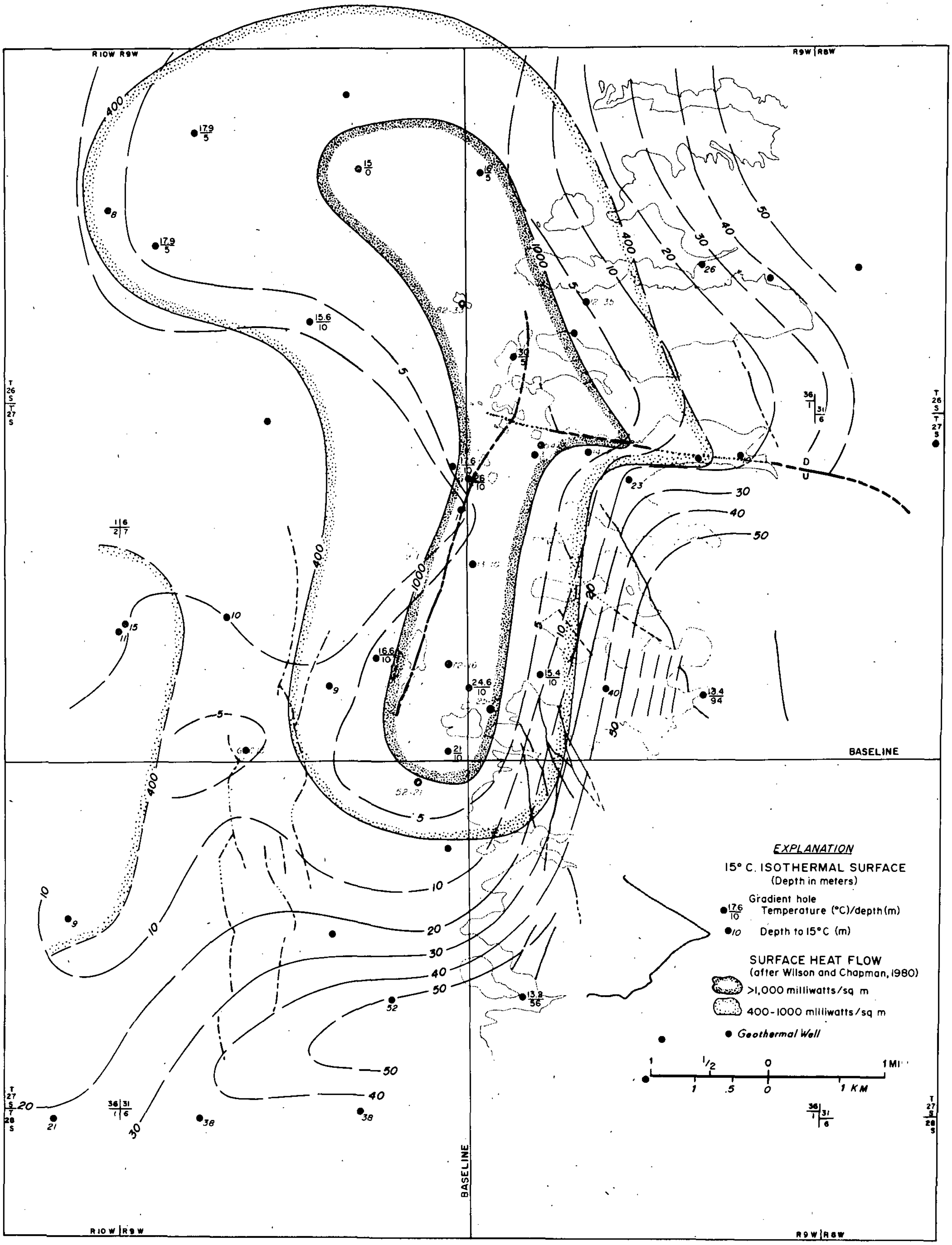

FIG. 11—Thermal studies map, Roosevelt Hot Springs KGRA.

Figure 11 compares the depth of the 15°C isothermal surface with the heat flow pattern of Wilson and Chapman (1980). Temperatures at a given depth are included for several holes. The 1,000 mW/m^2 area corresponds closely to the area where 15°C is mapped within 15 ft (5 m) of the surface. The steep contours along the flank of the range indicate depressed temperatures because of descending ground water, and a broad contour interval to the west and elongation to the northwest probably relate to the discharge of geothermal water within permeable strata in the alluvium. Although the contoured heat-flow map presents the thermal data in the most quantitative manner, the shallow occurrence of the high-temperature reservoir and its active discharge into the alluvium give rise to good definition of the system by contouring most thermal parameters.

RETROSPECTIVE EXPLORATION STRATEGY

The Roosevelt Hot Springs geothermal system has been used as a natural laboratory for the testing of exploration techniques. Using this relatively complete data base, it is possible to exercise 20/20 hindsight and define the most efficient strategy for the exploration of the Roosevelt Hot Springs geothermal system. Ward et al (1981) present a generalized exploration strategy which emphasizes the use of conceptual exploration models to solve specific exploration problems. The retrospective exploration strategy for Roosevelt Hot Springs involves five stages: literature study, geologic mapping, thermal gradient measurements, dipole-dipole resistivity, and drilling. The following strategy may not be applicable to other systems because the Roosevelt Hot Springs system is one of the hottest and largest hydrothermal systems in the Basin and Range province.

Phase I.—The first stage of the strategy would be a literature search and compilation. Review of existing geologic and hydrologic reports would have demonstrated that hot springs had been active in the recent past and would have provided chemical analyses of those waters. From those analyses, geothermometers could have been calculated to give an indication of possible reservoir temperatures. In addition, the presence of siliceous sinter is considered as evidence of a water-dominated geothermal system which has a base temperature of at least 356°F (180°C) (Renner et al, 1975). The presence of Quaternary rhyolites suggests the possibility of a high-level granitic pluton which supplies heat to the geothermal system. A review of the ground-water reports would allow plotting of potentiometric surfaces and chemical trends such as shown in Figure 3.

The approximate cost of such a literature search and compilation could vary widely depending on the expertise involved and the size of the region under consideration. The costs are thought to be between $5,000 and $20,000. Because the surface manifestations of the Roosevelt Hot Springs system are so well developed, a decision to initiate leasing activities could be made at the end of phase I.

Phase II.—This stage would involve the geologic mapping of the project area. Nielson et al (1978) found that mapping at a scale of 1:24,000 on an air-photo base was adequate for exploration purposes. The mapping should identify the extent of the hot-spring deposits and characterize their structural control. It would also identify the other structural systems within the geothermal area and allow a preliminary guess at the relative importance of the various fault systems in controlling the geothermal reservoir. Previous attempts at using air-photo interpretation alone to define these structural systems (Ward et al, 1978) were relatively unsuccessful.

The knowledge gained during this phase should be used to establish conceptual models of the geothermal system. In addition, the results will provide siting information for the subsequent thermal-gradient holes and electrical resistivity surveys.

The costs of the geologic mapping will probably be $15,000 to $25,000, one of the most cost-effective exploration options available.

Phase III.—The principal activities of phase III would be the delineation of the near-surface thermal anomaly and mapping of deeper resistivity structure. An area of 8 mi (12 km) north-south by 4 mi (6 km) east-west would be sampled by 25 to 30 shallow thermal-gradient holes drilled to depths of 300 to 600 ft (100 to 200 m). The deeper holes would logically be sited west of the Opal Mound fault where alluvium is predictably deeper and problems of near-surface water flow could be anticipated. This program would have detected temperature gradients in excess of 44°F/100 ft (800°C/km), and localized a high-temperature anomaly within 1.2 mi (2 km) of the Opal Mound fault. The cost of the survey, at 1980 prices, would have been between $130,000 and $160,000.

A slightly larger area, 10 by 5 mi (15 by 8 km), would have been selected for a dipole-dipole electrical resistivity survey. Cost-effective survey design, based on geologic mapping and in-progress temperature-gradient studies, would dictate perhaps 10 dipole-dipole lines of 1.8 to 3 mi (3 to 5 km) lengths, with perhaps 7 lines of 984-ft (300-m) dipole length and 3 lines of 492-ft (150-m) dipole length. The survey costs are estimated at $40,000 to $50,000, with selected numerical modeling estimated at an additional $10,000.

Without the prior knowledge of the valley-range transition, it would have been logical to plan several detailed gravity profiles in an attempt to define a range-front fault with major vertical offset. Such an effort would have cost $3,000 to $5,000, and would have contributed little to target definition.

It would be prudent to collect fluid samples from the gradient holes and to complete chemical analysis and geothermometric calculations on these samples. These costs would have added $1,000 to $2,000 to the exploration bill and provided valuable information on water type, mixing, and preliminary estimates of reservoir temperatures.

Phase IV.—After careful consideration of the geologic, thermal, and electrical data bases, the prudent exploration program would have sited two to four deep thermal-gradient holes, to depths of 1,000 to 2,000 ft (300 to 600 m). These could be expected to cost $15 to $18 per foot ($50 to $60 per meter). At Roosevelt Hot Springs these holes, if well chosen, might have intersected a major fracture at shallow depth (as in Utah State 72-16), or would at least indicate temperatures approaching 300°F (150°C) within 2,000 ft (600 m) depth. The program cost, with fluid collection and lithologic studies, is estimated at $100,000 to $200,000. A substantial effort in data interpretation and data integration would follow as a result of the strong encouragement for a high-temperature system. On the basis of this data analysis, production holes would be drilled.

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Western Exploration and Production Division

December 13, 1978

Mr. Howard P. Ross
University of Utah Research Institute
Earth Science Laboratory
391 Chipeta Way - Research Park
Salt Lake City, Utah 84108

Re: Directional Computation From Dip Log on Getty Oil Company's #52-21 Well at Roosevelt Hot Springs, Beaver County, Utah

Dear Howard:

Please find enclosed with this letter a hand computed directional log from Schlumberger. I am sorry for the delay in transmitting this data, but it was misplaced inadvertently and just re-surfaced.

An additional set of water analysis from flow test samples and a temperature log run in October will be forthcoming to you, which will be the final data from this well.

May I take this opportunity to wish you, your family, and your staff a very happy Holiday Season.

Sincerely yours,

GETTY OIL COMPANY

Wayne A. Shaw

Wayne A. Shaw

WAS/blb

Enclosures

Assumes start depth 2041 x = 0 Y = 0 Z = 0

Hole Azimuth corrected for magnetic declination

DEPTH	DEVIATION	HOLE AZIMUTH	COURSE LENGTH	X South	X West	Z	
2041	5.25	175	0	0	0	0	2041
2084	6	255	43	.34	3.92	42.81	
2114	5.5	255	30	2.69	4.73	72.65	
2132	5.5	255	18	4.36	5.18	90.57	
2186	5.75	255	54	9.36	6.52	144.32	
2216	5.5	255	30	12.26	7.30	174.17	
2246	5.75	245	30	15.04	8.04	204.03	
2276	6	245	30	17.76	7.31	233.88	
2306	6	250	30	20.60	10.64	263.72	
2336	6.25	255	30	23.55	11.71	293.56	
2367	5.5	255	31	26.81	12.58	324.38	
2397	6.25	255	30	29.59	13.32	354.24	
2426	6.5	250	29	32.64	14.14	383.01	
2456	6.5	250	30	35.83	15.30	412.88	
2516	6.5	240	60	42.21	17.62	472.49	2514
2546	7	240	30	45.15	17.32	502.30	
2575	7	240	29	48.21	21.09	531.08	
2605	7	235	30	51.38	22.92	560.86	
2635	7	245	30	54.37	25.02	590.64	
2665	7.75	245	30	57.68	26.57	620.42	
2695	8	240	30	61.35	28.28	650.15	
2725	8	240	30	64.85	30.30	679.86	
2755	8	240	30	68.35	32.32	709.57	
2785	8	240	30	71.85	34.34	739.28	
2846	8.25	235	61	79.28	38.58	799.69	
2967	8.5	240	21	81.75	40.31	820.47	2862
3087	9	240	120	97.11	49.18	939.15	
3207	9.25	240	120	113.37	58.57	1057.67	
3327	10.25	240	120	130.07	68.21	1176.11	
3449	10.75	240	122	148.87	79.06	1296.16	3347
3567	10	235	118	167.93	90.06	1412.09	
3728	10.5	235	159	190.55	105.90	1568.67	
3848	10.5	235	120	208.46	118.44	1686.66	
3999	10.5	230	151	231.00	134.22	1835.13	3876
4171	10.5	230	172	255.01	154.37	2004.25	
4241	11	230	70	264.78	162.57	2073.08	
4392	11	230	151	286.55	181.09	2221.31	
4512	11.25	225	120	304.39	195.81	2339.11	4380
4634	11.75	225	122	321.22	212.64	2458.77	
4751	11	220	117	345.05	229.49	2573.32	
4877	11	220	126	360.50	247.91	2697.01	
5002	11.5	210	125	375.83	266.18	2819.71	4861
5124	12	210	122	387.99	287.24	2939.26	
5246	12.5	220	122	400.67	309.21	3058.59	5409
5430	12.5	215	184	426.27	339.72	3238.23	
5553	13	210	133	442.78	363.30	3368.08	—
5676	14	205	123	456.61	387.26	3487.93	
6238	18	205	562	514.07	510.48	4033.24	
6252	18	200	14	- 515.90	- 514.40	4046.55	
						6087.55 True Vertical Depth	(164')

1282 ⟌26.00 1.022 23.64 2.360

2.2

989 / 1006

6252 / 6088 / 164

679

6087 / 4861 / 1226

6252 / 5002 / 1250

26'

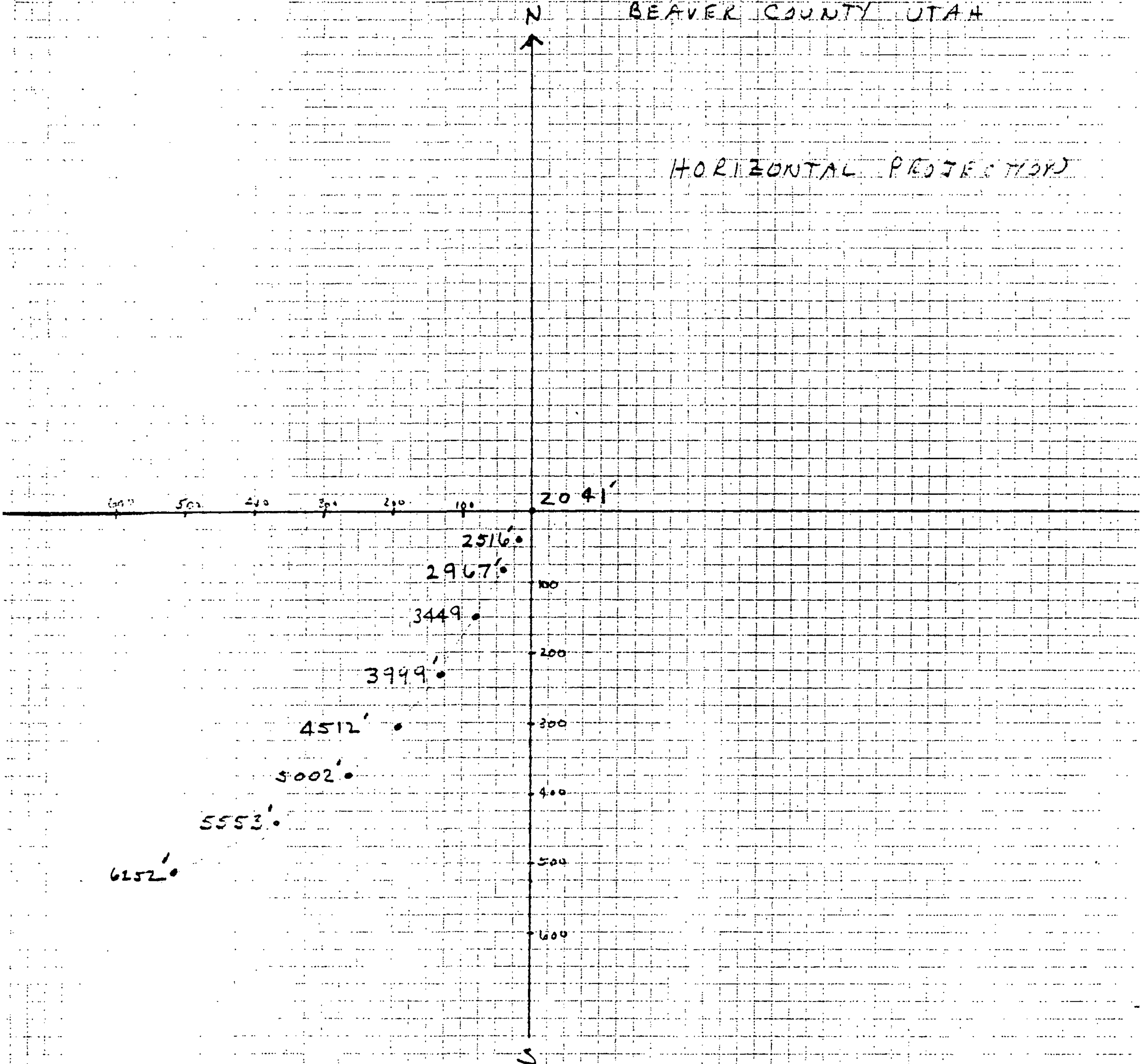
164 / 4211 ≈ .49

KGRA 52-21 ROOSEVELT HOT SPRINGS

SURFACE LOCATION: 2310' FEL & 990' FNL

S21 T27S R9W

BEAVER COUNTY UTAH



GETTY OIL COMPANY

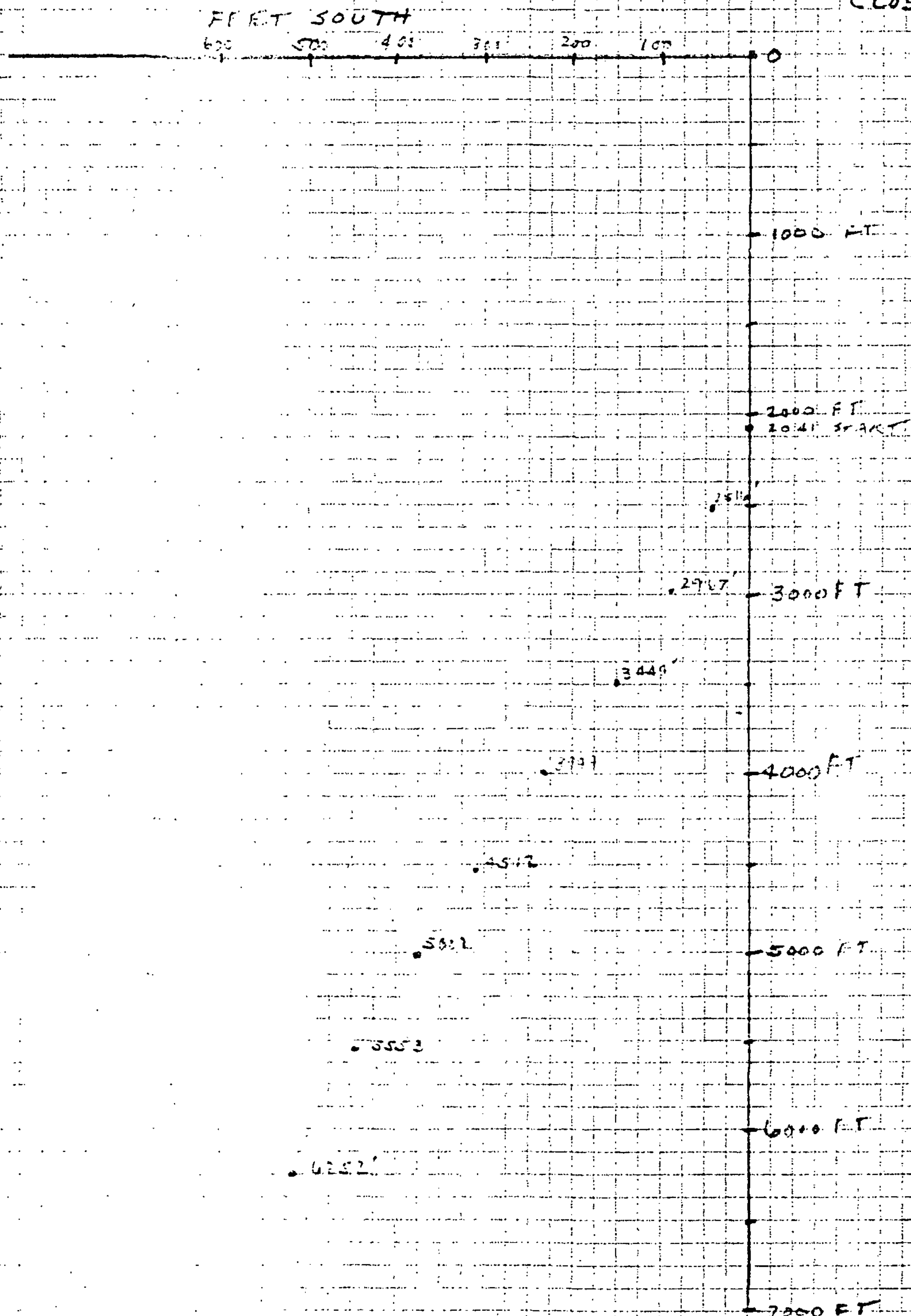
KGRA 52-21 ROOSEVELT HOT SPRINGS

SURFACE LOCATION: 2310 FEL & 990 FNL

S21 T27S R9W

BEAVER COUNTY, UTAH

CROSS-SECTION LOOKING WEST

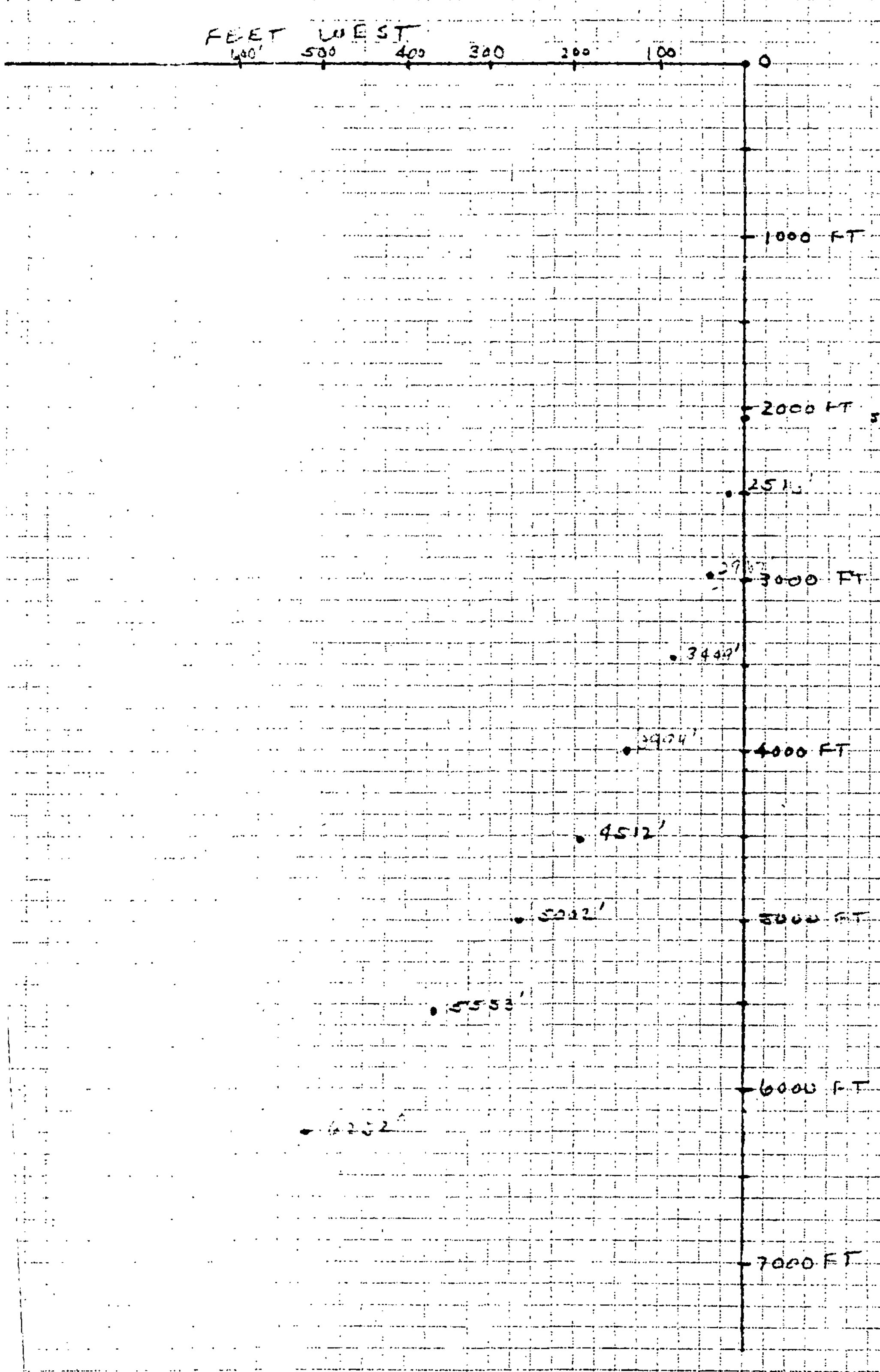


GETTY OIL COMPANY

KGRA 52-21 ROOSEVELT HOT SPRINGS

SURFACE LOCATION: 2310' FEL & 990' FNL
S21 T27S R9W
BEAVER COUNTY UTAH

CROSS-SECTION LOOKING NORTH



The American Association of Petroleum Geologists Bulletin
V. 66, No. 7 (July 1982), P. 879-902, 11 Figs., 6 Tables

Roosevelt Hot Springs Geothermal System, Utah—Case Study[1]

HOWARD P. ROSS, DENNIS L. NIELSON, and JOSEPH N. MOORE[2]

ABSTRACT

The Roosevelt Hot Springs geothermal system has been undergoing intensive exploration since 1974 and has been used as a natural laboratory for the development and testing of geothermal exploration methods by research organizations. This paper summarizes the geological, geophysical, and geochemical data which have been collected since 1974, and presents a retrospective strategy describing the most effective means of exploration for the Roosevelt Hot Springs hydrothermal resource.

The bedrock geology of the area is dominated by metamorphic rocks of Precambrian age and felsic plutonic phases of the Tertiary Mineral Mountains intrusive complex. Rhyolite flows, domes, and pyroclastic rocks reflect igneous activity between 0.8 and 0.5 m.y. ago. The structural setting includes older low-angle normal faulting and east-west faulting produced by deep-seated regional zones of weakness. North to north-northeast-trending faults are the youngest structures in the area, and they control present fumarolic activity. The geothermal reservoir is controlled by intersections of the principal zones of faulting.

The geothermal fluids that discharge from the deep wells are dilute sodium chloride brines containing approximately 7,000 ppm total dissolved solids and anomalous concentrations of F, As, Li, B, and Hg. Geothermometers calculated from the predicted cation contents of the deep reservoir brine range from 520 to 531°F (271 to 277°C). Hydrothermal alteration by these fluids has produced assemblages of clays, alunite, muscovite, chlorite, pyrite, calcite, quartz, and hematite. Geochemical analyses of rocks and soils of the Roosevelt Hot Springs thermal area demonstrate that Hg, As, Mn, Cu, Sb, W, Li, Pb, Zn, Ba, and Be have been transported and redeposited by the thermal fluids.

The geothermal system is well expressed in electrical resistivity and thermal-gradient data and these methods, coupled with geologic mapping, are adequate to delineate the fluids and alteration associated with the geothermal reservoir. The dipole-dipole array seems best suited to acquire and interpret the resistivity data, although controlled source AMT (CSAMT) may be competitive for near-surface mapping. Representations of the thermal data as temperature gradients, heat flow, and temperature are all useful in exploration of the geothermal system, because the thermal fluids themselves rise close to the surface. Self-potential, gravity, magnetic, seismic, and magnetotelluric survey data all contribute to our understanding of the system, but are not considered essential to its exploration.

INTRODUCTION

The Roosevelt Hot Springs geothermal system is located along the western side of the Mineral Mountains, approximately 12 mi (19 km) northeast of Milford, Utah (Fig. 1). The geothermal system is a high-temperature water-dominated resource, and is structurally controlled with permeability localized by faults and fractures cutting plutonic and metamorphic rocks.

The earliest exploration in the area was carried out in 1967 and 1968 by Eugene Davies of Milford who encountered boiling water in two holes on and adjacent to siliceous sinter deposits. Phillips Petroleum Co. initiated geothermal exploration at Roosevelt in 1972 and successfully bid for 18,871 acres (7,637 ha.) of the Roosevelt KGRA (Known Geothermal Resources Area) at the July 1974 competitive lease sale (Forrest, 1980). Exploratory drilling began in 1975 and the second well drilled, Roosevelt KGRA 3-1, encountered producible quantities of geothermal fluid. Additional drilling by Phillips Petroleum Co. and Thermal Power Co. has brought the total number of successful wells in the Roosevelt Hot Springs geothermal field to seven. A mobile 1.6 MWe generator has been installed in the field. Present development plans call for increasing capacity to 7 MWe in 1983 and 20 MWe by 1984. A discussion of the history of the Roosevelt Hot Springs KGRA and the exploration methods employed by Phillips Petroleum Co. is presented by Forrest (1980).

A large amount of geological, geophysical, geochemical, drilling, and reservoir data for the Roosevelt Hot Springs area has been made available in 1977-79 through the Department of Energy's Industry Coupled Program. A review of these data and earlier work helped to identify inconsistencies and weak areas in an already large geoscience data base. Additional work has helped to complete this data base. The integration of these data constitutes the present case study.

[1]Manuscript received, May 26, 1981; accepted, February 24, 1982. Paper presented at the AAPG Annual Meeting EMD Geothermal Session, June 2, 1981.

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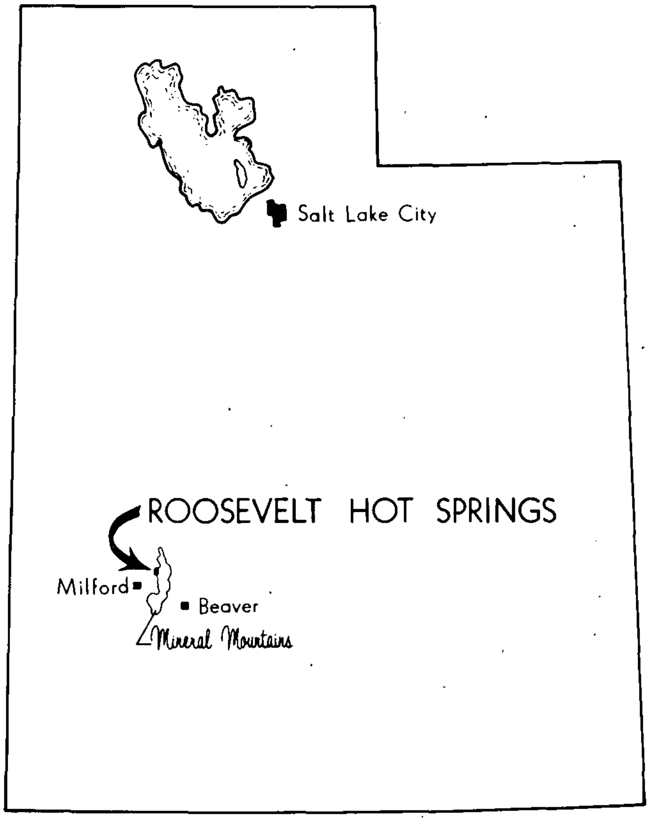

FIG. 1—Index map showing location of Roosevelt Hot Springs KGRA, Utah.

ROOSEVELT HOT SPRINGS GEOTHERMAL SYSTEM

Regional Setting

The Roosevelt Hot Springs KGRA is located on the western flank of the Mineral Mountains (Fig. 2). The Mineral Mountains are a north-trending horst bounded by Basin and Range normal faults and lie at the western edge of the transition zone between the Colorado Plateau and the Basin and Range physiographic provinces. The area is located on the western edge of the Intermountain seismic belt as defined by Smith and Sbar (1974). In addition, the Roosevelt Hot Springs KGRA lies along east-west-trending magnetic anomalies which follow the trend of the Wah Wah–Tushar mineral belt (Mabey et al, 1978). This belt has been the site of intrusive activity through the Tertiary and into Quaternary time. Associated with this igneous activity are deposits of uranium and base and precious metals.

The central part of the Mineral Mountains is a structural high relative both to adjacent ranges and also to the northern and southern parts of the range. In these northern and southern areas, sedimentary rocks of Cambrian through Cretaceous age are exposed. The southern area also contains volcanic and plutonic rocks of Tertiary age (Earll, 1957). In contrast, the central part of the Mineral Mountains contains Precambrian metasedimentary rocks and Tertiary plutonic rocks of the Mineral Mountains intrusive complex (Sibbett and Nielson, 1980). These Tertiary rocks possibly represent plutonic equivalents of the Marysvalle volcanic province which is exposed to the east, south, and southwest of the Mineral Mountains.

Local Setting

The geology in the vicinity of the Roosevelt Hot Springs geothermal system has been described in detail by Nielson et al (1978). The central part of the Mineral Mountains has been mapped by Sibbett and Nielson (1980). A simplified geologic map of the Roosevelt Hot Springs area is shown in Figure 2.

The oldest unit exposed in the area of the geothermal system is a banded gneiss which was formed from regionally metamorphosed quartzo-feldspathic sediments. The rock was metamorphosed to the upper amphibolite facies during middle Proterozoic time. The banded gneiss is strongly layered with adjacent layers distinguished principally on the content of mafic minerals. The rock is compositionally heterogeneous and contains thick sequences of quartzo-feldspathic rocks. The unit also contains metaquartzite and sillimanite schist layers which have been differentiated in the more detailed geologic study (Nielson et al, 1978).

The Mineral Mountains intrusive complex is the largest intrusive body exposed in Utah. K-Ar dating and regional relationships suggest that the intrusive sequence is middle to late Tertiary in age. In the vicinity of the geothermal system, the lithologies range from diorite and granodiorite through granite and syenite in composition (Fig. 2).

Rhyolite flows, pyroclastics, and domes were extruded along the spine of the Mineral Mountains 800,000 to 500,000 years ago (Lipman et al, 1978). The flows and domes are glassy, phenocryst-poor rhyolites. The pyroclastic rocks are represented by air-fall tuff and nonwelded ash-flow tuffs. Smith and Shaw (1975) hypothesized that young rhyolites such as these are indicative of an upper-level magma chamber which could serve as a heat source for geothermal systems.

Hot spring deposits in the vicinity of the geothermal system have been mapped as siliceous sinter, silica-cemented alluvium, hematite-cemented alluvium, and manganese-cemented alluvium. The principal areas of hot-spring deposition are along the Opal Mound fault and at the old Roosevelt Hot Springs. In both of these areas, the deposits consist of both opaline and chalcedonic sinter.

The geothermal reservoir at Roosevelt Hot Springs is structurally controlled. The controls are thought to be produced by the intersection of the faults mapped in the KGRA (Nielson et al, 1978, 1979). The structural evolution of the Roosevelt Hot Springs geothermal system is envisioned as follows. During rapid uplift of the Mineral Mountains structural block, westward-dipping low-angle normal faults formed. The most important of these is the Wild Horse Canyon fault (Fig. 2). This tectonic event produced intense zones of cataclasis both along low-angle fault planes and within the hanging wall of the principal fault block. The cataclastic zones within the hanging wall are steeply dipping zones up to 10 ft (3 m) wide which strike generally to the northwest. These zones in the hanging wall were produced by internal brecciation and interaction between rigid blocks during the low-angle faulting. East-west-trending high-angle normal faulting, represented by the Negro Mag fault, cuts the low-angle faults. The trend is parallel with the Wah Wah–Tushar mineral belt and is probably related to movement along this deep-seated structural trend. The Opal Mound fault and parallel structures are north-northeast-trending faults which are the youngest structures in the area. They localize siliceous hot-spring deposits and are often marked by zones of alteration and silicification of alluvium.

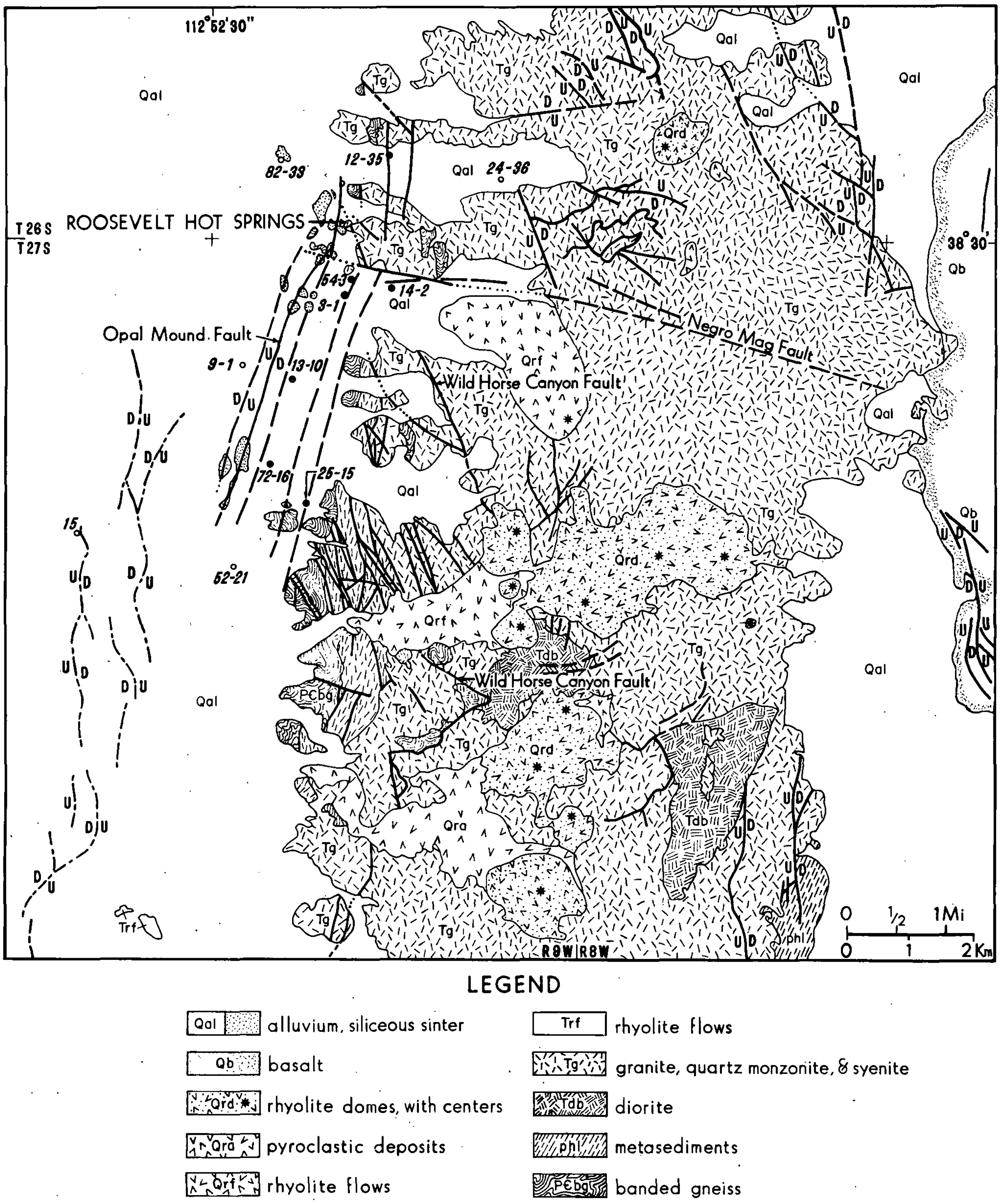

FIG. 2—Geologic map of Roosevelt Hot Springs KGRA and vicinity (after Sibbett and Nielson, 1980a).

Hydrothermal System

The limits of the hydrothermal system have been partially defined by deep drilling within the Roosevelt Hot Springs KGRA. The principal wells are shown on Figure 2, and important facts about the wells are presented in Table 1. The area of geothermal production is bounded on the east by the range front of the Mineral Mountains. On the west, the system is bounded by the Opal Mound fault, and on the south it terminates between Utah State 72-16, which is a producer, and Utah State 52-21, which is a hot but dry hole. The northern boundary of the system has not been determined.

The deep wells confirm that the host rocks for the hydrothermal system are the Tertiary plutonic rocks and Precambrian metamorphic rocks which have been mapped in the adjacent Mineral Mountains (Fig. 2). Thus, the rocks show little primary permeability and the system is controlled by faults and fractures (Lenzer et al, 1976; Nielson et al, 1978, 1979).

Fluid chemistry.—The hydrothermal fluids are relatively dilute sodium chloride brines which contain approximately 7,000 ppm total dissolved solids. Capuano and Cole (1982) provide the most comprehensive review of the chemistry of these fluids. Table 2 lists the composition of fluids collected from the wells and springs. These fluids are compositionally similar throughout the field, and differ mainly in their concentrations of calcium, magnesium, and bicarbonate which may be the result of mixing with calcium-rich nonthermal ground water.

Reservoir temperatures estimated from the Na-K-Ca contents of the brines range up to 554°F (290°C), and exceed the measured maximum temperatures by as much as 36 to 54°F (20 to 30°C). Capuano and Cole (1982) suggest that temperatures calculated by geothermometers may in fact be too high, reflecting changes in the fluid chemistry produced by flashing of the brines in the well bore. Their arguments are supported by geothermometer calculations based on the composition of the deep reservoir fluid determined from the combined analyses of brine, steam, and gas. These calculations suggest a reservoir temperature of 520°F (271°C) based on the Na-K-Ca geothermometer (Fournier and Truesdell, 1973) and a temperature of 531°F (277°C) based on the SiO_2 content of the fluid and the conductively cooled quartz saturation model of Fournier (1973).

The isotopic composition of the geothermal fluids indicates that they are of meteoric origin (Rohrs, 1980). A comparison of the fluid isotopic compositions with that of water from the mountain ranges to the east suggests that the thermal fluids could be derived either from the Mineral Mountains or ranges to the east.

Hydrothermal alteration.—Hydrothermal alteration in the geothermal system and the adjacent Mineral Mountains is localized along faults and fractures. There have been several periods of alteration and it is often

Table 1. Well Summary from Roosevelt Hot Springs, Utah

Well	Location	Depth	Status	T Max	Operator	Reference
O.H.-2	SW NW Sec. 10, T27S, R9W	2,250′	Deep T-gradient	NA	Phillips	Lenzer et al, 1977
O.H.-1	SE NE Sec. 17, T27S, R9W	2,321′	Deep T-gradient	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 9-1	NE NW Sec. 9, T27S, R9W	6,885′	dry	227°C	Phillips	Lenzer et al, 1977 Glenn et al, 1980
Roosevelt KGRA 3-1	SW NE Sec. 3, T27S, R9W	2,724′	producer	NA	Phillips	Lenzer et al, 1977
Roosevelt KGRA 54-3	SW NE Sec. 3, T27S, R9W	2,882′	producer	NA	Phillips	Lenzer et al, 1977
Roosevelt KGRA 12-35	NW NW Sec. 35, T26S, R9W	7,324′	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 13-10	SW NW Sec. 10, T27S, R9W	5,351′	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 82-33	NE NE Sec. 33, T26S, R9W	6,028′	dry	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Utah State 14-2	SW NW Sec. 2, T27S, R9W	6,108′	producer	254°C	Thermal Power	Lenzer et al, 1977 Glenn and Hulen, 1979
Roosevelt HSU 25-15	NW SW Sec. 15, T27S, R9W	7,500′	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Utah State 72-16	SE NE Sec. 16, T27S, R9W	1,254′	producer	243°C	Thermal Power O'Brien	Lenzer et al, 1977 Glenn and Hulen, 1979
Utah State 24-36	SW NW Sec. 36, T27S, R9W	6,119′	dry	NA	Thermal Power	
Utah State 52-21	NW NE Sec. 21, T27S, R9W	7,500′	dry	206°C	Getty Oil Co.	Glenn and Hulen, 1979
GPC-15	SE SE Sec. 18, T27S, R9W	1,900′	Deep T-gradient	72°C	Geothermal Power Corp.	Glenn and Hulen, 1979

not possible to separate these different periods on the basis of their mineralogy and chemistry. Older periods of hydrothermal alteration are associated with the intrusion of various phases of the Mineral Mountains intrusive complex. These episodes have produced minor copper mineralization which is generally associated with xenoliths of Paleozoic and Precambrian rocks. In addition, zones of sodium metasomatism have been identified with the contact zones between some of the felsic plutonic rocks (Sibbett and Nielson, 1980). The intrusion of the various phases of the pluton have also superimposed contact metamorphic assemblages on the older rocks.

A hydrothermal event which altered the fault zones and deposited pyrite and chalcopyrite occurred with the low-angle faulting. The hydrothermal alteration produced assemblages of quartz + chlorite + epidote + hematite. Hematite is commonly found as specularite veinlets and, where genetic relationships can be observed, hematite mineralization follows sulfide mineralization.

The hydrothermal alteration assemblages associated with the present geothermal system are crudely zoned with depth. The uppermost assemblage, occurring around the hot-spring deposits and fumeroles, is characterized by quartz, alunite, kaolinite, montmorillonite, hematite, and muscovite. Parry et al (1980) have studied the near-surface alteration and suggest that these minerals have formed above the water table by downward-percolating acid sulfate waters. Upward-convecting geothermal brines have produced, with increasing depth, alteration assemblages characterized by montmorillonite + mixed layer clays + sericite + quartz + hematite and chlorite + sericite + calcite + pyrite + quartz + anhydrite (Ballantyne, 1978). Thermochemical calculations and petrologic observations suggest that the brines are in equilibrium with the alteration assemblages produced by the upward-migrating fluids (Capuano and Cole, 1982).

Flow tests.—Most of the flow-test, production-logging, and reservoir-engineering data for the Roosevelt Hot Springs field is proprietary. However, data from Utah State 14-2 and Utah State 72-16 (Fig. 2, Table 1) are in the public domain and are summarized here.

A short flow test conducted in Utah State 72-16 on

Table 2. Chemical Analyses of Selected Thermal Waters[1]

		Wells				
		14-2	54-3	72-16	52-21	Hot Spring
Na	ppm	2,150	2,320	2,000	1,900	2,500
K	ppm	390	461	400	218	488
Ca	ppm	9.2	8	12.20	114	22
Mg	ppm	0.6	<2	0.29	3.9	0
Fe	ppm		0.03		6.9	
Al	ppm		<0.5		<0.1	0.04
Si	ppm	229	263	244	67	146
Sr	ppm		1.2	1.20		
Ba	ppm		<0.5			
Mn	ppm		<0.2			
Cu	ppm		<0.1			
Pb	ppm		<0.2			
Zn	ppm		<0.1			
As	ppm	3.0	4.3			
Li	ppm		25.3	16.0		0.27
Be	ppm		0.005			
B	ppm	29	29.9	27.2	27.0	38
Ce	ppm		0.27			
F	ppm	5.2	6.8	5.3	3.4	7.5
Cl	ppm	3,650	3,860	3,260	2,885.1	4,240
HCO_3	ppm		232	181	550.0	156
SO_4	ppm	78	72	32	86	73
NO_3	ppm				1.3	11
Total Dissolved Solids	ppm	>6,614	7,504	6,444	5,727	7,800
pH (at collection T)		5.9		7.53	7.3	7.9
T (at surface)	°C	14				55
T (bottom hole)	°C	268	>260	243	204	284
Geothermometer						
T (Na-K-Ca)	°C	286	297	288	210[2]	284
T (quartz cond.)	°C	276	263	256	158	212
T (quartz adiab.)	°C	244	234	229	150	194
T (chalcedony)	°C	274	259	250	134	197
Total Depth	m	1,862	878	382	2,289	

[1]From Capuano and Cole (1982). A blank indicates data not determined or information not available.
[2]Corrected for the Mg content of the fluid.

December 30, 1976, produced 454,546 kg/hr steam and hot water at a flowing wellhead pressure of 25 kg/cm^2 and a temperature of 432°F (222°C). A 24-hour test was conducted April 4 to 6, 1977, in 72-16. The total mass-flow rate was determined by the James method (James, 1966) as 595,000 kg/hr at a wellhead pressure of 20.7 kg/cm^2 psia and a temperature of 415°F (213°C). However, the mass-flow rate and wellhead pressure dropped throughout the test, indicating a longer flow test was needed to determine steady-state well production. Thermal Power Co. also concluded that the James method of determining the flow rate of a two-phase flow was unsatisfactory. On the basis of the longer test, it was calculated that Utah State 72-16 could yield 119,546 kg/hr of steam if flashed at 5.6 kg/cm^2. The well's electrical generating capacity was determined to be 12.5 Mw at a heat rate of 9,546 kg steam/hr/Mw.

Thermal Power Co. conducted a 48-hour flow test in geothermal well Utah State 14-2 between November 16 and 18, 1976. The last 35 hours of the test stabilized at 225,000 kg/hr and an enthalpy of 444.5 Btu/hr. Thermal Power Co. calculates that at a wellhead pressure of 4.9 kg/cm^2 and 17.8% flash, Utah State 14-2 has a generating capacity of 4.5 Mw. This capacity was calculated at a heat rate of 8,900 kg steam/hr/Mw. Well surges were observed during the test and their cause was attributed to temporary obstructions in the system that induced in-pipe flashing.

The Denver Research Institute (DRI) conducted several production logging tests in 1978 and 1979 (Butz and Plooster, 1979). Pressure and temperature data were logged versus depth during several constant flow-rate tests from 146,818 kg/hr to 263,636 kg/hr as measured by the James method. The flow rates are probably accurate to ±15%. Butz and Plooster (1979) were able to match the measured borehole temperature and pressure profiles by using a program that modeled two-phase flow in the well bore and determined the flash depth in the borehole and the productivity index of the well. The best models were those that included a component of compressible gas.

EXPLORATION METHODS

The Roosevelt Hot Springs geothermal area has served as a laboratory for the development and testing of geothermal exploration methods for some time (Ward et al, 1978). Since 1978 more exploration data have become available, including detailed geologic mapping, additional deep drill holes, additional electrical resistivity surveys, a reflection seismic profile, passive seismic data, and extensive application and testing of surface and subsurface geochemical methods. These new data allow us to expand upon earlier studies. With the results of the individual exploration methods and the supreme advantage of 20/20 hindsight, the most ef-

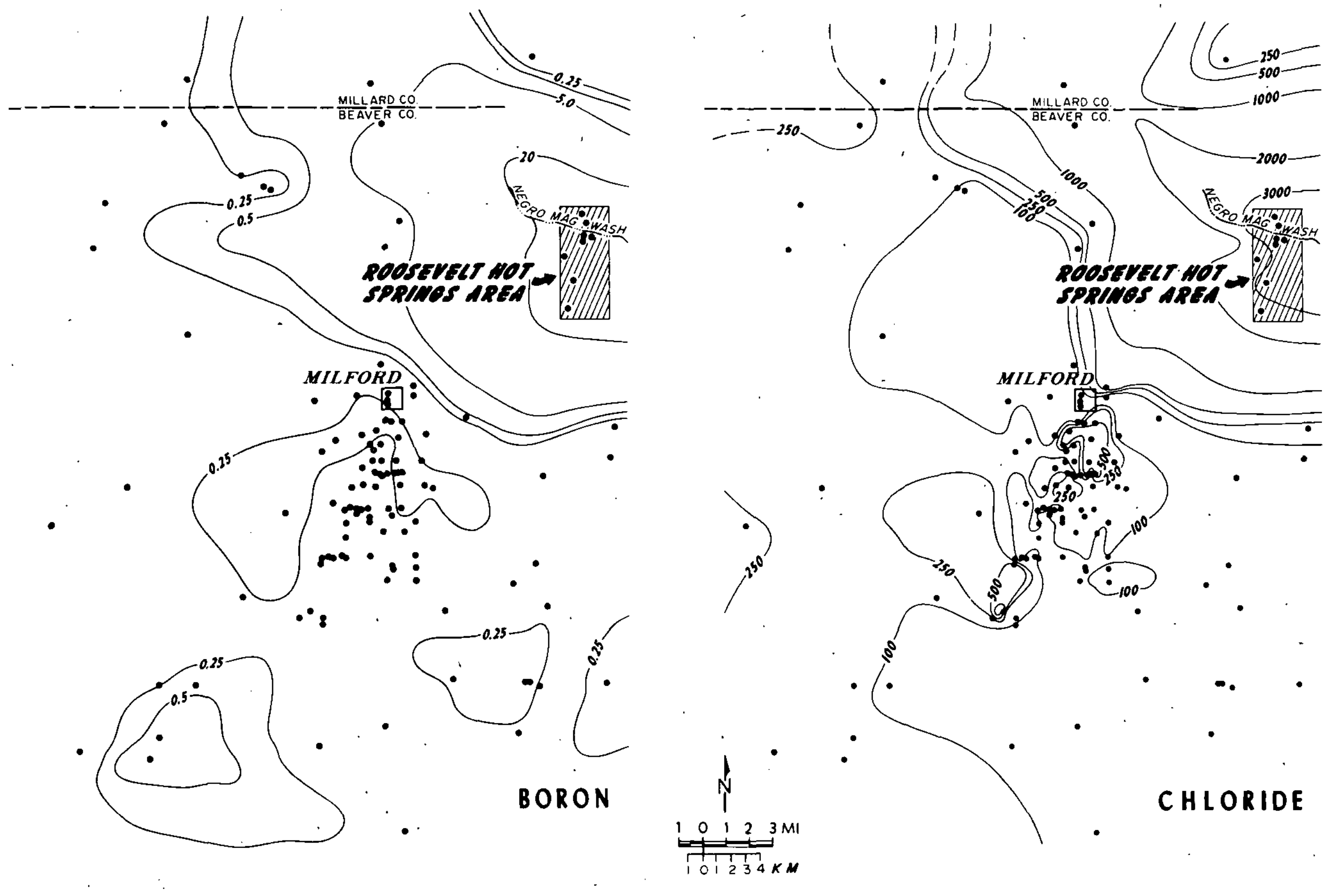

FIG. 3—Concentrations of boron and chloride in ground water in Roosevelt Hot Springs KGRA and adjacent Milford Valley. Data are from Mower and Cordova (1974) and Mower (1978) and was compiled by D. R. Cole.

ficient strategy for the exploration of the Roosevelt Hot Springs geothermal system will be presented. The results of some of the geological and geochemical exploration methods have been presented in the previous section.

Ground Water—Flow Systems and Chemistry

In contrast to the geothermal reservoir, the ground-water aquifer in the vicinity of the Roosevelt Hot Springs system is nonconfined. The study of this ground-water system is important from two aspects. First, it is probable that the present ground-water system provides the recharge to the thermal reservoir. This information bears on estimates of the longevity of the system and also whether the system will remain water-dominated during production or become vapor dominated (White et al, 1971) or possibly depleted. Second, studies of ground-water flow and chemistry are important exploration procedures which can be used to detect leakage of thermal waters into the regional ground-water systems. Hydrologic data can be inexpensively collected during thermal-gradient and heat-flow studies and for many areas are available in the ground-water literature.

Reconnaissance ground-water studies of the Milford and Beaver Valleys (Mower and Cordova, 1974; Mower, 1978) do not discuss the hydrology in the geothermal area but they do reveal the generalized ground-water configuration. The probable direction of ground-water movement within the KGRA is west-northwest toward Beaver Bottoms. The much higher elevation of the potentiometric surface in Beaver Valley than in Milford Valley—more than 984 ft (300 m) near the KGRA—prompts the speculation that if there were sufficiently transmissive fractures within the crystalline units of the Mineral Mountains, the difference in head would force ground water to flow through the range. The chances for this interbasin ground-water flow are enhanced by the cross-cutting structural zones like the Negro Mag fault zone. If ground water originating in the Beaver Valley drainage is recharging the geothermal reservoir, vast quantities of recharge may be available to the geothermal wells.

The ground-water papers by Mower and Cordova (1974) and Mower (1978) contain additional information which can be of value in exploration for the geothermal resource. Figure 3 is a map showing concentrations of chloride and boron in the vicinity of the Roosevelt Hot Springs. Lenzer et al (1976) show plots of ground water total dissolved solids which are similar to the patterns developed in Figure 3. These plots indicate that the chemical patterns produced by leakage of geothermal fluids into the ground-water system can be detected at significant distances from the geothermal system. These data are commonly available in the published literature, and its compilation and analysis can be a very efficient exploration method.

Solids Geochemistry

The geochemistry of hydrothermally altered rocks is routinely used by the minerals exploration industry as a guide to buried ore deposits. Yet despite the similarities between hydrothermal mineral deposits and active geothermal systems, the development of geochemical zoning models by the geothermal industry has been largely neglected. Recent studies (Browne, 1971; Ewers and Keayes, 1977) have suggested that trace-element zoning may be developed around the high-temperature centers of active geothermal systems. The studies at Broadlands have demonstrated that volatile and base metals are crudely zoned with depth as a result of decreasing temperature and boiling in permeable zones.

The trace- and major-element contents of surface and drill-hole samples from the Roosevelt Hot Springs thermal field were documented to determine if geochemical zoning models could be developed into an effective exploration tool (Bamford, 1978). Thirty-four elements were analyzed using an inductively coupled argon plasma spectrometer. Arsenic was determined by colorimetric methods, and mercury by a gold film detector. The analytical techniques are described by Christensen et al (1980a) and Capuano and Bamford (1978).

Geochemical analyses demonstrate that Hg, As, Sb, Mn, Cu, Co, W, Li, Pb, Zn, Be, Sr, and Ba have been transported and redistributed by the thermal fluids. These elements are strongly concentrated in hot-spring deposits (Table 3) with surface concentrations being typically more than an order of magnitude higher than

Table 3. Analyses of Hot Spring Deposits*

	Chalcedonic Sinter, Opal Mound	Mn-Cemented Alluvium	Altered Alluvium Over Fumarole
Na (%)	0.15	1.79	0.07
K (%)	0.14	3.12	0.26
Ca (%)	0.14	0.39	0.06
Mg (%)	0.01	0.10	0.03
Fe (%)	0.02	0.74	0.16
Al (%)	0.09	5.18	4.01
Ti (ppm)	19	560	2,040
P	—	651	336
Sr	33	386	266
Ba	—	4.9%	326
Cr	—	9	24
Mn	388	18.8%	173
Co	—	28	—
Ni	—	—	—
Cu	—	231	3
Mo	—	5	—
Pb	—	68	15
Zn	1	23	7
Cd	—	4	1
As	145	858	5
Sb	243	291	21
W	—	2,940	29
Li	11	17	8
Be	99.8	18.6	3.4
Zr	—	17	42
La	—	37	34
Ce	—	42	56
Hg (ppb)	352	2,210	49,300

*From Bamford et al (1980). Dash indicates that element was not detected.

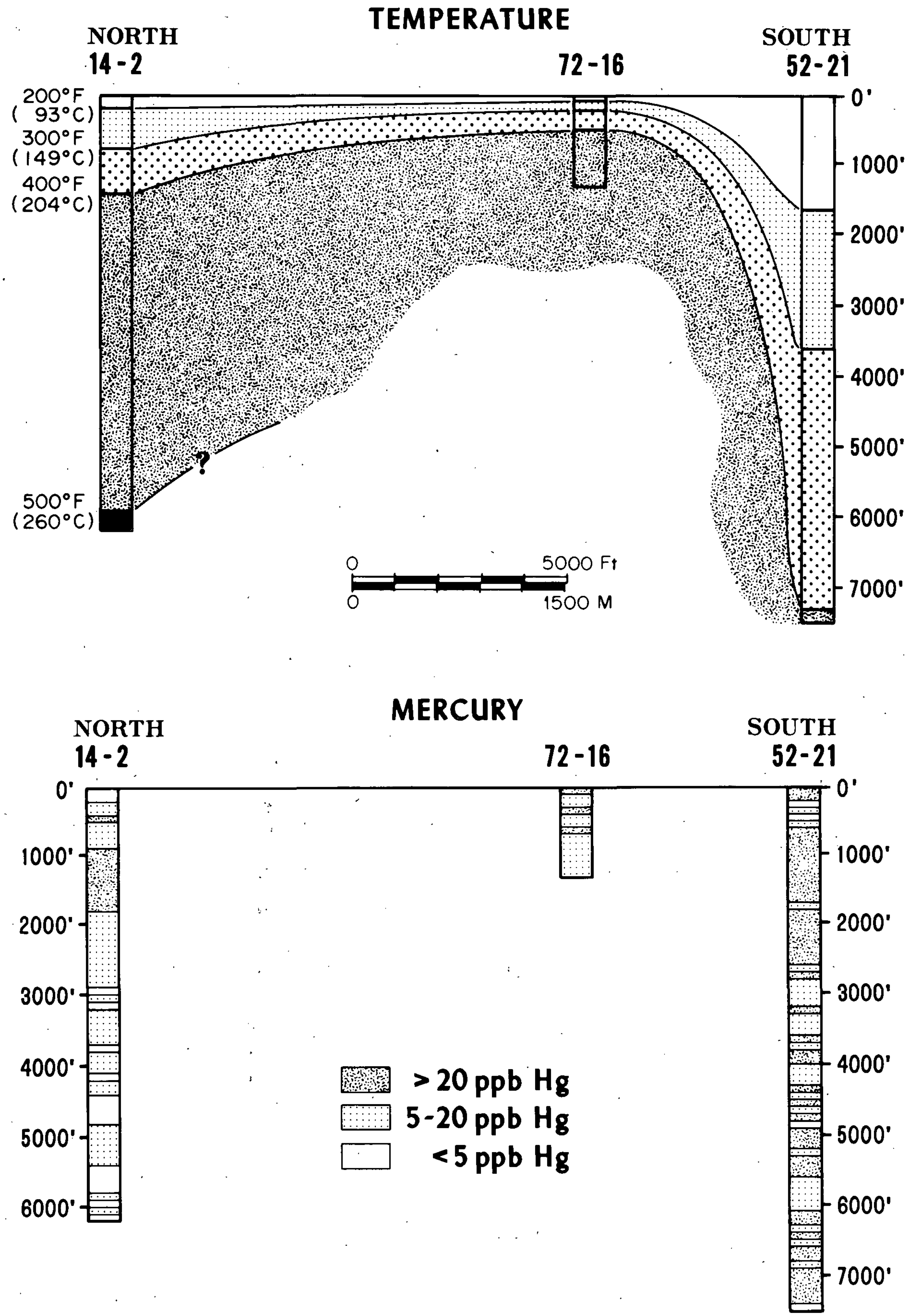

FIG. 4—Distribution of mercury and temperatures in wells 14-2, 72-16, and 52-21.

the subsurface abundances. However because of contamination during drilling, Hg, As, Sb, and Li are the only trace elements whose abundances in drill cuttings can be unambiguously related to chemical redistribution by geothermal activity.

Trace-element distributions, like the distribution of hydrothermal alteration minerals, are not pervasive but are localized along fractures and permeable alluvial horizons that have served as fluid channels. Detailed lithologic and geochemical logging has shown that at depth, arsenic is diagnostic of the hydrothermally altered rocks in the central part of the thermal field. Electron microprobe analyses and selective chemical leaching of the drill cuttings indicate that arsenic occurs primarily within pyrite and crystalline iron oxides formed from pyrite. Concentrations of arsenic as high as 4% have been detected in some pyrite crystals. In general, concentrations of arsenic are irregularly distributed within individual pyrite grains suggesting that the composition of the thermal brines has varied during the life of the system.

Mercury, in contrast to arsenic, is concentrated in the cooler parts of the thermal field, occurring in both weakly and highly altered rocks. Concentrations of mercury greater than 20 ppb define a broad envelope in the outer part of the system to depths approximately marked by the 420°F (215°C) isotherm (Fig. 4). This distribution reflects the extreme mobility of mercury at high temperatures within the thermal system.

Christensen et al (1980b) have experimentally investigated the mobility of mercury by measuring its progressive loss from drill cuttings and surface samples heated in air. These studies demonstrated that by 392°F (200°C) mercury loss had become significant, and had reached a maximum by 482°F (250°C). They concluded that mercury is present within the geothermal reservoir mainly as a native metal and suggested that its distribution reflects the present thermal configuration of the geothermal system.

At depths between 98.4 and 196.9 ft (30 and 60 m), pronounced enrichments in both arsenic and mercury occur in cuttings from thermal gradient holes located within the productive part of the thermal field. Despite the relatively small number of samples and the wide spacing between drill holes, geochemical data from shallow thermal-gradient holes nevertheless appear to be a useful means of prioritizing drilling targets that is independent of temperature measurements.

Concentrations of mercury and arsenic in soils over the thermal system are closely associated with hot-springs deposits and faults that are connected to the geothermal reservoir. The distribution of the anomalies and their shapes confirm that northeast- and west-northwest-trending (Negro Mag) structures control the near-surface hydrology of the thermal fluids.

Geophysics

Gravity Methods

Gravity methods are widely used in geothermal exploration to map structure and, in some cases, to detect directly the depositional products of hydrothermal systems. The latter use has been demonstrated in the Imperial Valley of California (Biehler, 1971) where the precipitation of silica and carbonates in sediments above the hydrothermal systems results in an increase in density of the sediments and thus a positive gravity anomaly.

The Bouguer gravity map, for the Roosevelt Hot Springs area (Fig. 5), is dominated by two major features: a broad, closed minimum centered over Milford Valley, and a north-trending elongate high over the Mineral Mountains. The data are fully terrain corrected. The gravity relief of approximately 36 mgal reflects a density contrast of 0.3 to 0.7 between valley fill and the bedrock (Table 4) within the range. A generally planar regional gradient of approximately 1 mgal/km decreasing eastward may be estimated for the Mineral Mountains area based on regional data (Crebs and Cook, 1976).

Bouguer gravity decreases gradually from the central part of the range to the valley low. The gravity high along the northern and western parts of the Mineral Mountains can be correlated with the Paleozoic limestones, Precambrian metasediments, and Tertiary diorites. In the vicinity of the geothermal field, denser lithologies such as the Precambrian banded gneiss and Tertiary diorites have been intruded by less dense felsic phases of the Mineral Mountains intrusive complex. Local perturbations on the smooth gradient (residuals), generally less than 2 mgal, can be readily explained by density variations of 0.1 to 0.3 g/cc within the crystalline rocks, as documented in Table 4. The interpretation of structural details beneath the alluvium within and west of the geothermal field is within the range of ambiguity due to uncertainties in the regional gradient and the variable density within the bedrock. A principal result of model studies supported by well control and density data is the absence of a large (>650 ft; 200 m) displacement in the bedrock surface along any single normal fault. Instead we see a gradual dip to the west, and possibly several minor faults near the Opal Mound fault and the outcropping range front. Ward et al (1978) present a similar interpretation, and these interpretations are in agreement with reflection and refraction seismic data presented later.

Magnetic Methods

The aeromagnetic map shown in Figure 6 is part, about 100 mi^2 (270 km^2), of a much larger 300 mi^2 (780 km^2) survey flown in 1975 (Ward et al, 1978). Data were obtained along east-west flight lines with an average line spacing of 1,380 ft (420 m). The flight path was smoothly draped at an average 1,000 ft (305 m) above ground level.

An extensive program of magnetic susceptibility measurements was undertaken in the summer of 1979 to provide support for the magnetic interpretation. The susceptibility data (Table 5) are in-situ susceptibility measurements at more than 60 locations on smooth, unweathered outcrop surfaces. A Bison magnetic susceptibility meter Model 3100 with in-situ coil accessory was used for the measurements. The data shown in Table 5 are fully corrected for outcrop surface

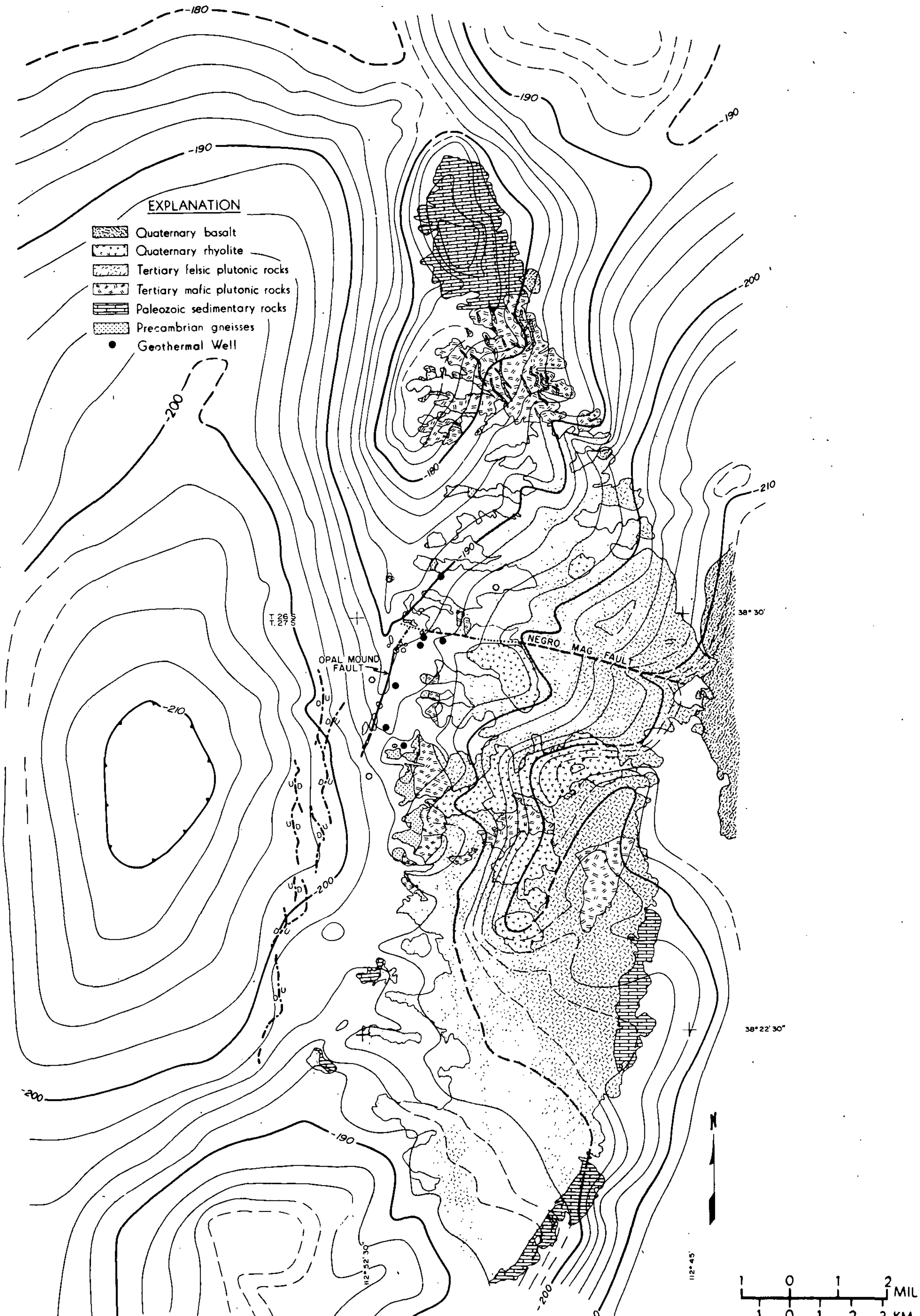

FIG. 5—Terrain-corrected Bouguer gravity map for Milford Valley and Mineral Mountains area (modified from Carter and Cook, 1978).

roughness. Additional susceptibility data determined from drill cuttings of Roosevelt KGRA 9-1 are reported by Glenn et al (1980).

A detailed inspection of the map reveals about 30 closed highs and lows over outcrop of the Mineral Mountains. The apparent complexity of the aeromagnetic map is not surprising in view of the complex igneous geology (Fig. 2) and the unavoidable variations in terrain clearance over rocks of varying magnetization. A study of the analog altimeter profiles shows actual terrain clearance values as little as 480 ft (150 m) over sharp topographic highs and as much as 1,200 ft (365 m) over canyons and between hills. This terrain clearance variation contributes to considerable east-west elongation and irregularities in the contoured map.

An in-depth discussion of the entire magnetic map is beyond the scope of this more general paper. An interpretation has been completed with integrated spatial correlation of magnetic and geologic maps, simple depth estimates, model comparisons, and extensive magnetic susceptibility data. On this basis, most of the magnetic features over the Mineral Mountain range can be attributed to mapped rock-type and altitude variations. Table 6 summarizes the main characteristics and interpreted sources for major anomalies identified in Figure 6. Most important to the present study is the interpretation of new information that relates to the struc-

Table 4. Densities of Lithologies from Mineral Mountains

Rock Type/Unit	g/cc ρ range	g/cc ρ (x)	No. of Samples	Reference
Tqm	2.43 - 2.80	2.64 ± .04	56	Glenn et al (1980)
"granitic"	2.45 - 2.71	2.59 ± .07	25	Carter and Cook (1978)
"granitic"	2.53 - 2.60	2.57 ± .02	11	Crebs and Cook (1976)
Td	2.54 - 2.90	2.76 ± .08	48	Glenn et al (1980)
Ts	2.43 - 2.63	2.55 ± .06	23	Glenn et al (1980)
Qrf/Qrd	2.16 - 2.24	2.22 ± .05	8	Crebs and Cook (1976)
Qrf/Qrd	2.22 - 2.38	2.31 ± .07	8	Carter and Cook (1978)
Obsidian	2.15 - 2.35	2.30 ± .07	8	Carter and Cook (1978)
P€bg (upper plate)	2.11 - 2.95	2.73 ± .28	136	Glenn et al (1980)
Gneiss and schist	2.63 - 2.74	2.69 ± .04	8	Crebs and Cook (1976)
Gneiss	2.63 - 2.74	2.69 ± .04	5	Carter and Cook (1978)
Limestone	2.55 - 2.97	2.71 ± .13	9	Carter and Cook (1978)
Quartzite[1]	2.50 - 2.74	2.62 ± .09	5	Carter and Cook (1978)
Alluvium		2.0 ± .1[2]		Carter and Cook (1978)
		2.05[3]		Glenn and Hulen (1979)

[1]Samples from Star Range, Rocky Range, Beaver Lake Mountains or Beaver Dam Mountains.
[2]Nettleton's method.
[3]Well log determination above water table.

Table 5. Magnetic Susceptibilities of Lithologies, Mineral Mountains, Utah[1]

Rock Type	Symbol#	No. Sites	No. Observations	Mean	Std. Dev.*	Range
				← Units of micro cgs →		
Alluvium	Qal	2	8	500	86	388-640
Rhyolite flows	Qrf	3	56	179	69	41-347
Rhyolite domes	Qrd	4	43	131	84	0-372
Rhyolite dikes	Trd	1	10	406	29	363-463
Diabase dikes	Tds	1	3	474	—	466-483
Granite dikes	Tgr-d	3	14	411	129	166-650
Granite	Tg	6; tr.	63	1,353	333	623-2,053
Seyenite	Ts	5; tr.	36	1,998	272	1,546-2,056
Porphyritic granite	Tpg	6; tr.	86	1,626	282	10-2,467
Quartz monzonite	Tqm	7; tr.	58	1,918	166	1,433-2,244
Hornblende gneiss	Tgn	4	24	2,310	830	1,520-3,440
Biotite gneiss	Tn	2	6	2,644	405	2,118-3,170
Sillmanite schist	P€s	tr.	10	22	32	0-89
Banded gneiss	P€bg	6; tr.	35	1,252	1,698	38-5,421

[1]Corrected in-situ susceptibilities, 6" diam. coil, Bison system.
Map symbol after Nielson et al, 1978.
*Standard deviation not always statistically significant, but indicative of susceptibility variability.
tr. = indicates observations along traverses, 500 to 2,500 ft long.

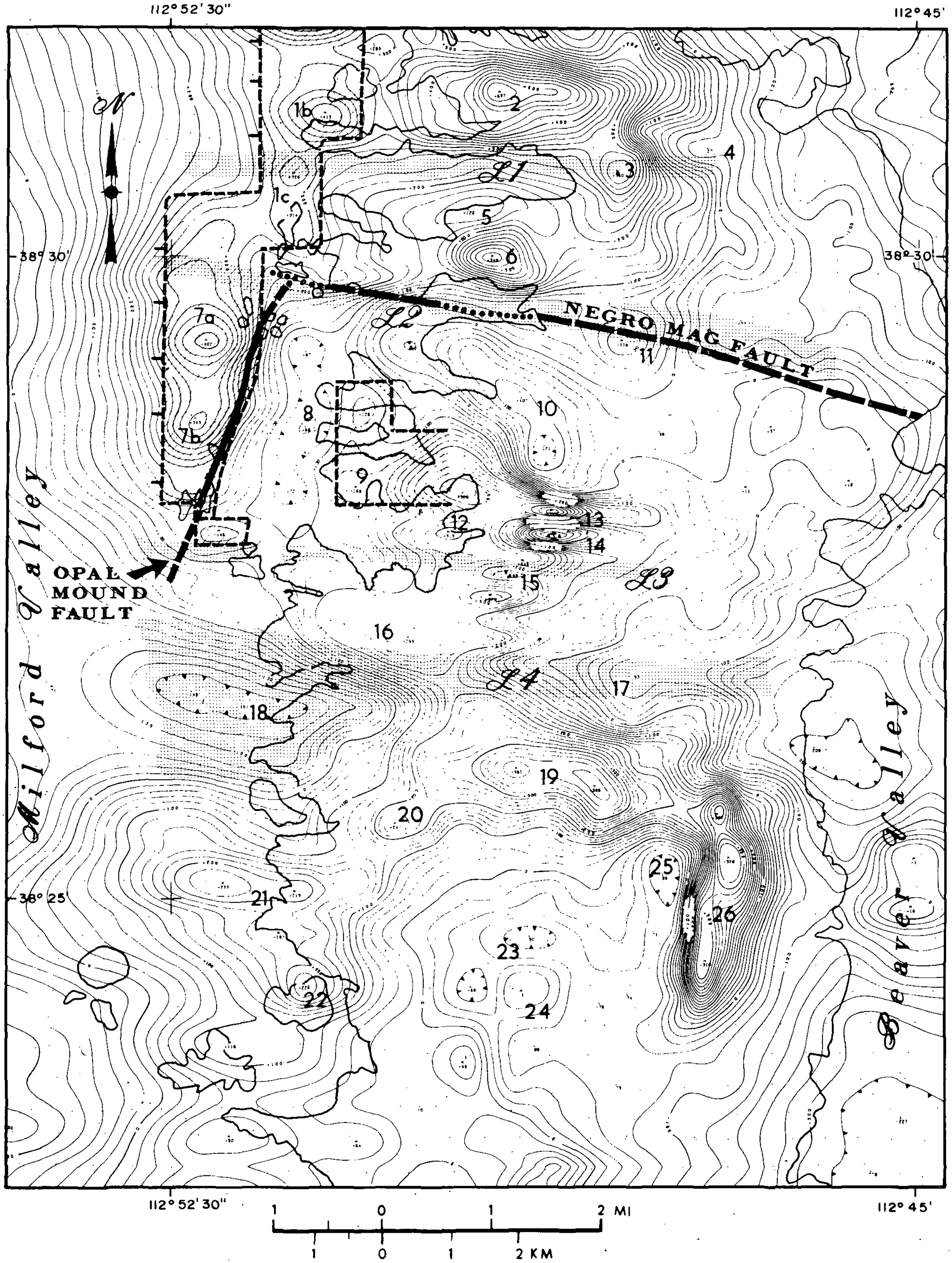

FIG. 6—Total magnetic intensity map of Roosevelt Hot Springs-Mineral Mountains area. Suballuvial magnetic sources and principal structural zones are indicated. Major anomalies are numbered for reference to text.

tural setting of the geothermal reservoir.

Several east-trending zones of low magnetic intensity cut across the Mineral Mountains (L1, L2, L3, L4). These magnetic trends are explained in part by the outcropping rock types and topographic features. The topography, geology, and magnetics are all expressions of east-west structural features. Feature L2 corresponds closely to the Negro Mag fault zone and its eastward continuation as mapped faults. It also appears to displace to the west a north-trending magnetic anomaly (7) closely associated with the Opal Mound fault. L3 terminates anomaly 7 between productive reservoir well Utah State 72-16 and the hot but dry hole, Utah State 52-21. The magnetic bodies appropriate to this magnetic field inclination are indicated on Figure 6.

The magnetic data over bedrock areas reflect the mapped geology and magnetic susceptibility. The identification of four east-trending structural zones, two of which (L1 and L3) may limit the north-south extent of the reservoir, is new information. The delineation of the alluvium-covered Opal Mound horst, bounded on the east by the reservoir-bounding Opal Mound fault, is the most important contribution to understanding the geothermal reservoir provided by the magnetic data. In view of known susceptibility values for the igneous and metamorphic rocks and the alluvium, it is not necessary to postulate an alteration low resulting from magnetite destruction to explain magnetic lows east of the Opal Mound fault (anomaly 8) and south of well Utah State 52-21 (anomaly 18).

Seismic Methods

A broad spectrum of seismic data is available at Roosevelt Hot Springs. Passive seismic data include long-term historical records of major earthquake activity, microearthquake surveys and, at a lower magnitude of naturally occurring seismic disturbance, seismic emissions or "noise" surveys. Single profiles of both refraction and CDP Vibroseis reflection data are also available for study.

In 1974 and 1975, an array of up to 12 portable high-gain seismographs was established within the Roosevelt Hot Springs and Cove Fort–Sulphurdale areas (Olson and Smith, 1976). In two survey periods totaling 49 days, 163 earthquakes of magnitude $-0.5 < M < 2.8$ were recorded. However, only four events could be associated with a 12-mi (20-km) length of the western flank of the Mineral Mountains which includes the Roosevelt Hot Springs area (Ward et al, 1978). Olson and Smith (1976) determined P-wave delays of up to 0.1 sec and detectable S-wave attenuation of ray paths across the Mineral Mountains. Ward et al (1978) interpreted the low-velocity effect and shear wave attenuation for these ray paths as possibly indicative of partial melting or major intense fracturing of crustal rocks beneath the southern part of the Mineral Mountains.

Robinson and Iyer (1981) observed P-wave residuals of up to 0.3 sec in a well-defined pattern corresponding to a region of anomalously low velocity (5 to 7%) centered under the geothermal area and extending from

Table 6. Magnetic Anomaly and Source Characteristics, Roosevelt Hot Springs Area, Utah

Anomaly	Amplitude (gammas)	Correlation[1] ρ topo/mag	Geologic Setting[2]	Probable Source[3]	Susc. Contrast (micro cgs)
1 b, c	120; 100	X	Qal, Qcal, Tqm	Tqm	2,000
2	280	H	Tqm, Tgr	Tqm, Ts; r.t.c.	~ 2,000
3	120+	M	Tqm	Tqm; r.t.c.	
4	− 80	R	Qrd, Qal	Qrd, Qal; r.t.c.	~ −1,400
5	−100	M-X	Qal, Tgr, Ts, Tqm	Qal, Tgr; i.t.c.	
6	190	H	Ts, Tqm	Ts, Tqm; r.t.c.	~ 2,000
7 a, b	300; 240	X	Qal, Qcal		2,000-5,000
8	− 40	X	Qal	Qal; i.t.c.	
9	50	H	Tgn, Ts, Tpg	Tgn, Ts; r.t.c.	
10	−200	L-R	Qrt, Qal	Qrt (reversed)	~ −1,500
11	50	H	Tqm	Tqm; r.t.c.	
12	190; 200	L-X	Ts, Qal	Ts;	~ 2,000
13*	250	H	Tqm	Tqm; * r.t.c.	
14*	−180	R?	Tqm	Tqm; * i.t.c.	
15*	− 80	R	Qrd	Qrd; * r.t.c.	
16	200	H-M	Tpg, Tn, Tgn, Ts, Tg	Ts, Tpg, Tgn; r.t.c.	~ 2,000
17	−150	X-R	Qrd, Tg	Qrd; i.t.c.	
18	− 90	L-X	Qal, Pєbg, Tpg	Qal; Pєbg, Tg	
19	350	H	Tg, Tn, Tpg, Qra, Qrd	Tn, Tg, Tpg; r.t.c.	~ 2,500
20	250	H	Tn, Tpg, Tgn	Tn, Tpg, Tgn; r.t.c.	~ 2,000
21	250	X	Qal, Tn, Tgn, Tpg	Tpg, Tgn, Tn	~ 2,000
22	150	L	Qal, Tbg, Tgr	Tbg, Tgr	
23	− 30	L-R	Qra, Qrd	Qra; i.t.c.	
24	20	H	Qrd	Qrd; r.t.c.	
25	− 50	H-R	Tq	— r.t.c.	
26	350	H	Tdb, Tg, Tbg	Tdb; r.t.c.	~ 3,500

[1]H = high, M = moderate, L = low, X = insignificant, R = reversed.
[2]Map symbol after Nielson et al (1978).
[3]r.t.c. = reduced terrain clearance; i.t.c. = increased terrain clearance; * indicates poor contour representation.

about 3 mi (5 km) depth down into the uppermost mantle. They preferred an explanation for these delays in terms of abnormally high temperatures and a small fraction of partial melt. Wechsler and Smith (1979) noted that these delays could arise from the fractured and possibly fluid-filled porosity of the western part of the Mineral Mountains pluton. Wechsler and Smith further noted the problems of accurate epicenter location and the limitations of P-wave studies in areas of complex near-surface lateral-velocity variations, such as exist at Roosevelt Hot Springs.

The relatively few earthquake locations determined for the western Mineral Mountains from this 49-day recording period and the complex near-surface velocity structure almost preclude the use of microearthquakes in delineating the Roosevelt Hot Springs geothermal system. P-wave delay studies may ultimately improve our understanding of the deep heat source at Roosevelt, but to this time have not contributed to delineation of the system.

Schaff (1981) has reported on seismic activity detected with the present nine-station seismograph array for the period October 1979 through January 1981. His results to date substantiate the earlier conclusion that seismicity in the immediate vicinity of Roosevelt Hot Springs is generally of a low level and somewhat episodic in nature. In the first 12-month period, no earthquakes were located within the anticipated production zone or along the Opal Mound fault. However, swarmlike activity was recorded east of the reservoir area in December 1980–January 1981. This trend of earthquake epicenters across the Mineral Mountains and east of the geothermal reservoir appears to lie along the eastern projection of the Negro Mag fault. More than 1,000 earthquakes of magnitude less than 1.5 have occurred in three major swarms between July and December 1981. The epicenters for these most recent events define a trend extending from the surface to 4 mi (6 km) depth parallel with the trend of the Negro Mag fault (L. McPherson and G. Zandt, personal commun.).

Seismic emissions survey.—Seismic emissions surveys have been promoted by several geophysical contractors as a geothermal exploration method in which the seismic emissions or "noise" would hopefully delineate active fault and fracture zones possibly associated with geothermal activity. The method employs an array of geophones (four or five) spaced approximately 2,000 ft (610 m) apart. In surveys at Roosevelt Hot Springs (Katz, 1977a, b) data were recorded at the array for 1 to 3 days, then moved to another station. Five such stations within a 14 to 16 mi^2 (36 to 41 km^2) area constitute a survey. The data were edited and processed with algorithms which determine the noise source locations based on delay times computed for a half-space velocity model and the correlation of these delays with the observed data.

A reevaluation of these data has been completed by Ross et al (1982). As employed at Roosevelt Hot Springs, the seismic emissions survey may indicate areas of geothermally induced seismic noise, but it is dominated by other noise sources and is imprecise in defining geothermal conduits. The correlation procedure is severely limited by model simplicity and velocity assumptions, and generally recognizes source direction more accurately than distance to the seismic noise source. A more refined velocity model could perhaps improve the resolution of the noise source areas through a higher correlation of source-to-geophone array delay times. It is unlikely that the velocity model could be refined enough to justify inclusion of the method in geothermal exploration in complex geologic environments.

Seismic refraction.—In April 1977, a 19 mi (30-km) long seismic refraction profile was recorded across the Roosevelt Hot Springs geothermal area (Gertson and Smith, 1979). Multiple shots at seven different shot locations were used to provide multiple subsurface coverage. Although the large station spacings of nearly 820 ft (250 m) were not adequate for locating narrow structural features, Gertson and Smith were able to define a somewhat generalized velocity model for the area and also determined that the first large displacement in range front faulting occurs at least 0.6 mi (1 km) west of the Opal Mound fault. P-wave attenuation across the geothermal reservoir was much less than attenuation in other parts of the profile, and Gertson and Smith (1979) concluded that the record sections did not appear to contain any evidence for seismic waves that had penetrated and returned from "hot rock" or magma chambers even at great depth. One complicating aspect of the refraction study is that the eastern part of the refraction profile, of necessity, followed the Negro Mag fault zone from the reservoir area eastward across the Mineral Mountains.

Seismic reflection.—One profile totaling 27 mi (43 line km) of detailed reflection seismic data was available for study of the Roosevelt Hot Springs area. Line 5 and 5 OPTW of a GSI speculation survey, recorded in March and April 1978, crosses the Mineral Mountains along the same path as the refraction survey. The data cannot be reproduced, but an interpretation of the data is presented here.

The GSI profiles are high-quality 24-fold CDP Vibroseis data with a 200 ft (60 m) group interval and a 400 ft (120 m) vibrator point interval. The source consisted of 16 sweeps per VP, using a 12-sec sweep and a 12 to 60 Hz sweep band. The sample rate was 2 msec. Processing included deconvolution, velocity analysis, CDP stack, and migration.

Our interpretation of the reflection profile is presented in Figure 7. The time-to-depth conversion is supported by 14 velocity analyses along the profile. These velocities are generalized in Figure 7. The refraction survey and sonic logs of three deep wells in the geothermal system—Utah State 52-21, 72-16, and 14-2—also provide velocity control. Although similar velocities are noted for alluvium, the refraction survey and well logs indicate much higher velocities at depth than velocity analyses of CDP reflection data. To those depths where good reflection quality persists and the coherence of the velocity analysis is well supported, the reflection survey velocities are considered more valid. At greater depths (times), the velocity analyses appear to be less valid and may be as much as 40% too low.

Figure 7 shows gently dipping layers in unconsolidated sediments to depths of 4,000 ft (1,220 m) in

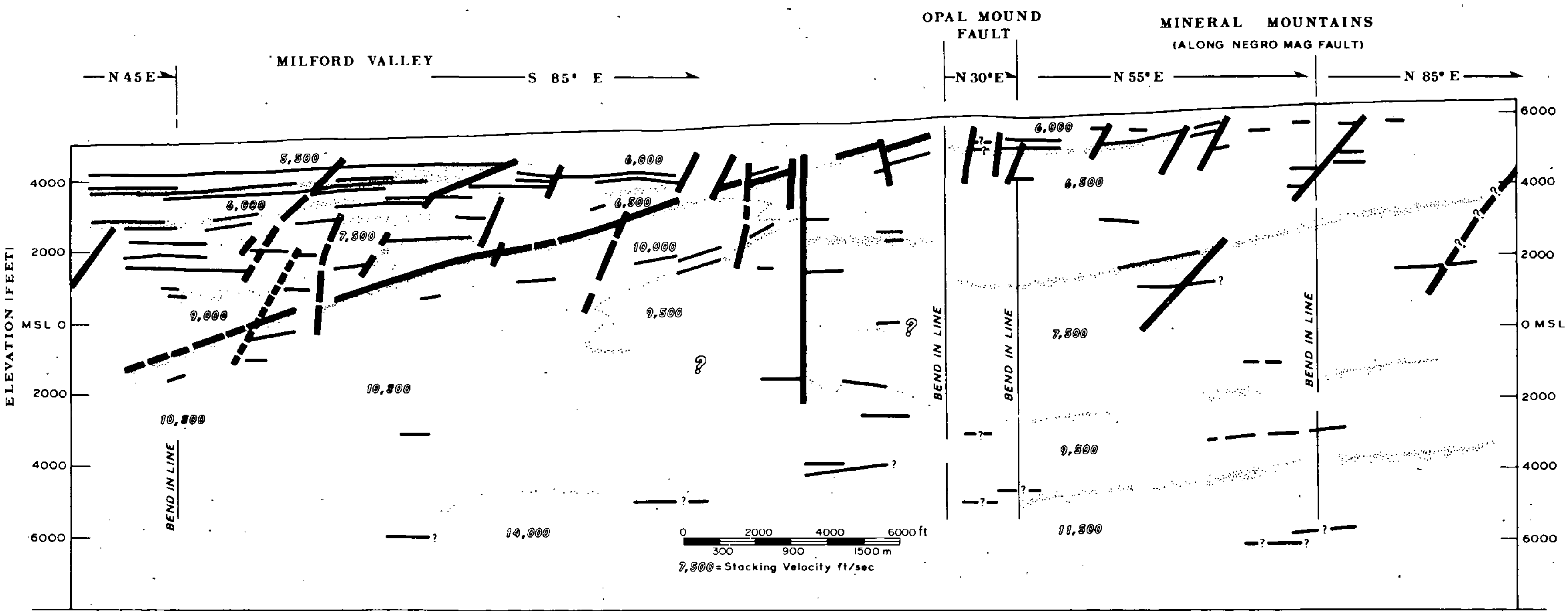

FIG. 7—Structural interpretation of Vibroseis profile GSI-5. Units as determined from velocity analysis and reflection continuity.

the Milford Valley. This sequence thins abruptly to the east as the Opal Mound fault is approached. Numerous faults cut the sedimentary sequence. A very prominent reflector which dips westward at approximately 20° from the Opal Mound fault indicates the base of the unconsolidated sediments. Few coherent reflectors are noted beneath this interface, presumably a sediment-igneous or sediment-metamorphic contrast.

The complexity of Basin and Range faulting is indicated by the reflection data. Unfortunately, the trend of the profile is along the southern edge of the horst block indicated by magnetic data, then N30°E along, and at a small angle to, the Opal Mound fault. The faulting is indicated, but less clearly than would be the case for an east-west line. East of the Opal Mound fault, the higher velocities and lack of coherent reflections indicate igneous rocks extending to depth. The interpretation of the vertical displacement along Basin and Range faults within the first 0.25 seconds is difficult because of high noise levels and probable lateral energy returns associated with the Opal Mound fault area. No single, major vertical displacement is indicated but a complex series of small displacements, of the order of 100 to 320 ft (30 to 100 m), is suggested by the data.

Electrical Studies

The recognition of Roosevelt Hot Springs as a relatively shallow geothermal resource of commercial potential has resulted in numerous electrical surveys designed to characterize the electrical resistivity distribution, and to test the effectiveness of various methods. To a large degree, Roosevelt Hot Springs has become a "test" area because of the type of resource and amount of publicly available supporting data. Included in the electrical surveys already completed are the following: magnetotelluric (MT), controlled source audio magnetotelluric (CSAMT), natural source AMT, controlled source EM, spontaneous polarization or self-potential (SP), induced polarization (IP), and electrical resistivity.

Electrical resistivity surveys.—Electrical resistivity surveys include both the dipole-dipole and bipole-dipole arrays. The bipole-dipole survey (Frangos and Ward, 1980) was undertaken with the knowledge of the resistivity distribution primarily to characterize the effectiveness of the method. The dipole-dipole survey includes at least 50 lines of various lengths, electrode separations, and depth penetrations, and constitutes one of the most complete resistivity data bases assembled. In view of this, and because all the other electrical methods are influenced by the resistivity distribution mapped by these data, a major effort was devoted to interpreting these data using two-dimensional numerical models (Ross et al, 1982).

Figure 8 shows the location of the resistivity lines which were numerically modeled and presents a summary of the intrinsic resistivity distribution for the depth interval of 330 to 500 ft (100 to 150 m). The modeled resistivity data are supplemented with a qualitative interpretation of more than 30 additional lines of data. The modeled resistivities at these depths have been transferred from the interpreted resistivity sections shown in Figure 9.

The resistivity distribution, even as generalized by numerical modeling, is quite complex. Generally high resistivities (100 to 500 ohm-m) are observed in the range over the Precambrian gneiss and Tertiary intrusive rocks. Very low (<10 ohm-m) to low (10 to 20 ohm-m) resistivities, often modeled as thin vertical conduits, occur along the trend of the Opal Mound fault. West of the Opal Mound fault moderate to high-resistivity (30 to 400 ohm-m) layers overlie layers of moderate to very low resistivity. This appears to represent dry alluvium (above the water table) overlying areas where fresh ground waters mix with warmer, more conductive geothermal fluids migrating downdip from the Opal Mound fault and other conduits. Very low resistivities (2.5 ohm-m) observed along UU line 8100N may represent geothermal outflow, dissolved salts in Lake Bonneville sediments, or some mixture of both. South of the survey base line and Utah State 52-21, high resistivities near the surface decrease with depth but do not indicate the presence of geothermal fluids noted to the north. Figure 8 provides a better comparison with key geologic features and drill-hole locations than does the section representation (Fig. 9), and shows that the low-resistivity area of the geothermal system is bounded by high resistivities southwest, south, and east of the geothermal system at these depths. Low and moderate resistivities extend well into the range along the Negro Mag fault zone. Most of the low-resistivity areas straddle or lie east of the Opal Mound fault at these depths. Resistivities of 3 to 10 ohm-m along the fault are bounded by 10 to 20 ohm-m on all sides as a crude zoning pattern. The resistivity of the alluvium is generally 20 to 50 ohm-m to the west, and much higher (200 to 400 ohm-m) to the south. The probable continuity of resistivity zones is indicated by heavy dashed lines (Fig. 8).

Ross et al (1982) present a similar map for the depth interval 1,500 to 2,000 ft (450 to 600 m) which is less complex because only 1,000 ft (300 m) dipole data could be used for this depth and because of the resistivity averaging inherent in modeling large separation data. The high resistivities east of the Opal Mound fault extend to the west but are substantially reduced to the south (GOC lines 1, 2, 4, 5). In fact, 50 to 400 ohm-m alluvium becomes quite conductive (5 to 20 ohm-m) with the increased depth. The Negro Mag Wash area is still a moderate resistivity reentrant into the range. The alluvium west of the Opal Mound fault has become quite low in resistivity (5 to 15 ohm-m), probably due to the downdip migration of conductive thermal fluids leaking from the geothermal system. The region of the Opal Mound fault itself is modeled as 30 to 50 ohm-m except at the southernmost part of the Opal Mound itself (8 ohm-m). This is in marked contrast to the low resistivity expression at 330 to 500 ft (100 to 150 m) depth. A high resistivity (450 ohm-m) body is indicated less than 1,250 ft (380 m) south of successful well Utah State 72-16, which reached total depth at 1,256 ft (383 m). The hot but dry Getty Oil Co. well, Utah State 52-21, is located in a resistive (100 to 400 ohm-m) area. Several productive wells (Roosevelt KGRA 12-35, Roosevelt KGRA 54-3, Roosevelt KGRA 3-1, Utah State 14-2, Roosevelt KGRA 13-10, and Roosevelt HSU

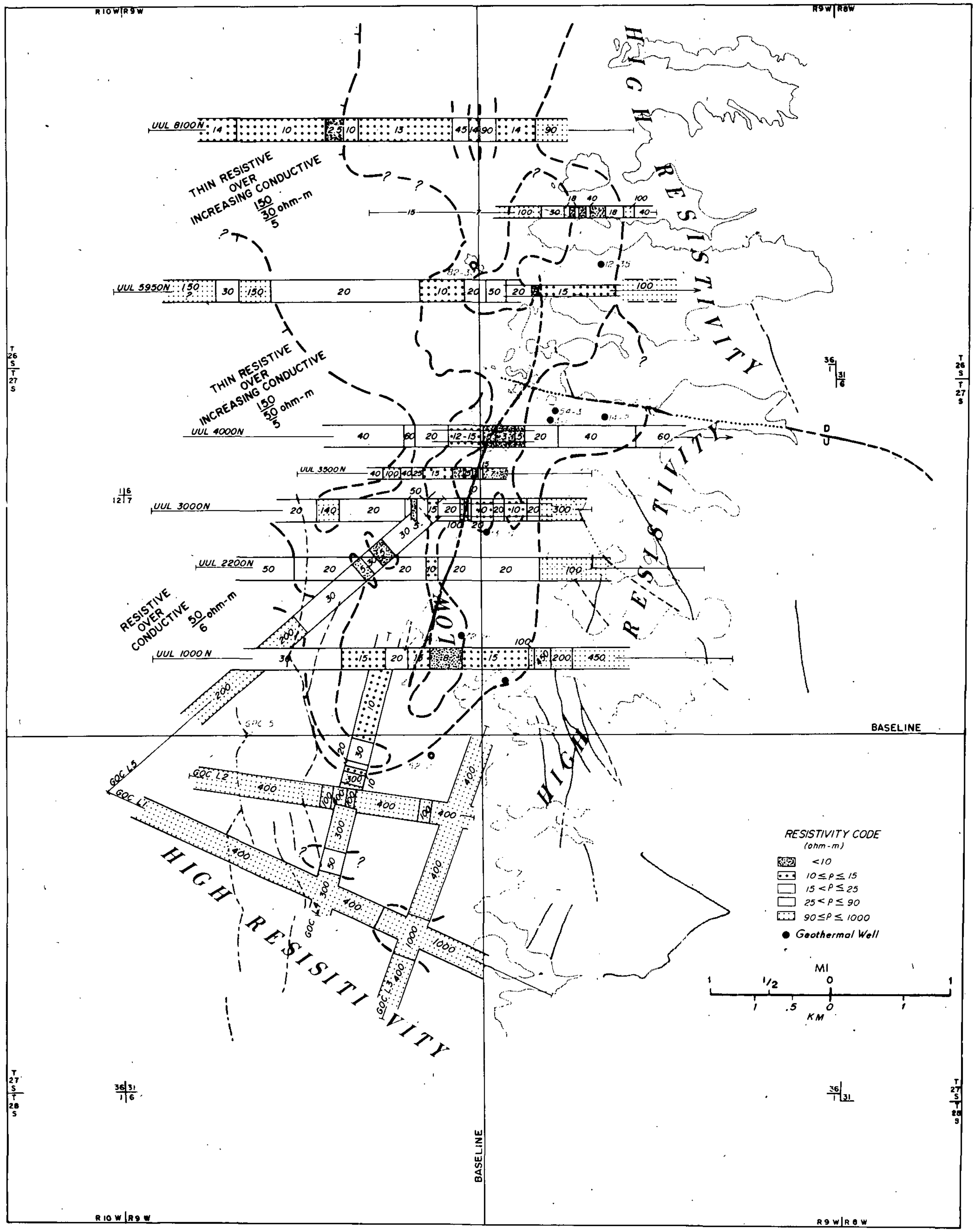

FIG. 8—Interpreted electrical resistivity distribution for depth interval 100 to 150 m, Roosevelt Hot Springs KGRA.

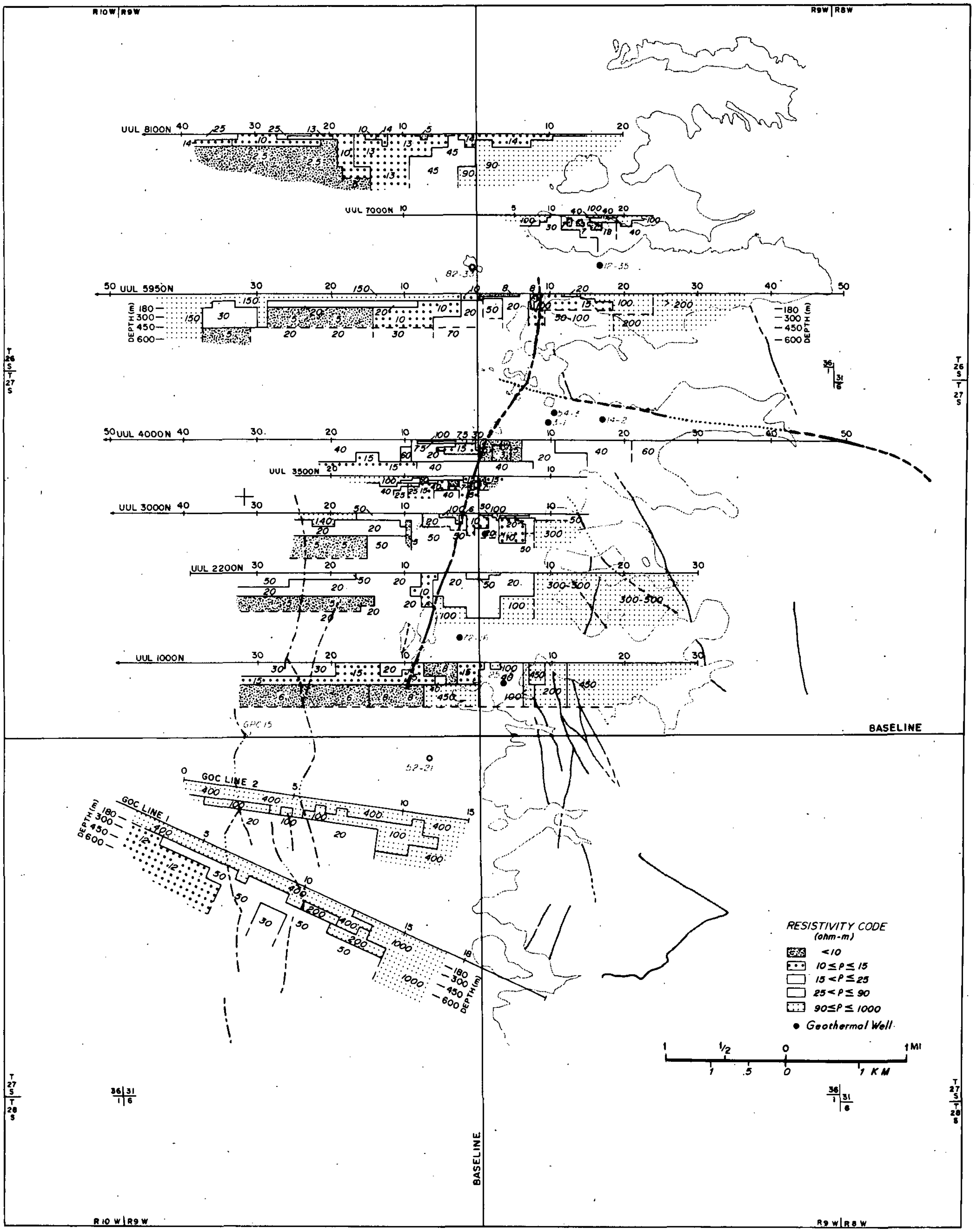

FIG. 9—Interpreted electrical resistivity sections, Roosevelt Hot Springs KGRA.

25-15) are sited in areas of moderate (15 to 40 ohm-m) resistivity which are relative lows.

The interpretations presented in Figure 9 are non-unique and are limited by the grid size which becomes coarser with depth, the validity of the two-dimensional model, the goodness of fit of computed to observed data values, and the choice of body size, position, and resistivity. Nonetheless, careful modeling of dipole-dipole resistivity data has resulted in a more accurate representation of earth resistivity distribution for a given cost than any other electrical method. The non-uniqueness is reduced by using a network of profiles, several of which intersect, and the integration of geologic data, such as the detailed (1:24,000) map of Nielson et al (1978).

Near-surface electrical methods.—Sandberg and Hohmann (1980) described a controlled source audio-magnetotelluric (CSAMT) survey at Roosevelt Hot Springs which compares well with dipole-dipole resistivity data for shallow depths ($\lessapprox$ 330 ft; 100 m). Minor disagreements in the interpretative models from these methods serve to indicate the noise and ambiguity levels for each method. Chu et al (1980) described induced polarization surveys which were unable to map clay alteration or pyritic zones related to the geothermal system.

Corwin and Hoover (1979) discussed thermoelectric coupling and electrokinetic coupling as possible mechanisms for generating self-potential anomalies observed over geothermal systems. They reported a dipolar anomaly located directly over the Opal Mound fault and concluded that it results from geothermally generated electrical activity along the fault. More detailed self-potential studies have been completed by Sill and Johng (1979). A complex "quadrapolar" anomaly of two lows (-100 mv) and two highs (+25 and +50 mv) was found to be associated with the southern part of the geothermal system. The 100-mv low which occurs over the Opal Mound fault is the most unambiguous expression of the geothermal system. Other self-potential features could arise from large resistivity contrasts, the movement of ground water from higher to lower elevation, or fluid movement along structures. Additional numerical modeling is required to obtain a more complete understanding of the SP response at Roosevelt Hot Springs, but the observed data document a complex expression of the geothermal system.

Magnetotelluric (MT) studies.—The magnetotelluric (MT) method is often used in both the reconnaissance and detailed stages of geothermal exploration. The earth's electric and magnetic fields vary as a function of frequency in response to natural electrical (telluric) currents flowing within the earth's crust. Through precise measurements of the electric and magnetic field components made at the surface, one may obtain information relating to the impedance distribution (i.e., electrical resistivity) to depths as great as 25 mi (40 km) within the earth's crust. The reader is referred to an excellent paper by Vozoff (1972) for a detailed description of the method.

In keeping with the detailed geophysical definition of the Roosevelt Hot Springs area an extensive MT network of 93 stations was established in the Milford Valley–Mineral Mountains area (Wannamaker et al, 1978, 1980).

The reduced data for a central 14 mi (22 km) east-west profile are presented as observed apparent resistivity versus frequency for the TM (transverse magnetic) mode in Figure 10a. The best-fit model results computed for a two-dimensional model geometry TM mode (Fig. 10b) are seen to be very similar. The comparison between observed and modeled impedance phase (not shown here) is also good but has a poorer fit in the central part of the profile. The two-dimensional model that produced this best fit is shown as Figure 10c. The rather fanciful 50,000 ohm-m prism in Figure 10c was inserted in an unsuccessful attempt to match the mid-frequency (10 to 0.1 Hz) data. A frequency dependent, three-dimensional current gathering effect involving the Milford Valley to the west is now believed responsible for these particular modeling difficulties (Wannamaker et al, 1980).

Stations 78-22 through 76-2 are located over low-resistivity alluvium and valley fill. Figure 10c indicates resistivities decreasing to near 1 ohm-m at depths of 1,300 ft (400 m). Dipole-dipole resistivity data and Schlumberger soundings also indicate a thin (330 ft; 100 m) resistive (100 ohm-m) near-surface layer, with resistivity decreasing to 5 ohm-m or less at depths of 1,000 ft (300 m). Station 76-3 near the center of the cross section occurs over a buried horst inferred from gravity and magnetic data and exhibits high resistivities at moderate depths. Station 76-4 was sited along the Opal Mound fault (Wannamaker et al, 1978), which has been identified earlier as a narrow, vertical conductive zone associated with ascending geothermal fluids. Silica cementation has locally decreased the porosity but fracture permeability remains.

Wannamaker et al (1978, 1980) noted great difficulty in obtaining good representations of both the TE and TM mode resistivities for any single one-dimensional or two-dimensional model. Extensive three-dimensional model studies indicate the limitations of one- and two-dimensional modeling at Roosevelt Hot Springs, and probably for most Basin and Range-type geothermal reservoir areas. A few of their more general conclusions are restated here.

1. Current gathering in the valley results in a regional distortion of the electric field affecting all stations at Roosevelt Hot Springs for lower frequencies.

2. The TM mode is most appropriate for two-dimensional interpretation, and has yielded good results for geometrically regular three-dimensional prisms.

3. Clays (ϱ = 1 to 2 ohm-m) may exist to depths of several hundred meters in the Milford Valley. These overlie more than 0.6 mi (1 km) of semiconsolidated and unconsolidated sediments and volcanics of moderate ($\varrho \simeq 25$ ohm-m) resistivity.

4. A geometrically regular reservoir of conductive brine beneath the thermal anomaly seems improbable, so the search for any economic hydrothermal reservoir at Roosevelt Hot Springs using MT must be considered unsuccessful at this time. If present, it is not resolved by the two-dimensional TM algorithm. The brine-saturated reservoir zone is clearly three-dimensional and difficult to model satisfactorily with present interpretation

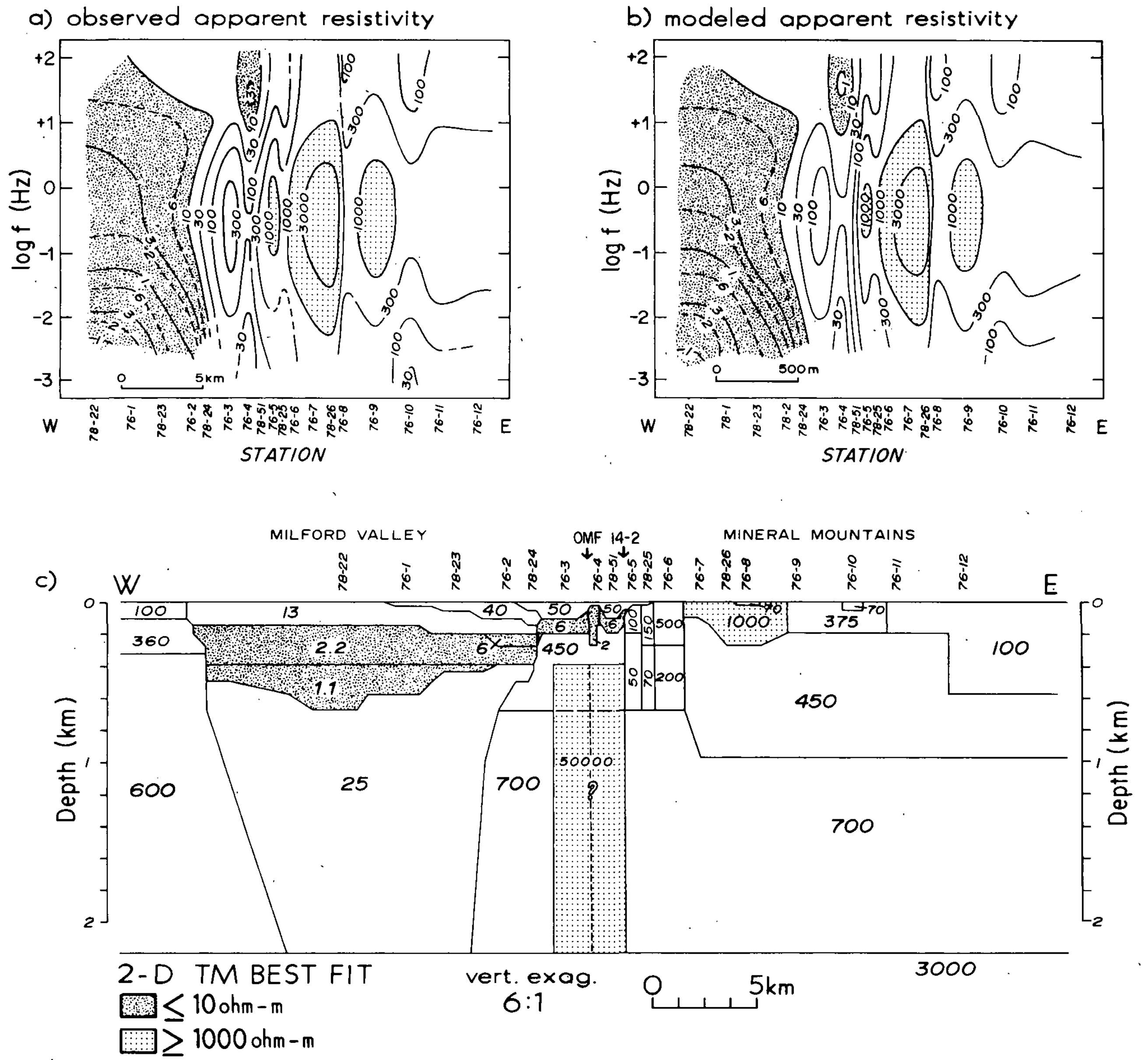

FIG. 10—Magnetotelluric profile CC′: (a) observed apparent resistivity, TM mode; (b) modeled apparent resistivity, TM mode; (c) finite element model for profile CC′ giving best fit between computed and observed TM mode resistivities. Intrinsic resistivity values in ohm-m; OMF = Opal Mound fault. (After Wannamaker et al, 1980.)

capabilities.

5. A deep heat source for the geothermal system also has not been discerned by MT interpretation to this time.

Thermal Studies

The thermal characteristics of the Roosevelt Hot Springs area have been defined by measurements in 53 thermal gradient, water well, and exploration drill holes (Wilson and Chapman, 1980). Most of the 53 gradient holes bottomed at depths of 200 to 360 ft (60-110 m). The observed near-surface (30 to 230 ft; 10 to 70 m depth) gradients range from 6°C/km to 3,330°C/km compared to a Great Basin average of 35 to 40°C/km.

Thermal conductivity measurements are listed in Wilson and Chapman (1980) and Glenn et al (1980). Wilson and Chapman (1980) assign the following average conductivities (W/m/K) to principal rock types: quartz monzonite, 2.54; opaline sinter, 2.00; biotite gneiss, 2.00; alluvium, 1.64. Using the appropriate thermal conductivities and gradients for each hole, Wilson and Chapman have produced a detailed map of the near-surface heat flow associated with the geothermal system. The 400 mW/m^2 and 1,000 mW/m^2 areas determined from their study are indicated in Figure 11. The 400 mW/m^2 contour, approximately four times background, encloses an area of 22 mi^2 (57 km^2), whereas the 1,000 mW/m^2 contour encloses an area of 6 mi^2 (16 km^2) including the Opal Mound fault and most of the successful production drill holes. The anomalous surface heat loss for the system was calculated at 64 MW.

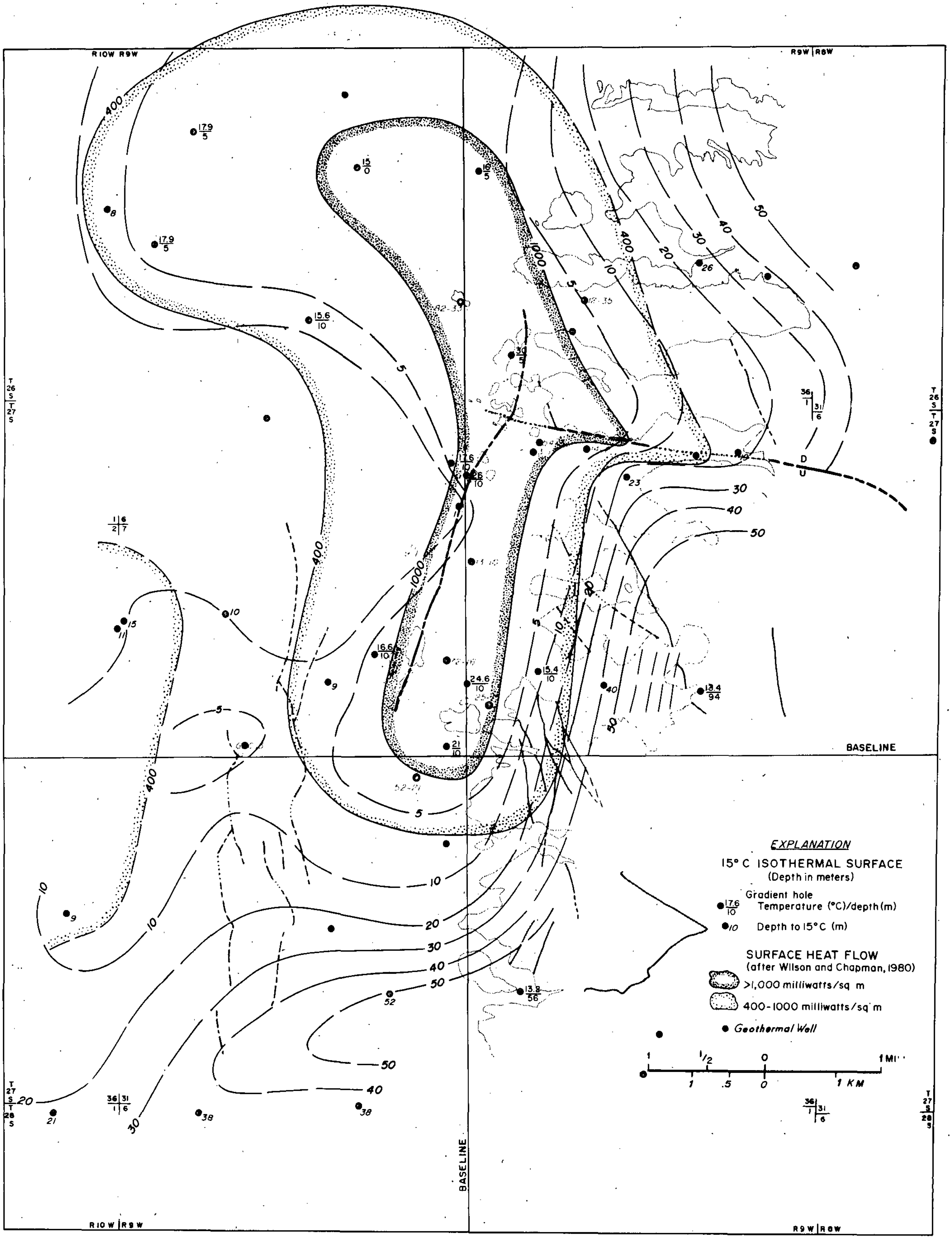

FIG. 11—Thermal studies map, Roosevelt Hot Springs KGRA.

Figure 11 compares the depth of the 15°C isothermal surface with the heat flow pattern of Wilson and Chapman (1980). Temperatures at a given depth are included for several holes. The 1,000 mW/m^2 area corresponds closely to the area where 15°C is mapped within 15 ft (5 m) of the surface. The steep contours along the flank of the range indicate depressed temperatures because of descending ground water, and a broad contour interval to the west and elongation to the northwest probably relate to the discharge of geothermal water within permeable strata in the alluvium. Although the contoured heat-flow map presents the thermal data in the most quantitative manner, the shallow occurrence of the high-temperature reservoir and its active discharge into the alluvium give rise to good definition of the system by contouring most thermal parameters.

RETROSPECTIVE EXPLORATION STRATEGY

The Roosevelt Hot Springs geothermal system has been used as a natural laboratory for the testing of exploration techniques. Using this relatively complete data base, it is possible to exercise 20/20 hindsight and define the most efficient strategy for the exploration of the Roosevelt Hot Springs geothermal system. Ward et al (1981) present a generalized exploration strategy which emphasizes the use of conceptual exploration models to solve specific exploration problems. The retrospective exploration strategy for Roosevelt Hot Springs involves five stages: literature study, geologic mapping, thermal gradient measurements, dipole-dipole resistivity, and drilling. The following strategy may not be applicable to other systems because the Roosevelt Hot Springs system is one of the hottest and largest hydrothermal systems in the Basin and Range province.

Phase I.—The first stage of the strategy would be a literature search and compilation. Review of existing geologic and hydrologic reports would have demonstrated that hot springs had been active in the recent past and would have provided chemical analyses of those waters. From those analyses, geothermometers could have been calculated to give an indication of possible reservoir temperatures. In addition, the presence of siliceous sinter is considered as evidence of a water-dominated geothermal system which has a base temperature of at least 356°F (180°C) (Renner et al, 1975). The presence of Quaternary rhyolites suggests the possibility of a high-level granitic pluton which supplies heat to the geothermal system. A review of the groundwater reports would allow plotting of potentiometric surfaces and chemical trends such as shown in Figure 3.

The approximate cost of such a literature search and compilation could vary widely depending on the expertise involved and the size of the region under consideration. The costs are thought to be between $5,000 and $20,000. Because the surface manifestations of the Roosevelt Hot Springs system are so well developed, a decision to initiate leasing activities could be made at the end of phase I.

Phase II.—This stage would involve the geologic mapping of the project area. Nielson et al (1978) found that mapping at a scale of 1:24,000 on an air-photo base was adequate for exploration purposes. The mapping should identify the extent of the hot-spring deposits and characterize their structural control. It would also identify the other structural systems within the geothermal area and allow a preliminary guess at the relative importance of the various fault systems in controlling the geothermal reservoir. Previous attempts at using air-photo interpretation alone to define these structural systems (Ward et al, 1978) were relatively unsuccessful.

The knowledge gained during this phase should be used to establish conceptual models of the geothermal system. In addition, the results will provide siting information for the subsequent thermal-gradient holes and electrical resistivity surveys.

The costs of the geologic mapping will probably be $15,000 to $25,000, one of the most cost-effective exploration options available.

Phase III.—The principal activities of phase III would be the delineation of the near-surface thermal anomaly and mapping of deeper resistivity structure. An area of 8 mi (12 km) north-south by 4 mi (6 km) east-west would be sampled by 25 to 30 shallow thermal-gradient holes drilled to depths of 300 to 600 ft (100 to 200 m). The deeper holes would logically be sited west of the Opal Mound fault where alluvium is predictably deeper and problems of near-surface water flow could be anticipated. This program would have detected temperature gradients in excess of 44°F/100 ft (800°C/km), and localized a high-temperature anomaly within 1.2 mi (2 km) of the Opal Mound fault. The cost of the survey, at 1980 prices, would have been between $130,000 and $160,000.

A slightly larger area, 10 by 5 mi (15 by 8 km), would have been selected for a dipole-dipole electrical resistivity survey. Cost-effective survey design, based on geologic mapping and in-progress temperature-gradient studies, would dictate perhaps 10 dipole-dipole lines of 1.8 to 3 mi (3 to 5 km) lengths, with perhaps 7 lines of 984-ft (300-m) dipole length and 3 lines of 492-ft (150-m) dipole length. The survey costs are estimated at $40,000 to $50,000, with selected numerical modeling estimated at an additional $10,000.

Without the prior knowledge of the valley-range transition, it would have been logical to plan several detailed gravity profiles in an attempt to define a range-front fault with major vertical offset. Such an effort would have cost $3,000 to $5,000, and would have contributed little to target definition.

It would be prudent to collect fluid samples from the gradient holes and to complete chemical analysis and geothermometric calculations on these samples. These costs would have added $1,000 to $2,000 to the exploration bill and provided valuable information on water type, mixing, and preliminary estimates of reservoir temperatures.

Phase IV.—After careful consideration of the geologic, thermal, and electrical data bases, the prudent exploration program would have sited two to four deep thermal-gradient holes, to depths of 1,000 to 2,000 ft (300 to 600 m). These could be expected to cost $15 to $18 per foot ($50 to $60 per meter). At Roosevelt Hot Springs these holes, if well chosen, might have intersected a major fracture at shallow depth (as in Utah State 72-16), or would at least indicate temperatures approaching 300°F (150°C) within 2,000 ft (600 m) depth. The program cost, with fluid collection and lithologic studies, is estimated at $100,000 to $200,000. A substantial effort in data interpretation and data integration would follow as a result of the strong encouragement for a high-temperature system. On the basis of this data analysis, production holes would be drilled.

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May 22, 1981

Mr. Paul D. James
Managing Editor, AAPG Bulletin
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Dear Mr. James:

My coauthors and I are pleased to submit the accompanying manuscript for your consideration for possible publication in the AAPG Bulletin. Dr. D. L. Nielson is scheduled to present a short form of this paper before the EMD Geothermal Session at the 1981 Annual AAPG meeting in San Francisco, June 2.

This manuscript, "An Integrated Case Study of the Roosevelt Hot Springs Geothermal System, Utah" by Ross, Nielson, Smith, Moore and Glenn was prepared under funding of the Department of Energy, Division of Geothermal Energy, but has not been distributed as a report. We would be pleased to consider format changes or minor revisions should the paper be favorably considered by your reviewers. Three copies of the text are enclosed to facilitate your review.

Sincerely,

Howard P. Ross

Howard P. Ross
Senior Geophysicist

HPR:jr

enclosure

cc D. L. Nielson

AN INTEGRATED CASE STUDY OF THE ROOSEVELT HOT SPRINGS GEOTHERMAL SYSTEM, UTAH

by

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ABSTRACT

The Roosevelt Hot Springs geothermal system is a hot water dominated, structurally controlled geothermal resource. Since 1974, the area has been undergoing intensive geothermal exploration and has been used as a natural laboratory for the development and testing of geothermal exploration methods by research organizations. This paper summarizes the geological, geophysical, geochemical, and hydrologic data which have been collected since 1974, presents an evaluation of the usefulness of each method, and concludes with a retrospective strategy describing the most effective means of exploration for the Roosevelt Hot Springs hydrothermal resource.

The bedrock geology of the area is dominated by metamorphic rocks of Precambrian age as well as felsic plutonic phases of the Tertiary Mineral Mountains intrusive complex. Rhyolite flows, domes, and pyroclastics reflect igneous activity between 0.8 and 0.5 million years ago. The structural development began with low-angle normal faulting which has produced low-angle, westward dipping cataclasites, steeply dipping, northwest-trending cataclasites, and brecciation localized in the hanging wall of the low-angle faults. East-west faulting is also present and has been produced by deep-seated regional zones of weakness. North to north-northeast trending faults are the youngest structures in the area, and they control present fumarolic activity which has deposited siliceous sinters. The geothermal reservoir is controlled by intersections of the principal zones of faulting.

The geothermal fluids that discharge from the deep wells are dilute sodium chloride brines containing approximately 7000 ppm TDS and anomalous concentrations of fluorine, arsenic, lithium, boron and mercury.

Geothermometers calculated from the predicted cation contents of the deep reservoir brine range from 271-277°C. Hydrothermal alteration by these fluids has produced assemblages of clays, alunite, muscovite, chlorite, pyrite, calcite, quartz and hematite.

Geochemical analyses of rocks and soils of the Roosevelt Hot Springs thermal area demonstrate that Hg, As, Mn, Cu Sb, W, Li, Pb, Zn, Ba and Be have been transported and redeposited by the thermal fluids. Hg, As, and Li are diagnostic of the hydrothermally altered rocks at depth. As and Li occur primarily along fracture zones and permeable horizons within the alluvium within the central portions of the thermal system. Hg, in contrast, is concentrated only within the outer, cooler portions of the thermal field to depths marked approximately by the 200°C isotherm. Knowledge of the distribution of these elements in soils and in thermal gradient and deep exploration wells can aid in prioritizing drilling targets, locating hydraulically significant fractures and estimating temperature distributions within the thermal system.

The geothermal system is well expressed in electrical resistivity and thermal gradient data and these methods, coupled with geologic mapping, are adequate to delineate the fluids and alteration associated with the geothermal reservoir. The dipole-dipole array seems best suited to acquire and interpret the resistivity data, although controlled source AMT (CSAMT) may be competitive for near surface mapping. Representations of the thermal data as temperature gradients, heat flow, depth to a given temperature, and temperature at a given depth are all useful in exploration of the geothermal system because the thermal fluids themselves rise close to the surface. Heat flow measurements were required, however, to quantify the energy potential and heat loss of the reservoir.

Self-potential, gravity, and magnetic survey data all contribute to our understanding of the system but are not considered essential to its exploration. Reflection seismic data, expensive on a cost per kilometer basis, provides substantial information on the structure beneath alluvial cover. Magnetotelluric data obtained and interpreted at considerable cost have to this date been unable to define deep resistivity structure associated with the system due to the complex 3-D near surface resistivity structure and to regional current gathering in the valley. A detailed study of geophysical well logs improved the lithologic logs and delineates porous and probable permeable zones corresponding to fractured zones within the crystalline rocks. The identification of fracture zones is instrumental in completion activities and flow testing.

INTRODUCTION

The Roosevelt Hot Springs geothermal system is located along the western side of the Mineral Mountains, approximately 19 km northeast of Milford, Utah (Fig. 1). The geothermal system is a high-temperature water dominated resource, and is structurally controlled with permeability localized by faults and fractures cutting plutonic and metamorphic rocks.

The earliest exploration in the area was carried out in 1967 and 1968 by Eugene Davies of Milford who encountered boiling water in two holes on and adjacent to siliceous sinter deposits. Phillips Petroleum Company initiated geothermal exploration at Roosevelt in 1972 and successfully bid for 18,871 acres of the Roosevelt KGRA (Known Geothermal Resource Area) at the July 1974 competitive lease sale (Forrest, 1980). Exploratory drilling began in 1975 and the second well drilled, PPC 3-1, encountered producible quantities of geothermal fluid. Additional drilling by Phillips Petroleum Co. and Thermal Power Company has brought the total number of successful wells in the Roosevelt Hot Springs geothermal field to seven. Present development plans call for one 20 megawatt (MWe) power plant to be on line by 1982 with subsequent expansion to perhaps 120 MWe. A discussion of the history of the Roosevelt Hot Springs KGRA and the exploration methods employed by Phillips Petroleum Co. is presented in Forrest (1980).

High temperatures, large flow rates and shallow reservoir depths make Roosevelt Hot Springs one of the most promising water dominated geothermal systems for near-term electric power production. Because of this a widespread interest has developed in the technical community and numerous studies have been carried out in the Roosevelt Hot Springs - Mineral Mountains area

(McKinney, 1978; Ward et al, 1978).

A large amount of geological, geophysical, geochemical, drilling, and reservoir data for the Roosevelt Hot Springs area has been made available in 1977-1979 through the DOE/DGE Industry Coupled Program. A review of these data and earlier work helped to identify inconsistencies and weak areas in an already large geoscience data base. A need for detailed structural and lithologic mapping seemed foremost, and this work was recently completed and reported (Nielson et al, 1978; Sibbett and Nielson, 1980a). Lithologic, geochemical and well-log studies have been completed for several drill holes and a detailed review of geophysical data was undertaken to expand upon the new geologic mapping. The integration of these data constitute the present case study.

ROOSEVELT HOT SPRINGS GEOTHERMAL SYSTEM

Regional Setting

The Roosevelt Hot Springs KGRA is located on the western flank of the Mineral Mountains (Fig. 2). The Mineral Mountains are a north-trending horst bounded by Basin and Range normal faults and lie at the western edge of the transition zone between the Colorado Plateau and the Basin and Range Physiographic Provinces. The area is located on the western edge of the Intermountain Seismic Belt as defined by Smith and Sbar (1974). In addition, the Roosevelt Hot Springs KGRA lies along east-west trending magnetic anomalies which follow the trend of the Wah Wah-Tushar mineral belt (Hilpert and Roberts, 1964; Mabey et al, 1978). This belt has been the site of intrusive activity through the Tertiary and into Quaternary time. Associated with this igneous activity are deposits of uranium and base and precious metals.

The central portion of the Mineral Mountains is a structural high relative both to adjacent ranges, and also to the northern and southern portions of the range. In these northern and southern areas sedimentary rocks of Cambrian through Cretaceous age are exposed. The southern area also contains volcanic and plutonic rocks of Tertiary age (Earll, 1957). In contrast, the central portion of the Mineral Mountains contains Precambrian metasedimentary rocks and Tertiary plutonic rocks of the Mineral Mountains intrusive complex (Sibbett and Nielson, 1980b). These Tertiary rocks possibly represent plutonic equivalents of the Marysvale volcanic province which is exposed to the east, south and southwest of the Mineral Mountains.

Local Setting

The geology in the vicinity of the Roosevelt Hot Springs geothermal system has been described in detail by Nielson et al (1978) and the geology of the central portion of the Mineral Mountains has been mapped by Sibbett and Nielson (1980a). The following is a summary of that study, and the reader is referred to those publications for a more detailed discussion of the lithologic units. A simplified geologic map of the Roosevelt Hot Springs area is shown in Figure 2.

The oldest unit exposed in the area of the geothermal system is a banded gneiss which was formed from regionally metamorphosed quartzo-feldspathic sediments. The rock was metamorphosed to the upper amphibolite facies during Early Proterozoic. The banded gneiss is strongly layered with adjacent layers distinguished principally on the content of mafic minerals. The rock is compositionally heterogeneous and contains thick sequences of quartzo-feldspathic rocks, which in drill cuttings resemble dikes of Tertiary intrusive units. The unit also contains metaquartzite and sillimanite schist layers which have been differentiated in the more detailed geologic study (Nielson et al, 1978).

The Mineral Mountains intrusive complex (Sibbett and Nielson, 1980b) is the largest intrusive body exposed in Utah. K-Ar dating and regional relationships suggest that the intrusive sequence is Middle-to Late-Tertiary in age. In the vicinity of the geothermal system, the lithologies range from diorite and granodiorite through granite and syenite in composition (Fig. 2).

Rhyolite flows, pyroclastics, and domes were extruded along the spine of the Mineral Mountains 800,000 to 500,000 years ago (Lipman et al, 1978). Discussions of the petrology and petrochemistry of these rocks have been presented by Evans and Nash (1978). The flows and domes are glassy phenocryst-poor rhyolites. The pyroclastic rocks are represented by air-fall tuff and non-welded ash-flow tuffs. Smith and Shaw (1975) hypothesized that young rhyolites such as these are indicative of an upper-level magma chamber which could serve as a heat source for geothermal systems.

Hot spring deposits in the the vicinity of the geothermal system have been mapped as siliceous sinter, silica cemented alluvium, hematite-cemented alluvium, and manganese-cemented alluvium. The principal areas of hot spring deposition are along the Opal Mound fault and at the old Roosevelt Hot Springs. In both of these areas the deposits consist of both opaline and calcedonic sinter. A detailed discussion of the phases associated with the sinter has been presented by Parry et al (1980) and summarized in Ward et al (1978).

The geothermal reservoir at Roosevelt Hot Springs is structurally controlled. The controls are thought to be produced by the intersection of the faults mapped in the KGRA (Nielson et al, 1978, 1979). The structural evolution of the Roosevelt Hot Springs geothermal system is envisioned as follows. During rapid uplift of the Mineral Mountains structural block, westward-dipping low-angle normal faults formed. The most important of these is the Wild Horse Canyon fault (Fig. 2). These types of features are termed denudation faults (Armstrong, 1972). This tectonic event produced intense zones of cataclasis both along low-angle fault planes and within the hanging wall of the principal fault block. The cataclastic zones within the hanging wall are steeply dipping zones up to 3 m wide which strike generally to the northwest. These zones in the hanging wall were produced by internal brecciation and interaction between rigid blocks during the low-angle faulting.

East-west trending high angle normal faulting cuts the low-angle faults. The best developed examples of this fault trend can be seen along Negro Mag Wash (Fig. 2). The trend is parallel to the Wah Wah-Tushar mineral belt and is probably related to movement along this deep-seated structural trend.

The Opal Mound fault and parallel structures are north-northeast-trending faults which are the youngest structures in the area. They localize siliceous hot spring deposits and are often marked by zones of alteration and silicification of alluvium.

Hydrothermal System

The limits of the hydrothermal system have been partially defined by deep drilling within the Roosevelt Hot Springs KGRA. The principal wells are shown on Figure 2, and important facts about the wells are presented in Table 1. The area of geothermal production is bounded on the east by the range front of the Mineral Mountains. On the west, the system is bounded by the Opal Mound fault, and on the south it terminates between 72-16 which is a producer and 52-21 which is a hot but dry hole. The system is presently open to the north.

The deep wells confirm that the host rocks for the hydrothermal system are the Tertiary plutonic rocks and Precambrian metamorphic rocks which have been mapped in the adjacent Mineral Mountains (Fig. 2). Thus the rocks show little primary permeability and the system is controlled by faults and fractures (Lenzer et al, 1976; Nielson et al, 1978, 1979). In addition, the compositions of the host rocks are not compatible with the development of secondary permeability by solution processes. The wells located outside of the producing zone such as 52-21 and 9-1 are hot, but because of low permeability, they do not produce geothermal fluids.

FLUID CHEMISTRY. The hydrothermal fluids are relatively dilute sodium chloride brines which contain approximately 7000 ppm total dissolved solids. Capuano and Cole (in prep.) provide the most comprehensive review of the chemistry of these fluids. Table 2 lists the composition of fluids collected from the wells and springs. These fluids are compositionally similar throughout the field, and differ mainly in their concentrations of calcium, magnesium and bicarbonate which may be the result of mixing with calcium-rich nonthermal ground water.

Reservoir temperatures estimated from the Na-K-Ca contents of the brines range up to 290°C, and exceed the measured maximum temperatures by as much as 20-30°C. Capuano and Cole (in prep.) suggest that temperatures calculated from geothermometers may in fact be too high, reflecting changes in the fluid chemistry produced by flashing of the brines in the well bore. Their arguments are supported by geothermometer calculations based on the composition of the deep reservoir fluid determined from the combined analyses of brine, steam and gas. These calculations suggest a reservoir temperature of 271°C based on the Na-K-Ca geothermometer (Fournier and Truesdell, 1973) and a temperature of 277°C based on the SiO_2 content of the fluid and the conductively cooled quartz saturation model of Fournier (1973).

The isotopic composition of the geothermal fluids indicates that they are of meteoric origin (Rohrs, 1980). A comparison of the fluid isotopic compositions with that of water from the mountain ranges to the east suggests that the thermal fluids could be derived either from the Mineral Mountains, the Tushar Mountains, or both.

HYDROTHERMAL ALTERATION. Hydrothermal alteration in the geothermal system and the adjacent Mineral Mountains is localized along faults and fractures. There have been several periods of alteration and it is often not possible to separate these different periods on the basis of their mineralogy and chemistry. Older periods of hydrothermal alteration are associated with the intrusion of various phases of the Mineral Mountains intrusive complex. These episodes have produced minor copper mineralization which is generally associated with xenoliths of Paleozoic and Precambrian rocks. In addition, zones of sodium metasomatism have been identified with the contact zones between some of the felsic plutonic rocks (Sibbett and Nielson, 1980a). The intrusion of the various phases of the pluton have also superimposed contact metamorphic assemblages on the older rocks. These contact assemblages are not restricted to the fault and fracture zones as are the hydrothermal alteration assemblages.

A hydrothermal event which altered the fault zones and deposited chalcopyrite in some places occurred with the low-angle faulting. The hydrothermal alteration produced assemblages of quartz + chlorite + epidote + hematite. Hematite is often found as specularite veinlets, and where genetic relationships can be observed, hematite mineralization follows sulfide mineralization.

The hydrothermal alteration assemblages associated with the present geothermal system are crudely zoned with depth. The uppermost assemblage, occurring around the hot spring deposits and fumeroles, is characterized by quartz, alunite, kaolinite, montmorillonite, hematite, and muscovite. Parry et al (1980) have studied the near-surface alteration and suggest that these minerals have formed above the water table by downward percolating acid sulfate waters. Upward convecting geothermal brines have produced, with increasing depth, alteration assemblages characterized by montmorillonite + mixed layer clays + sericite + quartz + hematite; and chlorite + sericite + calcite + pyrite + quartz + anhydrite (Ballantyne, 1978; Ballantyne and Parry, 1978; Glenn and Hulen, 1979; Parry et al, 1980; Capuano and Cole, in prep.). Thermochemical calculations and petrologic observations suggest that the brines are in equilibrium with the alteration assemblages produced by the upward migrating fluids (Capuano and Cole, in prep.).

FLOW TESTS. Most of the flow test, production logging and reservoir engineering data for the Roosevelt Hot Springs field is proprietory. However, data from Utah State 14-2 and Utah State 72-16 (Fig. 2, Table 1) is in the public domain and can be discussed to give some indication of the production capacity of the geothermal system.

A short flow test (no details available) was conducted in 72-16 on December 30, 1976. The test produced 454,546 kg/hr steam and hot water at a flowing wellhead pressure of 25 Kg/cm^2 and a temperature of 222°C.

A 24 hour test was conducted April 4 to 6, 1977 in 72-16. The total mass flow rate was determined by the James' method (James, 1966; Karamarakar and Cheng, 1980) as 595,000 kg/hr at a wellhead pressure of 20.7 Kg/cm^2 psia and a temperature of 213°C. However, the mass flow rate and wellhead pressure dropped throughout the test indicating a longer flow test was needed to determine steady state well production. Thermal Power Co. also concluded that the James' method of determining the flow rate of a two-phase flow was unsatisfactory. On the basis of the longer test, it was calculated that 72-16 could yield 119,546 kg/hr of steam if flashed at 5.6 Kg/cm^2. The well's electrical generating capacity was determined to be 12.5 Mw at a heat rate of 9,546 kg steam/hr/Mw.

Thermal Power Co. conducted a 48 hour flow test in Geothermal Well 14-2 between November 16, 1976 and November 18, 1976. The last 35 hours of the test stabilized at 225,000 kg/hr and an enthalpy of 444.5 BTU/hr. Thermal Power Co. calculates that at a wellhead pressure of 4.9 Kg/cm^2 and 17.8% flash, 14-2 has a generating capacity of 4.5 Mw. This capacity was calculated at a heat rate of 8,900 Kg steam/hr/Mw. Well surges were observed during the test and their cause was attributed to temporary obstructions in the system that induced in-pipe flashing.

The Denver Research Institute (DRI) conducted several production logging tests in 1978 and 1979 (Butz and Plooster, 1979). All tests were troubled by equipment failures, including the final test in May, 1979. Pressure and temperature data were logged versus depth during several constant flow-rate tests from 146,818 kg/hr to 263,636 kg/hr as measured by the James method. The flow rates are probably accurate to ± 15%.

The flow data were analyzed using a program that modeled two-phase flow in a well bore. By using this program, Butz and Plooster (1979) were able to match the measured borehole temperature and pressure profiles and to determine the flash depth in the borehole and the productivity index of the well. The best models were those that included a component of compressible gas.

EXPLORATION METHODS

The Roosevelt Hot Springs geothermal area has served as a laboratory for the development and testing of geothermal exploration methods for some time. Ward et al (1978) summarized the results of this work up to the time of their publication. Since 1978 more exploration data has become available including detailed geologic mapping, additional deep drill holes, additional electrical resistivity surveys, a reflection seismic profile, passive seismic data, and extensive application and testing of surface and subsurface geochemical methods. These new data allow us to expand upon the interim study of Ward et al (1978). The results of the individual exploration methods will be described, and, with all of this data available and the supreme advantage of

20-20 hindsight, the most efficient strategy for the exploration of the Roosevelt Hot Springs geothermal system will be presented. The results of some of the geological and geochemical exploration methods have been presented in the previous section.

Ground Water - Flow Systems and Chemistry

In contrast to the geothermal reservoir, the ground-water aquifer in the vicinity of the Roosevelt Hot Springs system is non-confined. The study of this ground-water system is important from two aspects. First, it is probable that the present ground-water system provides the recharge to the thermal reservoir. This information bears on estimates of the longevity of the system and also whether the system will remain water-dominated during production or become vapor dominated (White et al, 1971) or possibly depleted. Second, studies of ground-water flow and chemistry are important exploration procedures which can be used to detect leakage of thermal waters into the regional ground-water systems. Hydrologic data can be inexpensively collected during thermal-gradient and heat-flow studies and for many areas are available in the ground-water literature.

Reconnaissance ground-water studies of the Milford and Beaver Valleys (Mower and Cordova, 1974; Mower, 1978) do not discuss the hydrology in the geothermal area but they do reveal the generalized ground-water configuration. Figure 3 compiled from these reports is a map of the potentiometric surfaces of the alluvial aquifers in the two valleys. Figure 3 reveals the paucity of published water-level data near the KGRA. The direction of ground-water movement within the KGRA is likely to be west-northwest towards Beaver Bottoms.

The much greater elevation of the potentiometric surface in the Beaver Valley than in the Milford Valley, more than 300 m near the KGRA, prompts the speculation that if there were sufficiently transmissive fractures within the crystalline units of the Mineral Mountains, the difference in head would force ground water to flow through the range. The chances for this interbasin ground-water flow are enhanced by the cross-cutting structural zones like the Negro Mag fault zone. If ground water originating in the Beaver Valley drainage is recharging the geothermal reservoir, vast quantities of recharge may be available to the geothermal wells. Nearby examples of interbasin ground-water flow can be seen in the Snake Range (Hood and Rush, 1966) and Wah-Wah Mountains (Stephens, 1974).

To assess this possibility, Smith (1980) designed a finite-element model of the Mineral Mountains-Beaver Valley-Tushar Mountains hydrogeologic system. He concluded that unless there is a pronounced anisotropy in hydraulic conductivity (horizontal $\geq$ 10 vertical) in the granitic rocks of the Mineral Mountains it is unlikely that ground water flows through the range. Such anisotropy could readily be provided by east-west structural zones such as the Negro Mag fault. Smith's model is necessarily simple and does not account for thickness of the Tertiary valley fill in the Beaver Valley which enhances the possibility of recharge from the Tushar Mountains.

The ground-water papers by Cordova (1974) and Mower (1978) contain additional information which can be of value in exploration for the geothermal resource. Figure 4 is a map showing concentrations of Cl and B in the vicinity of the Roosevelt Hot Springs. Lenzer et al (1976) show plots of ground water TDS which are similar to the patterns developed in Figure 4.

These plots indicate that the chemical patterns produced by leakage of geothermal fluids into the groundwater system can be detected at significant distances from the geothermal system. This data is often available in the published literature and its compilation and analysis can be a very efficient exploration method.

Solids Geochemistry

The geochemistry of hydrothermally altered rocks is routinely used by the minerals exploration industry as a guide to buried ore deposits. Yet despite the similarities between hydrothermal mineral deposits and active geothermal systems, the development of geochemical zoning models by the geothermal industry has been largely neglected. Recent studies of the Broadlands field in New Zealand (Browne, 1971; Ewers and Keayes, 1977) and of Steamboat Springs, Nevada (White, 1980) have suggested that trace element zoning may be developed around the high-temperature centers of active geothermal systems. The studies at Broadlands have demonstrated that volatile and base metals are crudely zoned with depth as a result of decreasing temperature and boiling in permeable horizons.

The trace and major element content of surface and drill-hole samples from the Roosevelt Hot Springs thermal field were documented to determine if geochemical zoning models could be developed into an effective exploration tool (Bamford, 1978). Thirty-four elements were analyzed using an inductively coupled argon plasma spectrometer, arsenic was determined by colorimetric methods, and mercury by a gold film detector. The analytical techniques are described by Christensen et al (1980a) and Capuano and Bamford (1978). Detailed interpretations of the data from the Roosevelt Hot Springs thermal area have been presented by Christensen (1980), Christensen et al (1980b, in prep.), Bamford et al (1980) and Capuano and Moore (1980).

Geochemical analyses demonstrate that mercury, arsenic, antimony, manganese, copper, cobalt, tungsten, lithium, lead, zinc, beryllium, strontium and barium have been transported and redistributed by the thermal fluids. These elements are strongly concentrated in hot spring deposits. Surface concentrations are typically more than an order of magnitude higher than the subsurface abundances. However, with the exception of mercury, arsenic, antimony and lithium the trace element abundances of the drill cuttings cannot in general be unambiguously related to chemical redistribution by geothermal activity.

Attempts to enhance the geochemical signatures of the altered rocks at depth initially included the analysis of a high specific gravity (>3.3 g/cm^3) fraction separated from the drill cuttings. The purpose of preparing these samples was to concentrate sulfide minerals deposited by the thermal fluids. Because of the extremely low content of sulfide minerals in the altered rocks and the difficulty in mechanically separating sulfide minerals from drill steel and non-magnetic heavy minerals, routine analysis of the high specific gravity concentrates was discontinued.

Trace element distributions, like the distribution of hydrothermal alteration minerals, are not pervasive but are localized along fractures and permeable alluvial horizons that have served as fluid channels.

Detailed lithologic and geochemical logging has shown that at depth, arsenic is diagnostic of the hydrothermally altered rocks in the central portions of the thermal field. Electron microprobe analyses and selective chemical leaching of the drill cuttings indicate that arsenic occurs primarily within pyrite and crystalline iron oxides formed from pyrite. Concentrations of arsenic as high as 4% have been detected in some pyrite crystals. In general, concentrations of arsenic are irregularly distributed within individual pyrite grains suggesting that the composition of the thermal brines has varied during the life of the system.

Mercury, in contrast to arsenic, is concentrated in the cooler portions of the thermal field, occurring in both weakly- and highly-altered rocks. Concentrations of mercury greater than 20 ppb define a broad envelope in the outer portions of the system to depths approximately marked by the 215°C isotherm (Fig. 5). This distribution reflects the extreme mobility of mercury at high temperatures within the thermal system.

Christensen et al (1980b) have experimentally investigated the mobility of mercury by measuring its progressive loss from drill cuttings and surface samples heated in air. These studies demonstrated that by 200°C mercury loss had become significant and had reached a maximum by 250°C. They concluded that mercury is present within the geothermal reservoir mainly as a native metal and suggested that its distribution reflects the present thermal configuration of the geothermal system.

At depths between 30 and 60 meters pronounced enrichments in both arsenic and mercury occur in cuttings from thermal gradient holes located within the productive portion of the thermal field. Despite the relatively small number of samples and the wide spacing between drill holes, geochemical data from shallow thermal gradient holes nevertheless appear to be a useful means of prioritizing drilling targets that is independent of temperature measurements.

Concentrations of mercury and arsenic in soils over the thermal system are closely associated with hot springs deposits and faults that are connected to the geothermal reservoir. The distribution of the anomalies and their shapes confirm that northeast- and west-northwest-trending (Negro Mag) structures control the near-surface hydrology of the thermal fluids. Nielson et al (1978) reached a similar conclusion based on the distribution of faults mapped on the western flank of the Mineral Mountains. However, the geochemical anomalies suggest that faulting may be more prevalent in the alluvial covered portions of the thermal area than has previously been recognized (Capuano and Moore, 1980). Radon emanometry (Nielson, 1978) has also been demonstrated to be very useful in delineating faults covered by alluvium.

Hot spring deposits formed over the high temperature portions of the thermal system consist mainly of siliceous sinter. These deposits are enriched in mercury, arsenic, antimony, and beryllium (Table 3). In contrast, manganese-cemented alluvium located on the margin of the thermal system contains highly anomalous concentrations of manganese, barium, copper, lead, cobalt, arsenic, and tungsten. The highly oxidized nature of the manganese, absence of siliceous sinter, and high trace element contents, suggests that this deposit may have formed by cooling and oxidizing brines flowing away from the high temperature portion of the thermal system. Major concentrations of mercury also occur around fumaroles which represent degassing of boiling thermal brines beneath the surface.

Geophysics

GRAVITY METHODS. Gravity methods are widely used in geothermal exploration to map structure and, in some cases, to directly detect the depositional products of hydrothermal systems. The latter use has been demonstrated in the Imperial Valley of California (Biehler, 1971) where the precipitation of silica and carbonates in sediments above the hydrothermal systems results in an increase in density of the sediments and thus a positive gravity anomaly.

Gravity data from the Roosevelt Hot Springs KGRA and vicinity is presented in Ward et al (1978), Carter and Cook (1978), Crebs and Cook (1976), and Brumbaugh and Cook (1977). Figure 6 is a complete Bouguer gravity map compiled from these sources and superimposed on a simplified map of the geology of the Mineral Mountains.

The bouguer gravity map is dominated by two major features: a broad, closed minimum centered over Milford Valley, and a north trending elongate high over the Mineral Mountains. The data are fully terrain corrected - these features reflect a density contrast of 0.3-0.7 between valley fill and the bedrock (Table 4) within the range. The gravity relief is approximately 36 mgal. A generally planar regional gradient of approximately 1 mg/km decreasing eastward may be estimated for the Mineral Mountains area based on regional data (Crebs and Cook, 1976).

Bouguer gravity decreases gradually from the central portion of the range to the valley low. The gravity high along the northern and western portion of the Mineral Mountains can be correlated with the Paleozoic limestones, Precambrian metasediments, and Tertiary diorites. In the vicinity of the geothermal field, denser lithologies such as the Precambrian banded gneiss and Tertiary diorites have been intruded by less dense felsic phases of the Mineral Mountains intrusive complex. Local perturbations on the smooth gradient (residuals), generally less than 2 mgal, can be readily explained by density variations of 0.1-0.3 gm/cm^3 within the crystalline rocks, as documented in Table 4. The interpretation of structural details beneath the alluvium within and west of the geothermal field is within the range of ambiguity due to non-linear components of the regional gradient and the variable density within the bedrock. A principal result of model studies supported by well control and density data is the absence of a large (>200 m) displacement in the bedrock surface along any single normal fault. Instead we see a gradual dip to the west, and possibly several minor faults near the Opal Mound fault and the outcropping range front. Ward et al (1978) present a similar interpretation, and these interpretations are in agreement with reflection and refraction seismic data presented later.

<u>MAGNETIC METHODS</u>. The aeromagnetic map shown in Figure 7 is a portion, about 270 Km^2 of a much larger, about 780 Km^2 survey flown in 1975 (Ward et al, 1978; Brumbaugh and Cook, 1977). Data were obtained along east-west flight lines with an average line spacing of 420 m. The flight path was smoothly draped at an average 305 m above ground level.

An extensive program of magnetic susceptibility measurements was undertaken in the summer of 1979 to provide support for the magnetic interpretation. The susceptibility data, Table 5, are <u>in situ</u> susceptibility measurements at more than 60 locations on smooth, unweathered outcrop surface. A Bison magnetic susceptibility meter Model 3100 with <u>in situ</u> coil accessory was used for the measurements. The data shown in Table 4 are fully corrected for outcrop surface roughness. Additional susceptibility data determined from drill cuttings of well 9-1 are reported by Glenn et al (1980).

A detailed inspection of the map reveals about 30 closed highs and lows over outcrop of the Mineral Mountains. The apparent complexity of the aeromagnetic map is not surprising in view of the complex igneous geology (Fig. 2) and the unavoidable variations in terrain clearance over rocks of varying magnetization. A study of the analog altimeter profiles shows actual terrain clearance values as little as 150 m over sharp topographic highs and as much as 365 m over canyons and between hills. This terrain clearance variation contributes to considerable east-west elongation and irregularities in the contoured map.

An in-depth discussion of the entire magnetic map is beyond the scope of this more general paper. An interpretation has been completed with integrated spacial correlation of magnetic and geologic maps, simple depth estimates, model comparisons, and extensive magnetic susceptibility data. On this basis most of the magnetic features over the Mineral Mountain range can be attributed to mapped rock-type and altitude variations. Table 6 summarizes the main characteristics and interpreted sources for major anomalies identified in Figure 7. Most important to the present study is the interpretation of new information that relates to the structural setting of the geothermal reservoir.

Several east-trending zones of low magnetic intensity cut across the Mineral Mountains (L1, L2, L3, L4). These magnetic trends are explained in part by the outcropping rock types and topographic features. The topography, geology and magnetics are all expressions of east-west structural features. Feature L2 corresponds closely with the Negro Mag fault zone and its eastward continuation as mapped faults. It also appears to displace to the west a north-trending magnetic anomaly (7) closely associated with the Opal Mound fault. L3 terminates anomaly 7 between the productive reservoir well 72-16 and the hot-but-dry hole 52-21. The magnetic bodies appropriate to this magnetic field inclination are indicated on Figure 7.

The magnetic data over bedrock areas reflect the mapped geology and magnetic susceptibility. The identification of four east-trending structural zones, two of which (L1 and L3) may limit the north-south extent of the reservoir, is new information. The delineation of the alluvium covered Opal Mound horst bounded on the east by the reservoir bounding Opal Mound fault, is the most important contribution to understanding the geothermal reservoir provided by the magnetic data. In view of known susceptibility values for the igneous and metamorphic rocks and the alluvium it is not necessary to postulate an alteration low resulting from magnetite destruction to explain magnetic lows east of the Opal Mound fault (anomaly 8) and south of well Utah State 52-21 (anomaly 18).

SEISMIC METHODS. A broad spectrum of seismic data is available at Roosevelt Hot Springs. Passive seismic data includes long-term historical records of major earthquake activity, microearthquake surveys and at a lower magnitude of naturally occurring seismic disturbance, seismic emissions or "noise" surveys. Single profiles of both refraction and CDP "Vibroseis" reflection data are also available for study.

In 1974 and 1975 an array of up to 12 portable, high-gain seismographs was established within the Roosevelt Hot Springs and Cove Fort-Sulphurdale areas (Olson and Smith, 1976). One hundred sixty-three (163) earthquakes of magnitude $-0.5 < M < 2.8$ were recorded in two survey periods totaling 49 days. Most of the earthquake activity occurred as a series of swarms with shallow (less than 5 km) focal depths around the Cove Fort area, 20 km northeast of Roosevelt Hot Springs. Approximately 18 events were located along a north-south trend extending north of Milford along the west side of the Milford Valley. Only four events could be associated with a 20 km length of the western flank of the Mineral Mountains which includes the Roosevelt Hot Springs area (Ward et al, 1978). Olson and Smith (1976) determined P-wave delays of up to 0.1 sec and detectable S-wave attenuation of ray paths across the Mineral Mountains, Ward et al (1978) interpret the low-velocity effect and shear wave attenuation for these ray paths as possibly indicative of partial melting or major intense fracturing of crustal rocks beneath the southern part of the Mineral Mountains.

Robinson and Iyer (1979) observed P-wave residuals of up to 0.3 sec in a well defined pattern corresponding to a region of anomalously low velocity (5 to 7%) centered under the geothermal area and extending from about 5 Km depth down into the uppermost mantle. They prefer an explanation for these delays in terms of abnormally high temperatures and a small fraction of partial melt. Wechsler and Smith (1979) note that these delays could arise from the fractured and possibly fluid-filled porosity of the western portion of the Mineral Mountains pluton. Wechsler and Smith further note the problems of accurate epicenter location and the limitations of P-wave studies in areas of complex near surface lateral velocity variations, such as exist at Roosevelt Hot Springs.

The relatively small number of earthquake locations determined for the western Mineral Mountains from this 49 day recording period and the complex near surface velocity structure almost precludes the use of microearthquakes in delineating the Roosevelt Hot Springs geothermal system. P-wave delay studies may ultimately improve our understanding of the heat source at Roosevelt but to this time have not contributed to delineation of the system.

Schaff (1981) has reported on seismic activity detected with the present nine-station seismograph array for the period October 1979 through January 1981. His results to date substantiate the earlier characterization of Olson and Smith (1978) that seismicity in the immediate vicinity of Roosevelt Hot Springs is of a low level and somewhat episodic in nature. In the first 12 month period no earthquakes were located within the anticipated production zone or along the Opal Mound fault. Swarmlike activity was recorded east of the reservoir area in December 1980-January 1981. This trend of earthquake epicenters across the Mineral Mountains and east of the geothermal reservoir may lie along the eastern projection of the Negro Mag fault (Schaff, 1981).

Seismic Emissions Survey. Seismic emissions surveys have been promoted by several geophysical contractors as a geothermal exploration method in which the seismic emissions or "noise" would hopefully delineate active fault and fracture zones possibly associated with geothermal activity. The method employs an array of geophones (four or five) spaced approximately 610 m apart. In surveys at Roosevelt Hot Springs (Katz, 1977a; 1977b) data were recorded at the array for one to three days, then moved to another station. Five such stations within a 36 to 41 Km^2 area constitutes a survey. The data were edited and processed with algorithms which determine the noise source locations based on delay times computed for a half-space velocity model and the correlation of these delays with the observed data (Katz, 1977a; 1977b). This procedure was completed for a northern and a southern survey block at Roosevelt Hot Springs, Figure 8.

An interpretative composite anomaly map for the southern survey based on the distribution of correlation values greater than 50% of the maximum correlation suggests that more than 20 Km^2 of the 36 Km^2 area is anomalous (i.e., may be the source of seismic emissions). This broad area features: 1) a north-northeast trending zone which includes the Opal Mound fault and extends two miles east to the border of the correlation grid; and 2) areas to the southwest. A similar interpretation for the northern block indicates over 15 Km^2 as potential source areas for seismic emissions.

A reevaluation of these data (Ross et al, 1981) based on the upper 10 percent of maximum source correlations would reduce the anomalous area of the southern survey block to approximately five Km^2 including a coherent north-northeast trending zone three Km long and about 1000 m wide which is 1.6 Km east of the Opal Mound fault, Fig. 8. The correlation with the geothermal reservoir is obscure. In the northern survey block the anomalous area is reduced to less than five Km^2 with one coherent pattern well north of the known thermal area, and several small source areas located along the northern projection of the Opal Mound fault. A coherent area of emissions mapped only by Station 5 of this survey suggests a source area along the western 1000 m of Negro Mag Wash, an area which includes the producing wells Utah State 14-2, Roosevelt KGRA 3-1, and Roosevelt KGRA 54-3. This region, together with several small high correlation blocks along the northern projection of the

Opal Mound fault offer some support that interpretable seismic emissions are associated with the Roosevelt Hot Springs geothermal system.

As employed at Roosevelt Hot Springs the seismic emissions survey may indicate areas of geothermally induced seismic noise, but clearly records other noise sources and is imprecise in defining geothermal conduits. The correlation procedure is severely limited by model simplicity and velocity assumptions, and generally recognizes source direction more accurately than distance to the seismic noise source. A more refined velocity model could perhaps improve the resolution of the noise source areas through a higher correlation of source-to-geophone array delay times. It unlikely that the velocity model could be refined enough to justify inclusion of the method in geothermal exploration in complex geologic environments.

Seismic refraction. In April, 1977 a 30 km long seismic refraction profile was recorded across the Roosevelt Hot Springs geothermal area (Gertson and Smith, 1979). Multiple shots at seven different shot locations were used to provide multiple subsurface coverage. Although the large station spacings of nearly 250 m were not adequate for locating narrow structural features, Gertson and Smith (1979) were able to define a somewhat generalized velocity model for the area and also determined that the first large displacement in range front faulting occurs at least one km west of the Opal Mound fault. P-wave attenuation across the geothermal reservoir was much less than attenuation in other portions of the profile, and Gertson and Smith (1979) concluded that the record sections did not appear to contain any evidence for seismic waves that had penetrated and returned from 'hot rock' or magma chambers even at great depth. One complicating aspect of the refraction study is that the eastern portion of the refraction profile, of necessity, followed the Negro Mag fault zone from the reservoir area east across the Mineral Mountains.

Seismic reflection. One profile totaling 43 line km (27 miles) of detailed reflection seismic data was available for study of the Roosevelt Hot Springs area. Line 5 and 5 OPTW of a GSI speculation survey, recorded in March and April 1978, crosses the Mineral Mountains along the same path as the refraction survey. The data cannot be reproduced but an interpretation of the data is presented here.

The GSI profiles are high quality 24 fold CDP Vibroseis data with a 60 m (200 ft) group interval and a 120 m (400 ft) vibrator point interval. The source consisted of 16 sweeps per VP, using a 12 second sweep and a 12-60 Hz sweep band. The sample rate was 2 ms. Processing included deconvolution, velocity analysis, CDP stack, and migration.

Our interpretation of the reflection profile is presented in Figure 9. The time-to-depth conversion is supported by 14 velocity analyses along the profile. These velocities are generalized in Figure 9. The refraction survey and sonic logs of three deep wells in the geothermal system - Utah State 52-21, 72-16, and 14-2 also provide velocity control. Although similar velocities are noted for alluvium, the refraction survey and well logs indicate much higher velocities at depth than velocity analyses of CDP reflection data. To those depths where good reflection quality persists and the coherence of the velocity analysis is well supported, the reflection survey velocities are considered more valid. At greater depths (times) the velocity analyses appear to be less valid and may be as much as 40 per cent too low.

Figure 9 shows gently dipping layers in unconsolidated sediments to depths of 1220 m in the Milford Valley. This thickness thins rapidly to the east as the Opal Mound fault is approached. Numerous faults cut the sedimentary sequence. A very prominent reflector which dips westward at approximately 20 degrees from the Opal Mound fault indicates the base of the unconsolidated sediments. Few coherent reflectors are noted beneath this interface, presumably a sediment-igneous or sediment-metamorphic contrast.

The complexity of basin and range faulting is indicated by the reflection data. Unfortunately the trend of the profile is along the southern edge of the horst block indicated by magnetic data, then N30°E east along, and at a small angle to, the Opal Mound fault. The faulting is indicated, but less clearly than would be the case for an east-west line. East of the Opal Mound fault, the higher velocities and lack of coherent reflections indicate igneous rocks extending to depth. The interpretation of the vertical displacement along basin and range faults within the first 0.25 seconds is difficult because of high noise levels and probable lateral energy returns associated with the Opal Mound fault area. No single, major vertical displacement is indicated but a complex series of small displacements, of the order of 30-100 m is suggested by the data.

<u>ELECTRICAL STUDIES</u>. The recognition of Roosevelt Hot Springs as a relatively shallow geothermal resource of commercial potential has resulted in numerous electrical surveys designed to a) characterize the electrical resistivity distribution and b) test the effectiveness of various methods. To a large degree Roosevelt Hot Springs has become a "test" area because of the

type of resource and amount of publically available supporting data. Included in the electrical surveys already completed are the following: magnetotelluric (MT), controlled source audio magnetotelluric (CSAMT), natural source AMT, controlled source EM, spontaneous polarization or self-potential (SP), induced polarization (IP) and electrical resistivity.

Electrical resistivity surveys. Electrical resistivity surveys include both the dipole-dipole and bipole-dipole arrays. The bipole-dipole survey (Frangos and Ward, 1980) was undertaken with the knowledge of the resistivity distribution primarily to characterize the effectiveness of the method. The dipole-dipole work completed to date includes at least 50 lines of various lengths, electrode separations, and depth penetrations and constitutes one of the most complete resistivity data bases assembled. In view of this, and because all the other electrical methods are influenced by the resistivity distribution mapped by these data, a major effort was devoted to interpreting these data using 2-D numerical models (Ross et al, 1981). Much of the data were obtained as research projects and theses studies by the Department of Geology and Geophysics, University of Utah through funding of the National Science Foundation (Ward and Sill, 1976a) and the Department of Energy, Division of Geothermal Energy -- formerly ERDA (Ward and Sill, 1976b). A third survey was completed by Geonomics, Inc. under contract to Getty Oil Company (Katzenstein and Jacobson, 1976) and was made public through the DOE/DGE Industry Coupled Case Study Program.

The dipole-dipole array has been widely used in mineral and geothermal exploration. It is favored for its high lateral resolution and multiplicity of data along a profile which also permits interpretation in terms of depth.

The geometry of the array and the data plot or "pseudosection" are shown in Figure 10. The data are well suited to numerical modeling to arrive at an interpretation of the resistivity distribution both along the profile and at depth (Pelton et al, 1978).

Figure 10 illustrates several aspects of the dipole-dipole resistivity data and the interpretation used in this study. A more complete description of the interactive computer modeling process can be found in Ross et al (1981). The observed field data (lower values) and corresponding computed values for the final numerical model solutions (upper values) are shown for two resistivity profiles, G.O.C. lines 4 and 5 (Fig. 11). These are lines of 300m dipole length which trend northeast in the southern part of the thermal area. The lines intersect several east-west-trending profiles which are more nearly perpendicular to the geologic trend and resistivity structure. The numerical model which gives rise to the computed resistivity values is shown above each pseudosection. Figure 10 illustrates the agreement or "goodness of fit" between observed and modeled resistivity distributions. Vertical discontinuities between bodies of different resistivities often indicate faulting. Low resistivity zones of 10 and 5 ohm-m in the models for lines GOC 4 and GOC 5, respectively, correspond to conductive thermal fluids associated with the geothermal system which have flowed into the alluvium.

Figure 11 shows the location of all lines and presents a summary of the intrinsic resistivity distribution for the depth interval of 100-150 meters (0.3-0.6a, 300 m dipoles and 1.0-1.5a, 100 m dipoles). The interpreted resistivity sections are presented in Figure 12.

The resistivity distribution even as generalized by numerical modeling is quite complex. Generally high resistivities (100-500 ohm-m) are observed in the range over the Precambrian gneiss and Tertiary intrusive rocks. Very low (< 10 ohm-m) to low (10-20 ohm-m) resistivities often modeled as thin vertical conduits, occur along the trend of the Opal Mound fault. West of the Opal Mound fault moderate- to high-resistivity (30-400 ohm-m) layers overlie layers of moderate to very low resistivity. This appears to represent dry alluvium (above the water table) overlying areas where fresh ground waters mix with warmer, more conductive geothemal fluids migrating down dip from the Opal Mound fault and other conduits. Very low resistivities (2.5 ohm-m) observed along UU line 8100N may represent geothermal outflow, dissolved salts in Lake Bonneville sediments, or some mixture of both. South of the survey base line and GOC well 52-21 high resistivities near surface decrease with depth but do not indicate the presence of geothermal fluids noted to the north.

Figure 11 shows the interpreted resistivity distribution in map form for the depth interval 100-150 meters. The modeled resistivities at these depths have been transferred from Figure 12 and supplemented with a qualitative interpretation of more than 30 additional lines of dipole data. Figure 11 provides a better comparison with key geologic features and drill hole locations than does the section representation, Figure 12, and 11 shows that the low resistivity area of the geothermal system is bounded by high resistivities southwest, south and east of the geothermal system at these depths. Low and moderate resistivities extend well into the range along The Negro Mag fault zone. Most of the low-resistivity areas straddle or lie east of the Opal Mound fault at these depths. Resistivities of 3 to 10 ohm-m along the fault are bounded by 10-20 ohm-m on all sides as a crude zoning pattern. The resistivity of the alluvium is generally 20-50 ohm-m to the west, much higher (200-400 ohm-m) to the south. The probable continuity of resistivity zones is indicated by heavy dashed lines (Figure 12).

Ross et al (1981) present a similar map for the depth interval 450-600 m which is less complex because only 300 m dipole data could be used for this depth (450-600 m) and because of the resistivity averaging inherent in modeling large separation data. The high resistivities east of the Opal Mound fault extend to the west but are substantially reduced to the south (GOC Lines 1, 2, 4, 5). In fact, 50 to 400 ohm-m alluvium becomes quite conductive (5-20 ohm-m) with the increased depth. The Negro Mag Wash area is still a moderate resistivity re-entrant into the range. The alluvium west of the Opal Mound fault has become quite low in resistivity, 5-15 ohm-m, probably due to the downdip migration of conductive thermal fluids leaking from the geothermal system. The region of the Opal Mound fault itself is modeled as 30-50 ohm-m except at the southernmost portion of the Opal Mound itself (8 ohm-m). This is in marked contrast to the low resistivity expression at 100-150 m depth. A high resistivity (450 ohm-m) body is indicated less than 380 m south of successful well Utah State 72-16, which reached t.d. at 383 m. The hot but dry Getty Oil Co. well Utah State 52-21 is located in a resistive (100-400 ohm-m) area. Several productive wells Roosevelt KGRA 12-35, Roosevelt KGRA 54-3, Roosevelt KGRA 3-1, Utah State 14-2, Roosevelt KGRA 13-10, and Roosevelt HSU 25-15 are sited in areas of moderate (15-40 ohm-m) resistivity which are relative lows.

The complex resistivity distribution varies in three dimensions as shown by Figures 12 and 13. The dominant low resistivity material appears to be alluvium at depth where flushing with geothermal effluent or lake-bed clays gives rise to low resistivities.

The reader is reminded of the non-uniqueness of the interpreted resistivity distribution for a given profile of observed data. The interpretations presented in Figure 12 are limited by the grid size which becomes coarser with depth, the validity of the two-dimensional model, the goodness of fit of computed to observed data values, and the choice of body size, position and resistivity. Nonetheless, careful modeling of dipole-dipole resistivity data offers a more accurate representation of earth resistivity distribution for a given cost than any other electrical method. The non-uniqueness is further reduced by utilizing a network of profiles, several of which intersect. The integration of geologic data, such as the detailed (1:24,000) map of Nielson et al (1978) can further reduce interpretational ambiguities.

<u>Near Surface Electrical Methods</u>. Sandberg and Hohmann (1980) describe a controlled source audio-magnetotelluric (CSAMT) survey at Roosevelt Hot Springs which compares well with dipole-diople resistivity data for shallow depths (less than 100 m). Minor disagreements in the interpretative models from these methods serve to indicate the noise and ambiguity levels for each method. Chu et al (1979) describe induced polarization surveys which were unable to map clay alteration or pyritic zones related to the geothermal system.

Corwin and Hoover (1979) discuss thermoelectric coupling and electrokinetic coupling as possible mechanisms for generating self-potential anomalies observed over geothermal systems. They report a dipolar anomaly located directly over the Opal Mound fault and conclude that it results from geothermally generated electrical activity along the fault. More detailed self-potential studies have been completed by Sill and Johng (1980) and Sill (1981). A complex "quadrapolar" anomaly of two lows (-100 mv) and two highs (+25 and +50 mv) was found to be associated with the southern portion of the geothermal system. The 100 mv low which occurs over the Opal Mound fault is the most unambiguous expression of the geothermal system. Other self-potential features could arise from large resistivity contrasts, the movement of ground water from higher to lower elevation, or resistivity or fluid movement along structures (Sill, 1981). Additional numerical modeling is required to obtain a more complete understanding of the SP response at Roosevelt Hot Springs, but the observed data document a complex expression of the geothermal system.

Magnetotelluric (MT) Studies. The magnetotelluric (MT) method is routinely used in both the reconnaissance and detailed stages of geothermal exploration. The earth's electric and magnetic fields vary as a function of frequency in response to natural electrical (telluric) currents flowing within the earth's crust. Through precise measurements of the electric and magnetic field components made at the surface one may obtain information relating to the impedance distribution (i.e. electrical resistivity) to depths as great as 40 km within the earth's crust. The reader is referred to an excellent paper by Vozoff (1972) for a detailed description of the method.

In keeping with the detailed geophysical definition of the Roosevelt Hot Springs area an extensive network of MT stations was established in the Milford Valley-Mineral Mountains area. This work has been reported by Wannamaker (1978) and Wannamaker et al (1980). A summary of this complex topic is included here to add completeness to the case study. Twenty-five MT soundings were obtained in the central Mineral Mountains-Milford Valley area during the summer of 1976. A second survey of 50 stations was completed in the spring of 1978. Additional work in 1977 and 1979 brings the total number of MT soundings to 93. This discussion is primarily limited to the results of an east-west profile 22 km in length which extends from the Milford Valley on the west, crosses the Opal Mound fault and the heart of the geothermal system, and continues east across the Precambrian gneiss and Tertiary granitic rocks of the Mineral Mountains (Fig. 13).

The reduced data are presented as observed apparent resistivity versus frequency for the TM (transverse magnetic) mode in Figure 13a. The best fit model results computed for a two-dimensional model geometry, TM mode (Fig. 13b) are seen to be very similar. The comparison between observed and modeled impedance phase (not shown here) is also good but a poorer fit in the central portion of the profile. The two-dimensional model that produced this best fit is shown as Figure 13c.

Stations 78-22 through 76-2 are located over low resistivity alluvium and valley fill. Figure 13 indicates resistivities decreasing to near 1 ohm-m at depths of 400 m. Dipole-dipole resistivity data and Schlumberger soundings also indicate a thin (100 m) resistive (100 ohm-m) near surface layer with resistivity decreasing to 5 ohm-m or less at depths of 300 m.

Station 76-3 near the center of the cross section occurs over a buried horst inferred from gravity and magnetic data and exhibits high resistivities at moderate depths. Station 76-4 was sited along the Opal Mound fault (Wannamaker, 1978) which has been identified earlier as a narrow, vertical conductive zone associated with ascending geothermal fluids. Silica cementation has locally decreased the porosity but fracture permeability remains. Both the TE and TM mode observed resistivities and the 1-D "stitch" model reflect these geologic conditions throughout the entire frequency range. The rather fanciful 50,000 ohm-m prism in Figure 13 was inserted in an unsuccessful attempt to match the mid-frequency (10 to 0.1 Hz) data. A frequency dependent, 3-D current gathering effect involving the Milford Valley to the west is now believed responsible for these particular modeling difficulties (Wannamaker et al, 1980).

Stations 76-5 through 76-10 are located over Precambrian gneiss and Tertiary granitic-intrusives of the Mineral Mountains. These soundings record high resistivity values at high frequencies (shallower depth) which increase with lower frequencies and hence greater depths. Both TE and TM mode suggest resistivities of several hundred to over 1000 ohm-m. Two local low-resistivity zones occur at Stations 76-4 and 76-19. Wannamaker (1978) notes that a local low-resistivity zone of probable finite depth extent may give rise to low apparent resistivity values even at the lowest frequencies.

Wannamaker et al (1978, 1980) note great difficulty in obtaining good representations of both the TE and TM mode resistivities for any single one-dimensional or two-dimensional model. Extensive three-dimensional model studies (Wannamaker et al, 1980) clearly indicate the limitations of 1-D and 2-D modeling at Roosevelt Hot Springs, and probably for most Basin and Range type geothermal reservoir areas. A few of their more general conclusions should be restated here:

1. Current gathering in the valley results in a regional distortion of the electric field affecting all stations at Roosevelt Hot Springs for lower frequencies.

2. The TM mode is most appropriate for 2-D interpretation, and has yielded good results for geometrically regular 3-D prisms.

3. Clays (ρ = 1-2 ohm-m) may exist to depths of several hundred meters in the Milford Valley. These overlie more than 1 km of semi- and unconsolidated sediments and volcanics of moderate ($\rho \simeq$ 25 ohm-m) resistivity.

4. A geometrically regular reservoir of conductive brine beneath the thermal anomaly seems improbable, so the search for any economic hydrothermal reservoir at Roosevelt Hot Springs using MT must be considered unsuccessful to this time. If present, it is not resolved by the 2-D TM algorithm. The brine saturated reservoir zone is clearly 3-D and difficult to model satisfactorily with present interpretation capabilities.

5. A deep heat source for the geothermal system also has not been discerned by MT interpretation to this time. Uncertainties about the physiochemical state of this source (a hot but solidified magma chamber is not likely detectable by any electrical method) as well as the probable 3-D nature of its geometry again makes difficult its understanding with present modeling expertise.

<u>THERMAL STUDIES</u>. The thermal characteristics of the Roosevelt Hot Springs area have been reported by Sill and Bodell (1977), Ward et al (1978),

and Wilson and Chapman (1980). The study by Wilson and Chapman (1980) includes the results of 53 thermal gradient, water wells and exploration drill holes. Most of the 53 gradient holes bottomed at depths of 60-110 m. The thermal gradients reported by Wilson and Chapman (1980) generally correspond to uniform gradients within the 10-70 m depth range. The observed near surface gradients range from 6°C km^{-1} to 3331°C km^{-1} compared to a Great Basin average of 35°-40°C km^{-1}. Based on gradients and degree of curvature these temperature profiles were classified into three spatially consistent patterns. These patterns are interpreted as representing recharge in the Mineral Mountains, convective heat transfer above the geothermal system, and discharge of the thermal fluids into the alluvium to the west of the thermal system.

Thermal conductivity measurements are listed in Wilson and Chapman (1980) and Glenn et al (1980). Wilson and Chapman (1980) assign the following average conductivites (Wm^{-1} K^{-1}) to principal rock types: quartz monzonite, 2.54; opaline sinter, 2.00; biotite gneiss, 2.00; alluvium, 1.64. Using the appropriate thermal conductivities and gradients for each hole Wilson and Chapman have produced a detailed map of the near-surface heat flow associated with the geothermal system. The 400 mWm^{-2} and 1000 mWm^{-2} areas determined from their study are indicated in Figure 14. The 400 mWm^{-2} contour, approximately four times background, encloses an area of 57 km^2 while the 1000 mWm^{-2} contour encloses an area of 16 Km^2 including the Opal Mound fault and most of the successful production drill holes. The anomalous surface heat loss for the system was calculated at 64 MW.

Relatively few geothermal exploration projects have the benefit of such a detailed thermal study as noted above. More typically detailed exploration proceeds on the basis of contour maps of thermal gradients, depth to a given temperature, or temperature at a given depth. The latter two have little uncertainty resulting from the interpretation of complex gradients or gradient projections. Figure 14 compares the depth to the 15°C isothermal surface with the heat flow pattern of Wilson and Chapman (1980). Temperatures at a given depth are included for selected holes. The 1000 mWm^{-2} area corresponds closely to the area where 15°C is mapped within 5 m of the surface. The steep contours along the flank of the range indicate depressed temperatures because of descending ground water, while a broad contour interval to the west and elongation to the northwest probably relate to the discharge of geothermal water within permeable strata in the alluvium. Although the contoured heat flow map presents the thermal data in the most quantitative manner, the shallow occurrence of the high temperature reservoir and its active discharge into the alluvium give rise to good definition of the system by contouring several thermal parameters.

WELL LOG DATA AND ANALYSIS. Well logs from four drill holes (Utah State 52-21, Utah State 14-2 and Utah State 72-16, and thermal gradient hole GPC-15) have been studied by Glenn and Hulen (1979) and hole Roosevelt KGRA 9-1 by Glenn et al (1980).

A fairly complete suite of open hole logs, including electric, nuclear, acoustic, caliper and temperature logs, were obtained in all drill holes except in 72-16 where only the mud log, several temperature logs and a spontaneous potential-cement bond log in casing were obtained. Except for caliper logs, temperature effects do not appear to be severe in any of the holes, primarily because the hotter holes were circulated before logging. Log interpretation is difficult because the logs are calibrated for and recorded in oil field units whereas the rocks are igneous and metamorphic rather than sedimentary. Standard interpretation techniques can be used with satisfactory results, but these interpretations would be improved if adequate calibration data existed for these rock types. Glenn and Hulen (1979) note that calibration data are particularly needed for borehole compensated neutron and density tools and for the different types of neutron tools.

The hydrous mafic minerals, predominantly biotite and hornblende, strongly influence the neutron, density and acoustic log responses. The higher density and the contained water of these minerals are the principal properties influencing the log response. Trace element effects have been investigated in Roosevelt KGRA 9-1 and are found in concentrations that could cause only a 2-3 percent change in the thermal neutron capture cross section (Glenn et al, 1980). The unfractured rocks exhibit 100 to greater than 1500 ohm-meter resistivities on induction resistivity logs.

EVALUATION OF EXPLORATION TECHNIQUES

Geologic Mapping

The detailed geologic mapping at Roosevelt Hot Springs has been an important contribution to exploration for the geothermal resource. This data base includes data that stands alone in addition to information which contributes to the interpretation of geophysical and geochemical surveys. The following specifics outline the contributions of geologic mapping to resource discovery and definition.

Siliceous sinters mapped along the Opal Mound fault zone are considered _prima facie_ evidence of a water-dominated geothermal system which has a base temperature of at least 180°C (Renner et al, 1975). In the Roosevelt Hot Springs area, two major episodes of hot spring activity are evident, and both are controlled by the same structure, the Opal Mound fault. It is therefore evident that the Opal Mound fault has been an important controlling structure of the geothermal system for some period of time. The other area of hot spring deposition and present fumarolic activity is along the Negro Mag fault, particularly near the intersection of the Negro Mag and the Opal Mound fault zones. From the above evidence, drill holes could be sited to intersect either the Opal Mound or the Negro Mag faults or the zones of intersection of those faults. The dip of these structures cannot be well established from surface mapping.

Rhyolite flows, domes, and pyroclastics which are 500,000 - 800,000 years old are generally interpreted to indicate systems which are driven by a high-level granitic pluton which still contains much of its heat. At Roosevelt Hot Springs the area of extrusive activity is offset from the geothermal system and is characterized by thermal gradients which suggest recharge (Wilson and Chapman, 1980). It has not been possible from mapping the rhyolites to site drill holes. The presence of young extrusive rocks remains of importance on a district scale to focus exploration on a specific area.

Mapping has allowed for the characterization of lithologies and the delineation of their areal extent. This readily leads to the conclusion that none of the lithologies present show primary permeability and therefore, the system must be structurally controlled. The characterization of the lithologies provides an important data base for the logging of lithologies in exploration drill holes. From these logs and the geologic maps, cross sections can be constructed which contribute to structural models of the geothermal system.

Structural studies and the mapping of faults allows inferences to be made concerning the specific structures which control the geothermal system. This in itself would allow the siting of exploration holes. In a more prudent exploration program, this information should be utilized for the planning of heat-flow and resistivity surveys and geochemical surveys which have as their objective the delineation of prospective zones of permeability in the subsurface. From our structural interpretation, we also propose that there may be two geothermal reservoir zones at Roosevelt. The upper zone is defined by the structural style in the hanging wall of the Wildhorse Canyon fault. Below the Wildhorse Canyon fault the fluid flow paths are controlled by the high-angle structures of the Opal Mound and Negro Mag fault systems. Drilling targets in the footwall of the Wildhorse Canyon fault will obviously be more difficult to define than those in the hanging wall.

The geologic mapping contributes to the interpretation of the gravity and the aeromagnetic data. The correlation of gravity and aeromagnetic maps with the geologic map, coupled with density and magnetic susceptibility data, constitutes much of the interpretation of these data in bedrock areas. The geologic map has been used as the basis for physical property measurements.

The geologic mapping allows for an understanding of temporal and geometric relationships of periods of alteration. This is critical for understanding the trace element geochemical relationships and the interpretation of the alteration phases found in studies of lithologies from exploration drill holes.

Geochemistry

Studies of the Roosevelt Hot Springs thermal area have documented both geochemical and mineralogical zoning about the high-temperature portion of the thermal field. These studies show that the distributions of arsenic and mercury in even widely-spaced thermal gradient holes and soils can, for example, help prioritize drilling targets and delineate structures of hydrologic importance. During the later stages of exploration, the geochemistry of deep wells can aid in defining the margins of the thermal system and in mapping the distribution of fluid pathways at depth. The development of effective geochemical zoning models will be very important in the search for buried geothermal systems.

Mineralogical data can place constraints on the temperature and chemistry of the reservoir fluid and provide insight into the historical development of the thermal system. At Roosevelt Hot Springs, the alteration minerals appear to be in equilibrium with the calculated composition of the present deep reservoir fluids.

The composition of the deep reservoir fluid can be calculated from the analyses and relative abundances of brine, steam and gases collected at the wellhead. This calculated composition provides baseline data for monitoring chemical changes that may occur in the reservoir during production. Reservoir temperatures determined from the cation abundances of this model fluid are in close agreement with down-hole measured temperatures and are internally consistent. Because the composition of fluids sampled at the wellhead have been modified by flashing they do not accurately reflect the composition of the reservoir brine.

Geophysical Surveys

The number of different geophysical methods used and the amount of detail obtained substantially exceed the effort of a cost-effective, sequential exploration program. Much of this work has been oriented at an evaluation of the methods themselves rather than the delineation of the geothermal reservoir.

Dipole-dipole electrical resistivity and thermal gradient surveys best delineate the geothermal system. Drilling to depths as little as 20 meters has been shown by Wilson and Chapman (1980) to define three basic patterns of thermal behavior, one of which is closely associated with the reservoir area. Some interpretation problems were initially presented by the down-gradient migration of thermal waters within the alluvium. The discrete nature of drill hole data require integration of geologic or other continuous data to help pinpoint reservoir tests.

The dipole-dipole resistivity results, either as contoured by Ward et al (1978) or as modeled quantitatively in this study define an anomalous area almost identical to that of the thermal studies. Low-resistivity thermal water, probable wall rock alteration products and the shallow depth to the reservoir facilitate mapping and interpretation of the near-surface reservoir area, faults under alluvial cover, and the down-gradient movement of thermal waters. Low-resistivity values associated with Quaternary sediments in the Milford Valley are associated with different geometries and geology and should not be confused with the expression of the thermal system itself. A more limited resistivity program could have given rise to equivalent exploration results at Roosevelt Hot Springs.

Controlled source audio-magnetotelluric surveys (CSAMT) have been shown to produce a similar near-surface resistivity map to dipole-dipole surveys. The CSAMT (Sandberg and Hohmann, 1980) appeared to be cost competitive but was completed with the prior knowledge of the resistivity distribution which allowed a minimum number of well planned stations. The interpretation of detailed resistivity structure, particularly at depths of 300-600 m is considered more costly and tenuous than for dipole-dipole surveys. Bipole-dipole resistivity surveys (Frangos and Ward, 1980) are very sensitive to transmitting dipole location and lack the geometric resolution which is available in the dipole-dipole method. Induced polarization profiles (Chu et al, 1980) designed to map pyrite or clay distributions associated with the geothermal reservoir were not useful at Roosevelt Hot Springs and will generally be subject to the confusion of older, unrelated mineralization or alteration, and to other unrelated geologic effects (i.e. zeolite formation in volcanic rocks).

The magnetotelluric method has been used at Roosevelt Hot Springs in great detail (93 tensor stations). The result to this date of extensive computer modeling (Wannamaker et al, 1980) is a 2-D TM mode resistivity distribution which is consistent with the much less costly dipole-dipole model. No deep-seated (35-65 Km) resistivity feature definitely associated with the geothermal system could be interpreted because of the current gathering phenomena associated with the Milford Valley as well as other complex near-surface resistivity distributions. Given the present-day limitations in interpretation capabilities, the MT method cannot be considered cost effective at Roosevelt Hot Springs. The value of MT studies at Roosevelt Hot Springs lies in the new understanding of the method developed as a result of 2-D and 3-D numerical modeling of this extensive data base. The concerns and shortcomings of MT interpretation demonstrated here undoubtedly apply to many Basin and Range type geothermal systems, and to MT surveys in other complex geologic environments.

The gravity method is used extensively in geothermal and mineral exploration in the Basin and Range province, principally to define normal faults under alluvial cover and to estimate alluvial thickness. An interesting result at Roosevelt Hot Springs is the gradual westward dip of higher density bedrock beneath the alluvium of the east side of Milford Valley. There does not appear to be a major fault displacement of the buried bedrock surface near Roosevelt Hot Springs which could indicate alternate fluid paths for the geothermal fluids, but rather a series of minor displacements close to the Opal Mound fault. Density variations within the crystalline rocks exposed in the range are adequate to explain, or to confuse, the detailed interpretation of the gravity data. The gravity work then is interesting and contributes to our understanding of the resource area, but does not contribute to drill hole selection or reservoir definition.

The aeromagnetic survey, supplemented by ground profiles (Brumbaugh and Cook, 1978) defines several east-west trends which cut the Mineral Mountains and extend westward, roughly bracketing the reservoir area. One magnetic source beneath alluvial cover defines a lithologic or structural block which appears to limit the reservoir on the west. The magnetic data play a supporting role rather than a major role in delineating the geothermal resource.

The passive seismic methods have played a minor role at best in delineating the geothermal resource. The seismic emissions surveys were non-specific and microearthquake data have been limited, to date, by few events and the episodic nature of activity at Roosevelt Hot Springs. The after-the-fact location of a precise microearthquake monitoring network, and subsequent relocation of several stations within this network, will provide useful information on pre-production seismic activity. A recent earthquake swarm suggests an important role for an east-trending structural zone, probably the Negro Mag fault, in the plumbing and possibly the recharge of the reservoir. This again is after-the-fact information, but may contribute substantially to our ultimate understanding of the resource.

One detailed reflection seismic profile provides our best structural view of the Milford Valley and pediment area near the resource. The nearly uniform dip of the bedrock surface, crystalline or volcanic rocks, beneath the alluvium is well defined. The termination of several near-surface (0-600 m) reflections indicates numerous small displacement high-angle faults in the pediment area, many of which correspond to mapped delineations in alluvium or to the projection of bedrock faults. Unfortunately, the seismic profile bends and trends nearly parallel to the Opal Mound fault and adjacent reservoir area, precluding a detailed evaluation of this most important area. Few reflections would be expected in the crystalline rocks of this range, and no significant structural or lithologic information is interpreted.

A single refraction profile (Gertson and Smith, 1979) addresses regional structure and velocity data and does not contribute significantly to our understanding of the geothermal reservoir.

The detailed interpretation of geophysical well logs, though not strictly an exploration technique, is seen to be essential to the identification of production zones, well and formation conditions and to a proper interpretation of lithologic data determined from cuttings. In addition the resistivity, acoustic and density logs have contributed to a more complete interpretation of surface geophysical data.

RETROSPECTIVE EXPLORATION STRATEGY

The Roosevelt Hot Springs geothermal system has been used as a natural laboratory for the testing of exploration techniques. Using this relatively complete data base it is possible to exercise 20/20 hindsight and define the most efficient strategy for the exploration of the Roosevelt Hot Springs geothermal system and provide an update on the strategy suggested by Ward et al (1978). More general exploration strategies are presented in Ward et al (1981). The 1981 paper also outlines our philosophy of utilizing conceptual exploration models and application of methods to solve specific exploration problems. The retrospective exploration strategy for Roosevelt Hot Springs involves five stages: literature study, geologic mapping, thermal gradient measurements, dipole-dipole resistivity, and drilling.

A search of existing literature would indicate that hot springs had been active in the recent past and that young rhyolites were present in the area. A review of ground-water reports would allow the plotting of chemical trends such as shown in Figure 4.

Geologic mapping would identify the extent of the hot spring deposits and also characterize their structural control. It would also identify the other structural systems within the geothermal area and allow a preliminary guess at the relative importance of the various fault systems in controlling the geothermal reservoir.

Thermal gradient measurements should be acquired in intermediate depth holes. From these measurements large anomalous target areas will be defined. Heat flow determinations would allow a more quantitative evaluation of the thermal potential of the reservoir, but may not be essential to the exploration of the geothermal system when as close to the surface as it is at Roosevelt.

Since the Roosevelt system is structurally controlled, dipole-dipole resistivity would be used to define fluid bearing fault zones within the thermal anomalies. The resistivity data can be used to define the near surface trend of these zones and knowledge of their attitudes from the geologic mapping could be used to approximate their dips and allow for planning of drill holes to intersect these structures at an optimum depth.

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Table Captions

Table 1 - Well summary from Roosevelt Hot Springs, Utah.

Table 2 - Representative chemical analyses of selected thermal waters.

Table 3 - Analyses of hot spring deposits from Bamford et al (1980) and Christensen et al (in prep.). A dashed line indictes that the element was not detected.

Table 4 - Densities of lithologies from the Mineral Mountains.

Table 5 - Magnetic susceptibilities of lithologies from the Mineral Mountains.

Table 6 - Magnetic anomaly and source characteristics, Roosevelt Hot Springs Area, Utah.

Table 1. Well summary from Roosevelt Hot Springs, Utah.

Well	Location	Depth	Status	T Max	Operator	Reference
O.H.-2	SW NW Sec. 10 T27S R9W	2250'	Deep T-gradient	NA	Phillips	Lenzer et al, 1977
O.H.-1	SE NE Sec. 17 T27S R9W	2321'	Deep T-gradient	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 9-1	NE NW Sec. 9 T27S R9W	6885'	dry	227°C	Phillips	Lenzer et al, 1977 Glenn et al, 1981
Roosevelt KGRA 3-1	SW NE Sec. 3 T27S R9W	2724'	producer	NA	Phillips	Lenzer et al, 1977
Roosevelt KGRA 54-3	SW NE Sec. 3 T27S R9W	2882'	producer	NA	Phillips	Lenzer et al, 1977
Roosevelt KGRA 12-35	NW NW Sec. 35 T26S R9W	7324'	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 13-10	SW NW Sec. 10 T27S R9W	5351'	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Roosevelt KGRA 82-33	NE NE Sec. 33 T26S R9W	6028'	dry	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Utah State 14-2	SW NW Sec. 2 T27S R9W	6108'	producer	254°C	Thermal Power	Lenzer et al, 1977 Glenn and Hulen, 1979
Roosevelt HSU 25-15	NW SW Sec. 15 T27S R9W	7500'	producer	NA	Phillips	Lenzer et al, 1977 Geothermex, 1977
Utah State 72-16	SE NE Sec. 16 T27S R9W	1254'	producer	243°C	Thermal Power O'Brien	Lenzer et al, 1977 Glenn and Hulen, 1979
Utah State 24-36	SW NW Sec. 36 T27S R9W	6119'	dry	NA	Thermal Power	
Utah State 52-21	Sec. 21 T27S R9W	7500'	dry	206°C	Getty Oil Co.	Glenn and Hulen, 1979
GPC-15	Sec. 18 T27S R9W	1900'	Deep T-gradient	72°C	Geothermal Power Corp.	Glenn and Hulen, 1979

Table 2. Chemical analysis of selected thermal waters[1]

Well: Utah State		14-2	54-3	72-16	52-21	Hot Spring
Na	ppm	2150	2320	2000	1900	2500
K	ppm	390	461	400	218	488
Ca	ppm	9.2	8	12.20	114	22
Mg	ppm	0.6	<2	0.29	3.9	0
Fe	ppm		0.03		6.9	
Al	ppm		<0.5		<0.1	0.04
Si	ppm	229	263	244	67	146
Sr	ppm		1.2	1.20		
Ba	ppm		<0.5			
Mn	ppm		<0.2			
Cu	ppm		<0.1			
Pb	ppm		<0.2			
Zn	ppm		<0.1			
As	ppm	3.0	4.3			
Li	ppm		25.3	16.0		0.27
Be	ppm		0.005			
B	ppm	29	29.9	27.2	27.0	38
Ce	ppm		0.27			
F	ppm	5.2	6.8	5.3	3.4	7.5
Cl	ppm	3650	3860	3260	2885.1	4240
HCO_3	ppm		232	181	550.0	156
SO_4	ppm	78	72	32	86	73
NO_3	ppm				1.3	11
Total Dissolved Solids	ppm	>6614	7504	6444	5727	7800
pH (at collection T)		5.9		7.53	7.3	7.9
T (at surface)	°C	14				55
T (bottom hole)	°C	268	>260	243	204	284
Geothermometer						
T (Na-K-Ca)	°C	286	297	288	210[2]	284
T (quartz cond.)	°C	276	263	256	158	212
T (quartz adiab.)	°C	244	234	229	150	194
T (chalcedony)	°C	274	259	250	134	197
Total depth	m	1862	878	382	2289	

[1]From Capuano and Cole (in prep.). A blank indicates data not determined or information not available.

[2]Corrected for the Mg content of the fluid.

Table 3. Analyses of hot spring deposits

	Chalcedonic sinter, Opal Mound	Mn-cemented alluvium	Altered alluvium over fumarole
Na (%)	0.15	1.79	0.07
K (%)	0.14	3.12	0.26
Ca (%)	0.14	0.39	0.06
Mg (%)	0.01	0.10	0.03
Fe (%)	0.02	0.74	0.16
Al %)	0.09	5.18	4.01
Ti (ppm)	19	560	2040
P	---	651	336
Sr	33	386	266
Ba	---	4.9%	326
Cr	---	9	24
Mn	388	18.8%	173
Co	---	28	---
Ni	---	---	---
Cu	---	231	3
Mo	---	5	---
Pb	---	68	15
Zn	1	23	7
Cd	---	4	1
As	145	858	5
Sb	243	291	21
W	---	2940	29
Li	11	17	8
Be	99.8	18.6	3.4
Zr	---	17	42
La	---	37	34
Ce	---	42	56
Hg (ppb)	352	2210	49300

Table 4. Densities of lithologies from Mineral Mountains

Rock Type/Unit	g/cc ρ range	g/cc ρ(x)	No. of Samples	Reference
Tqm	2.43 - 2.80	2.64 ± .04	56	Glenn et al (1981)
"granitic"	2.45 - 2.71	2.59 ± .07	25	Carter and Cook (1978)
"granitic"	2.53 - 2.60	2.57 ± .02	11	Crebs and Cook (1976)
Td	2.54 - 2.90	2.76 ± .08	48	Glenn et al (1981)
Ts	2.43 - 2.63	2.55 ± .06	23	Glenn et al (1981)
Qrf/Qrd	2.16 - 2.24	2.22 ± .05	8	Crebs and Cook (1976)
Qrf/Qrd	2.22 - 2.38	2.31 ± .07	8	Carter and Cook (1978)
obsidian	2.15 - 2.35	2.30 ± .07	8	Carter and Cook (1978)
pЄbg (upper plate)	2.11 - 2.95	2.73 ± .28	136	Glenn et al (1981)
Gneiss and schist	2.63 - 2.74	2.69 ± .04	8	Crebs and Cook (1976)
Gneiss	2.63 - 2.74	2.69 ± .04	5	Carter and Cook (1978)
Limestone	2.55 - 2.97	2.71 ± .13	9	Carter and Cook (1978)
Quartzite[1]	2.50 - 2.74	2.62 ± .09	5	Carter and Cook (1978)
Alluvium		2.0 ± .1[2]		Carter and Cook (1978)
		2.05[3]		Glenn and Hulen (1979)

[1]Samples from Star Range, Rocky Range, Beaver Lake Mountains or Beaver Dam Mountains

[2]Nettleton's method

[3]Well log determination above water table

Table 5. Magnetic susceptibilities[1] of lithologies
Mineral Mountains, Utah

Rock Type	# Symbol	No. Sites	No. Observations	Mean	* Std. Dev.	Range
				← Units of micro cgs →		
Alluvium	Qal	2	8	500	86	388-640
Rhyolite flows	Qrf	3	56	179	69	41-347
Rhyolite domes	Qrd	4	43	131	84	0-372
Rhyolite dikes	Trd	1	10	406	29	363-463
Diabase dikes	Tds	1	3	474	-	466-483
Granite dikes	Tgr-d	3	14	411	129	166-650
Granite	Tg	6;tr.	63	1353	333	623-2053
Seyenite	Ts	5;tr.	36	1998	272	1546-2056
Porphyritic Granite	Tpg	6;tr.	86	1626	282	10-2467
Quartz Monzonite	Tqm	7;tr.	58	1918	166	1433-2244
Hornblende Gneiss	Tgn	4	24	2310	830	1520-3440
Biotite Gneiss	Tn	2	6	2644	405	2118-3170
Sillmanite Schist	Pϵs	tr.	10	22	32	0-89
Banded Gneiss	Pϵbg	6;tr.	35	1252	1698	38-5421

[1]Corrected in situ susceptibilities, 6" diam coil, Bison system

#Map symbol after Nielson et al, 1978

*Standard deviation not alway statistically significant, but indicative of susceptibility variability.

tr. indicates observations along traverses, 500-2500 feet long

Table 6. Magnetic anomaly and source charcteristics
Roosevelt Hot Springs Area, Utah

Anomaly	Amplitude (gammas)	Correlation[1] ρtopo/mag	Geologic Setting[2]	Probable Source[3]	SUSC. Contrast micro cgs
1 b,c	120;100	X	Qal,Qcal;Tqm	Tqm	2000
2	280	H	Tqm,Tgr	Tqm,Ts;r.t.c.	~ 2000
3	120+	M	Tqm	Tqm r.t.c.	
4	- 80	R	Qrd,Qal	Qrd,Qal;r.t.c.	~-1400
5	-100	M-X	Qal,Tgr,Ts,Tqm	Qal,Tgr;i.t.c.	
6	190	H	Ts,Tqm	Ts,Tqm;r.t.c.	~ 2000
7 a,b	300;240	X	Qal,Qcal		2000-5000
8	- 40	X	Qal	Qal; i.t.c.	
9	50	H	Tgn,Ts,Tpg	Tgn,Ts r.t.c.	
10	-200	L-R	Qrt,Qal	Qrt(reversed)	~-1500
11	50	H	Tqm	Tqm; r.t.c.	
12	190;200	L-X	Ts,Qal	Ts;	~ 2000
13*	250	H	Tqm	Tqm; * r.t.c.	
14*	-180	R?	Tqm	Tqm; * i.t.c.	
15*	- 80	R	Qrd	Qrd; * r.t.c.	
16	200	H-M	Tpg,Tn,Tgn,Ts,Tg	Ts,Tpg,Tgn;r.t.c.	~ 2000
17	-150	X-R	Qrd,Tg	Qrd; i.t.c.	
18	- 90	L-X	Qal,Pεbg,Tpg	Qal;Pεbg,Tg	
19	350	H	Tg,Tn,Tpg,Qra,Qrd	Tn,Tg,Tpg r.t.c.	~ 2500
20	250	H	Tn,Tpg,Tgn	Tn,Tpg,Tgn r.t.c.	~ 2000
21	250	X	Qal,Tn,Tgn,Tpg	Tpg,Tgn,Tn	~ 2000
22	150	L	Qal,Tbg,Tgr	Tbg,Tgr	
23	- 30	L-R	Qra,Qrd	Qra; i.t.c.	
24	20	H	Qrd	Qrd; r.t.c.	
25	- 50	H-R	Tq	--- r.t.c.	
26	350	H	Tdb,Tg,Tbg	Tdb, r.t.c.	~ 3500

[1]H = high, M = moderate, L = low, X = insignificant, R = reversed.
[2]Map symbol after Nielson et al (1978).
[3]r.t.c. = reduced terrain clearance; i.t.c. = increased terrain clearance; * indicates poor contour representation

Figure Captions

Fig. 1 - Index map showing the location of the Roosevelt Hot Springs KGRA, Utah.

Fig. 2 - Geologic map of the Roosevelt Hot Springs KGRA and vicinity (after Sibbett and Nielson, 1980a).

Fig. 3 - Map of Milford and Beaver Valleys showing the elevation of the non-thermal aquifer (after Mower and Cordova, 1974; and Mower, 1978).

Fig. 4 - Concentrations of chlorine (Fig 4A) and boron (Fig. 4B) in ground water in the Roosevelt Hot Springs KGRA and adjacent Milford Valley. Data is from Cordova (1974) and Mower (1978) and was compiled by D. R. Cole.

Fig. 5 - Distribution of mercury and temperatures in wells 14-2, 72-16, and 52-21.

Fig. 6 - Terrain corrected bouguer gravity map for Milford Valley and Mineral Mountains area (modified from Cook et al, 1980).

Fig. 7 - Total magnetic intensity map of the Roosevelt Hot Springs - Mineral Mountains area. Sub-alluvial magnetic sources and principal structural zones are indicated. Major anomalies are numbered for reference to the text.

Fig. 8 - Seismic emission sources, Roosevelt Hot Springs KGRA and structures interpreted from Vibroseis* profile GSI-5 (see Fig. 9).

Fig. 9 - Structural interpretation of Vibroseis* profile GSI-5. Units as determined from velocity analysis and reflection continuity.

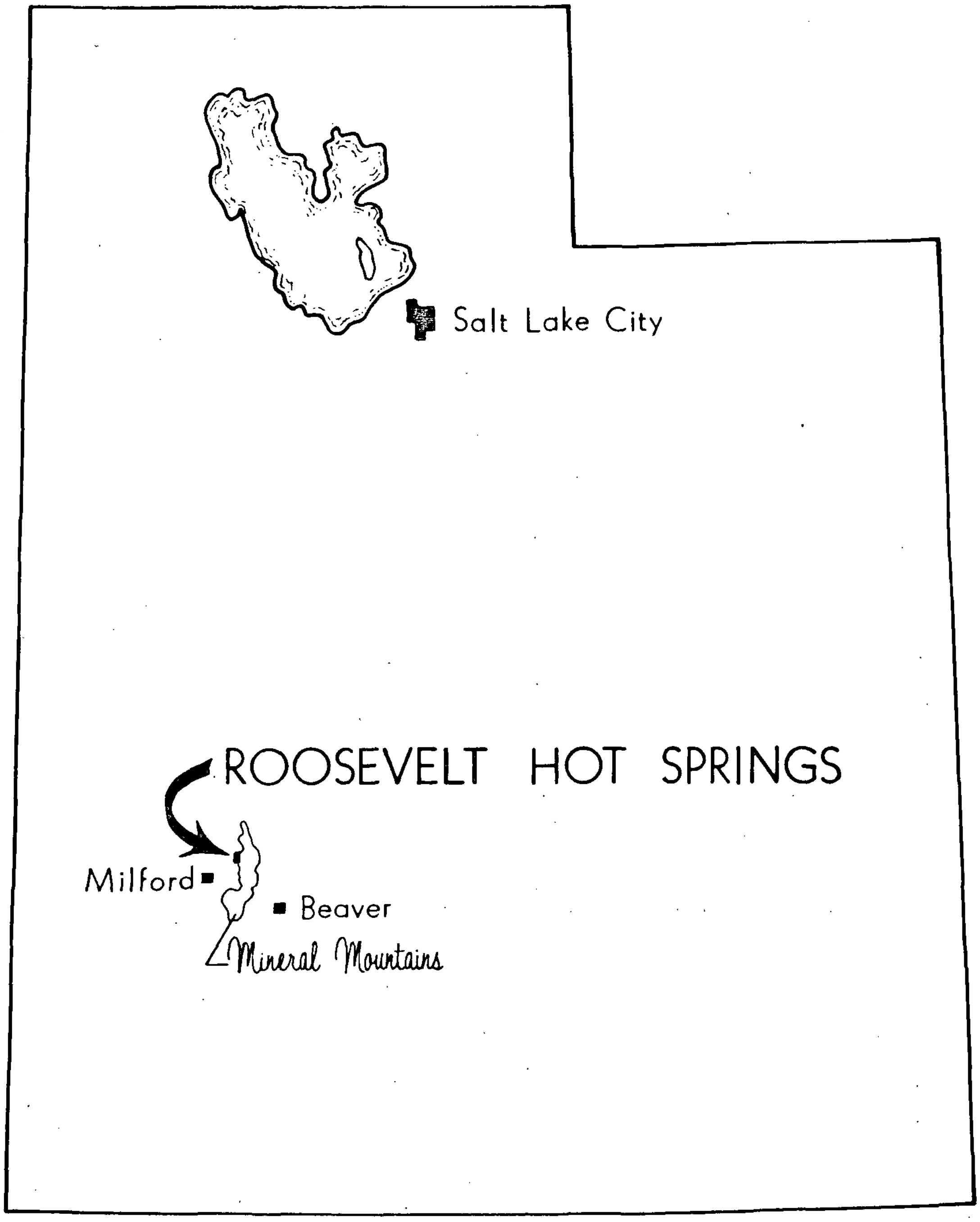
Fig. 10 - Dipole-dipole resistivity array geometry and plotting convention. Observed data, numerical models and computed resistivity values are shown for Lines GOC 4 and GOC 5.

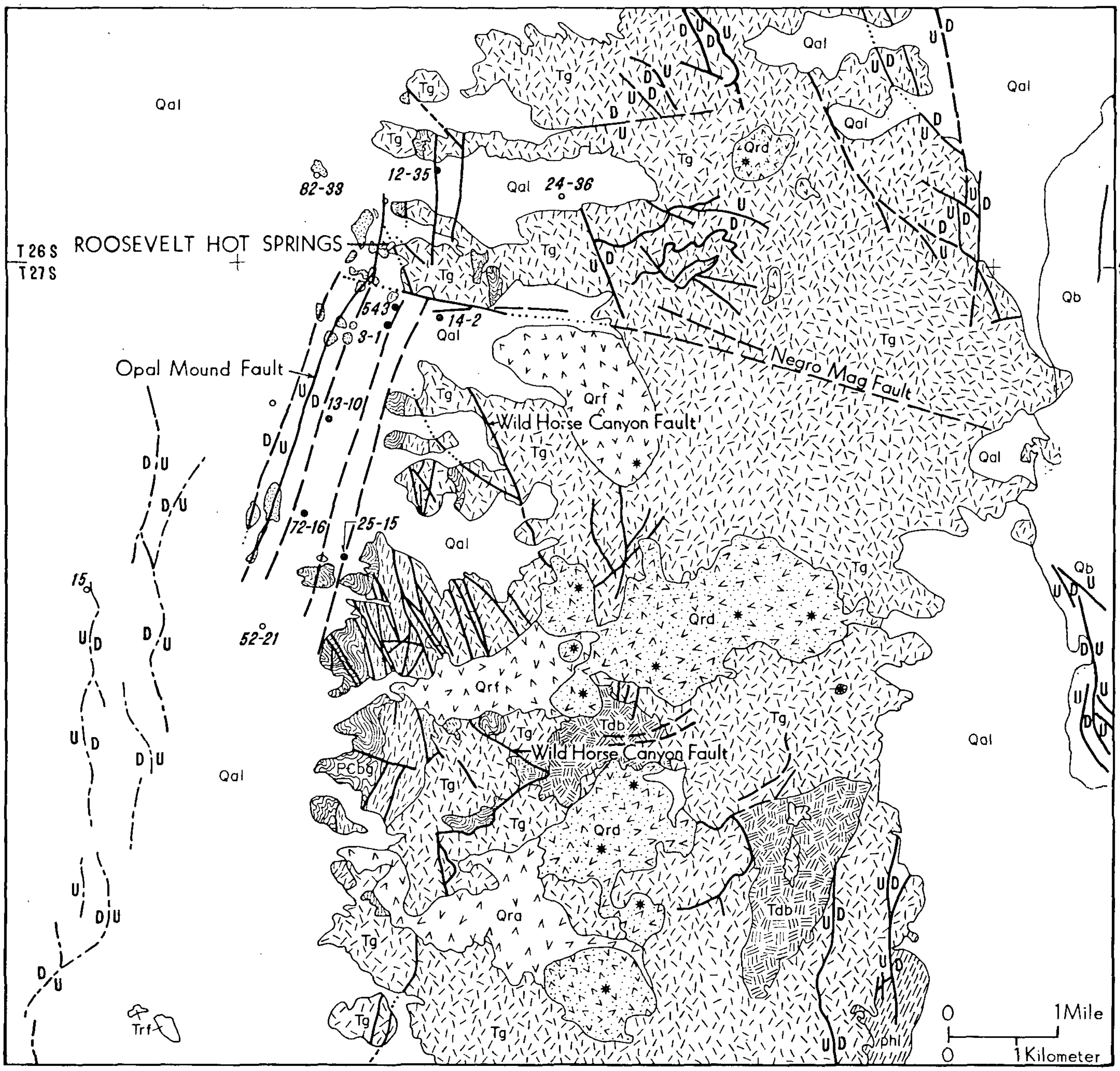
Fig. 11 - Interpreted electrical resistivity distribution for the depth interval 100-150 meters, Roosevelt Hot Springs KGRA.

Fig. 12 - Interpreted electrical resistivity sections, Roosevelt Hot Springs KGRA.

Fig. 13 - Magnetotelluric profile C-C': a) Observed apparent resistivity, TM mode; b) Modeled apparent resistivity, TM mode; c) Finite element model for profile C-C' giving best fit between computed (c) and observed TM mode resistivities. Intrinsic resistivity values in ohm-m; OMF = Opal Mound fault. (After Wannamaker et al, 1980).

Fig. 14 - Thermal studies map, Roosevelt Hot Springs KGRA.


Salt Lake City
ROOSEVELT HOT SPRINGS
Milford
Beaver
Mineral Mountains



Qal - alluvium

Qb - basalt

Qrd - rhyolite domes

Qra - pyroclastic deposits

Qrf - rhyolite flows

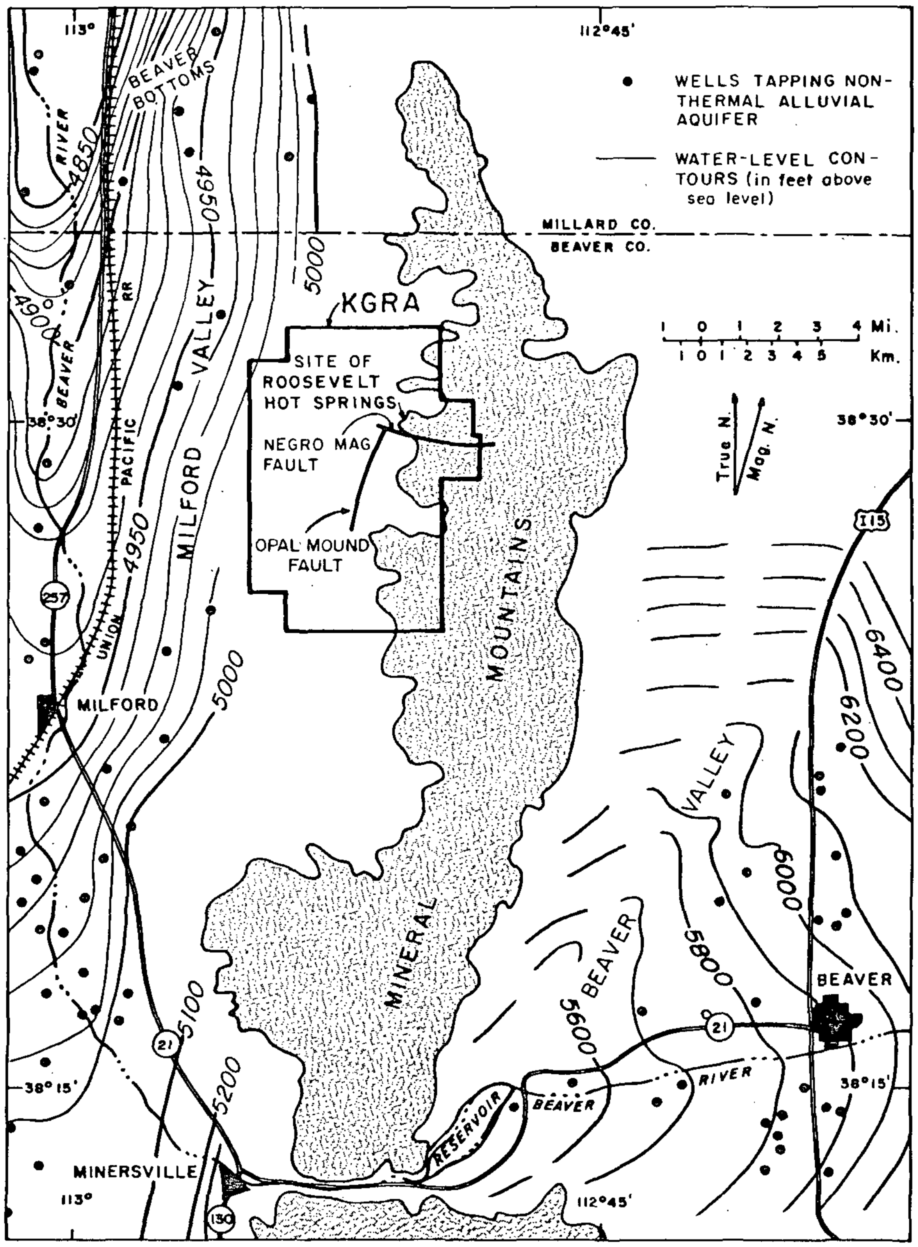
Trf - rhyolite flows

Tg - granite, quartz monzonite, and syenite

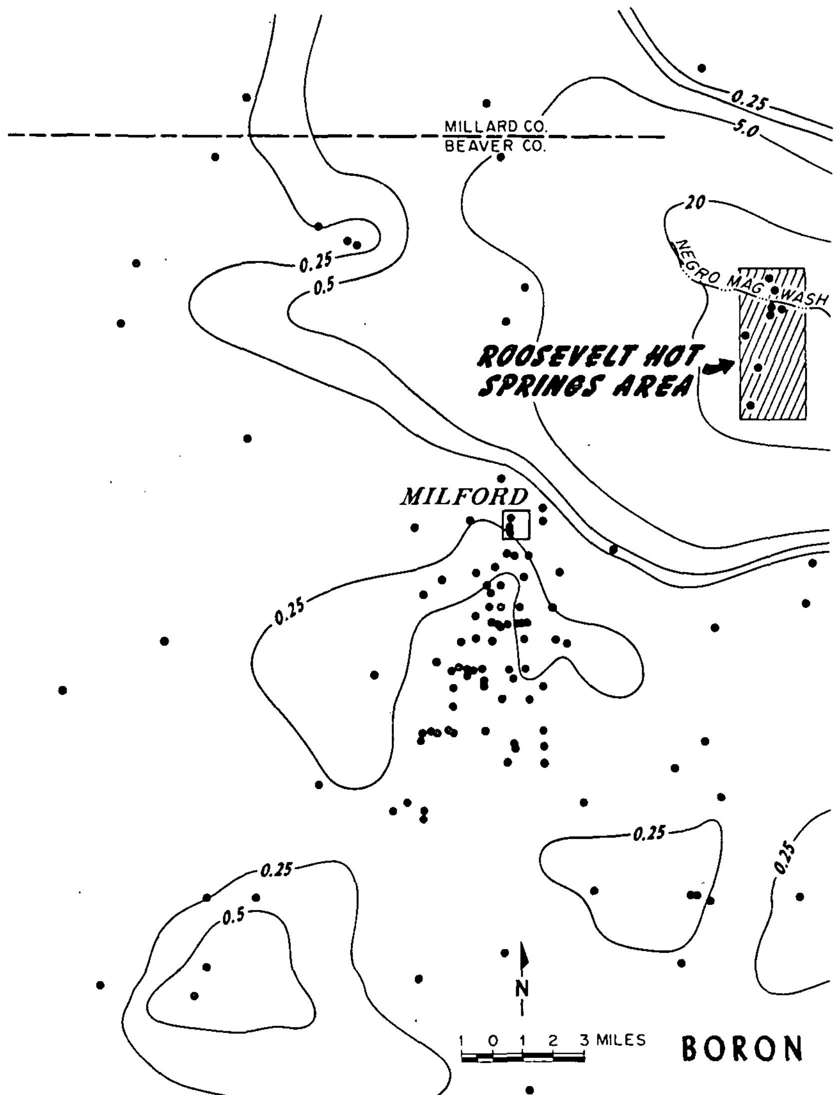
Tdb - diorite

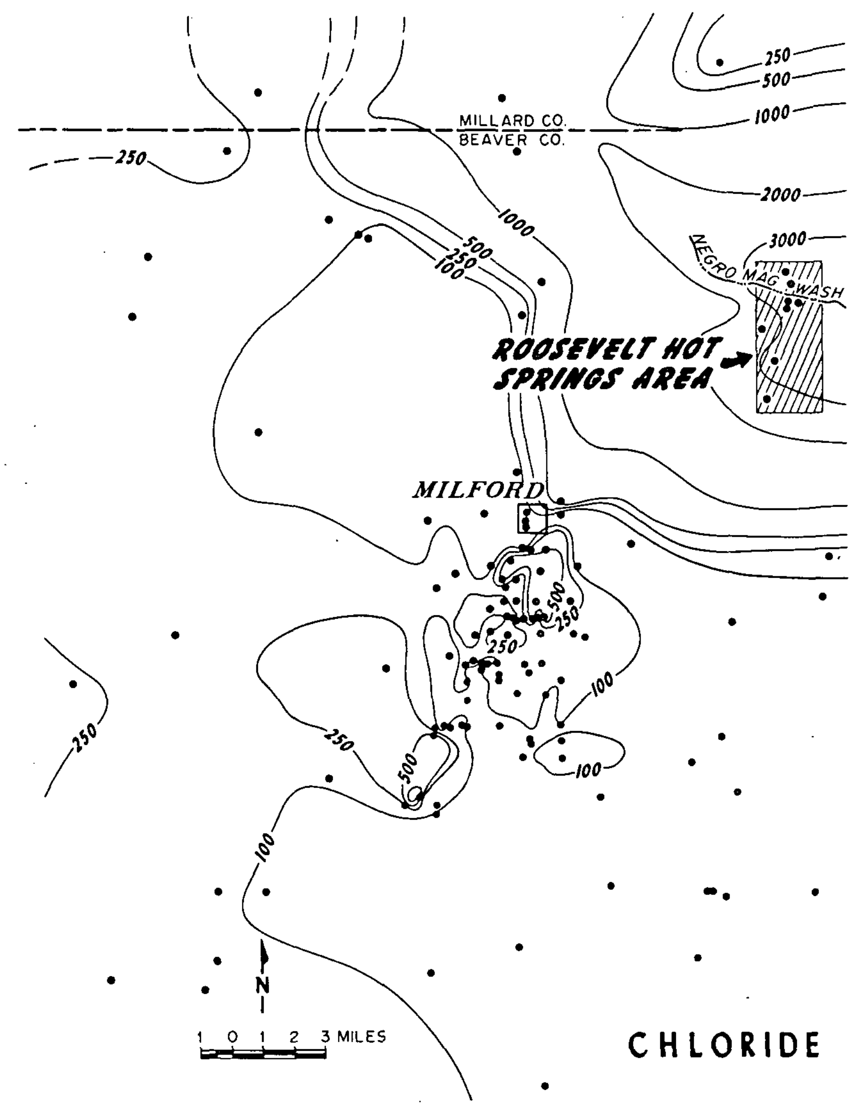
phl - metasediments

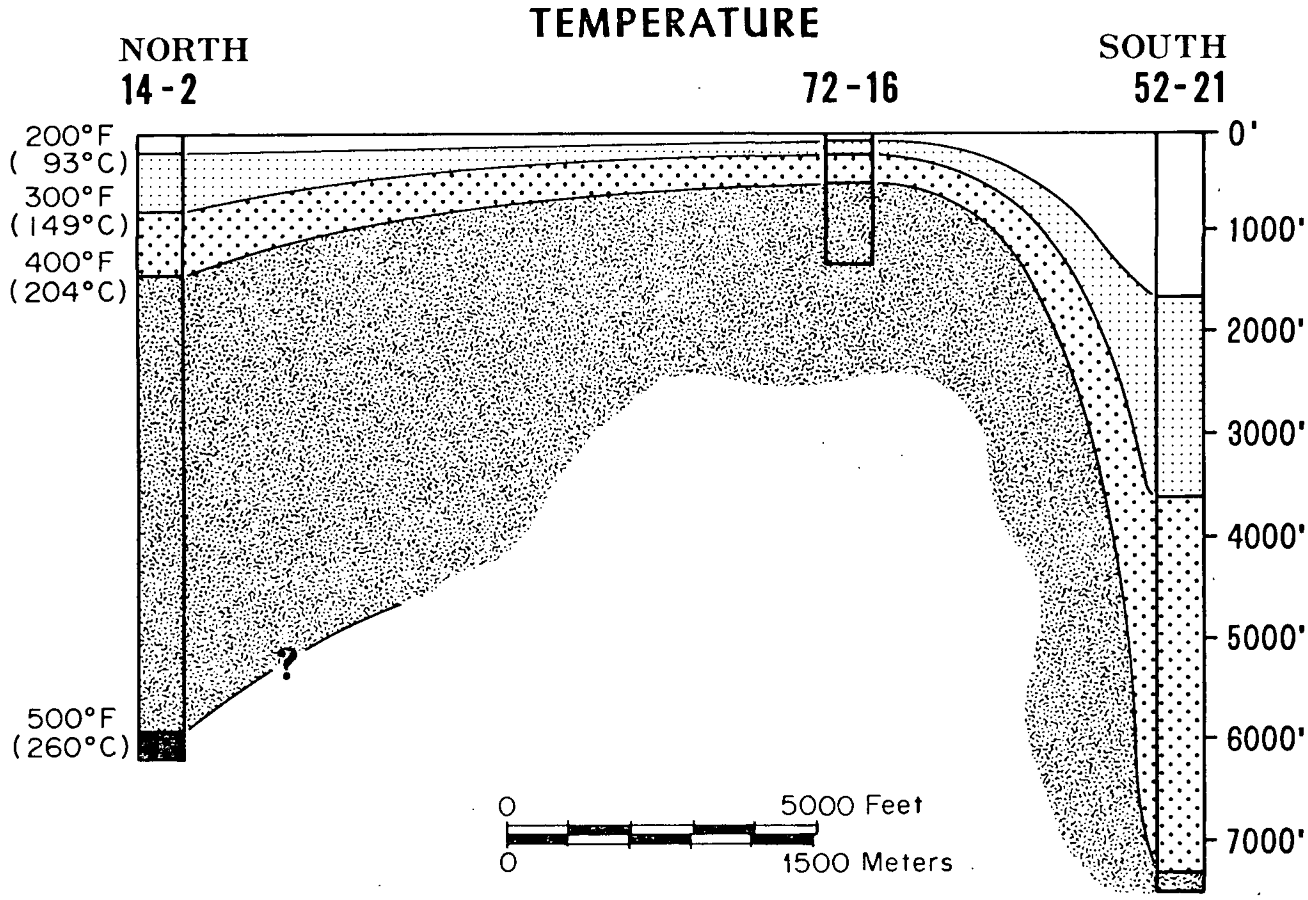
Pεbg - bonded gneiss

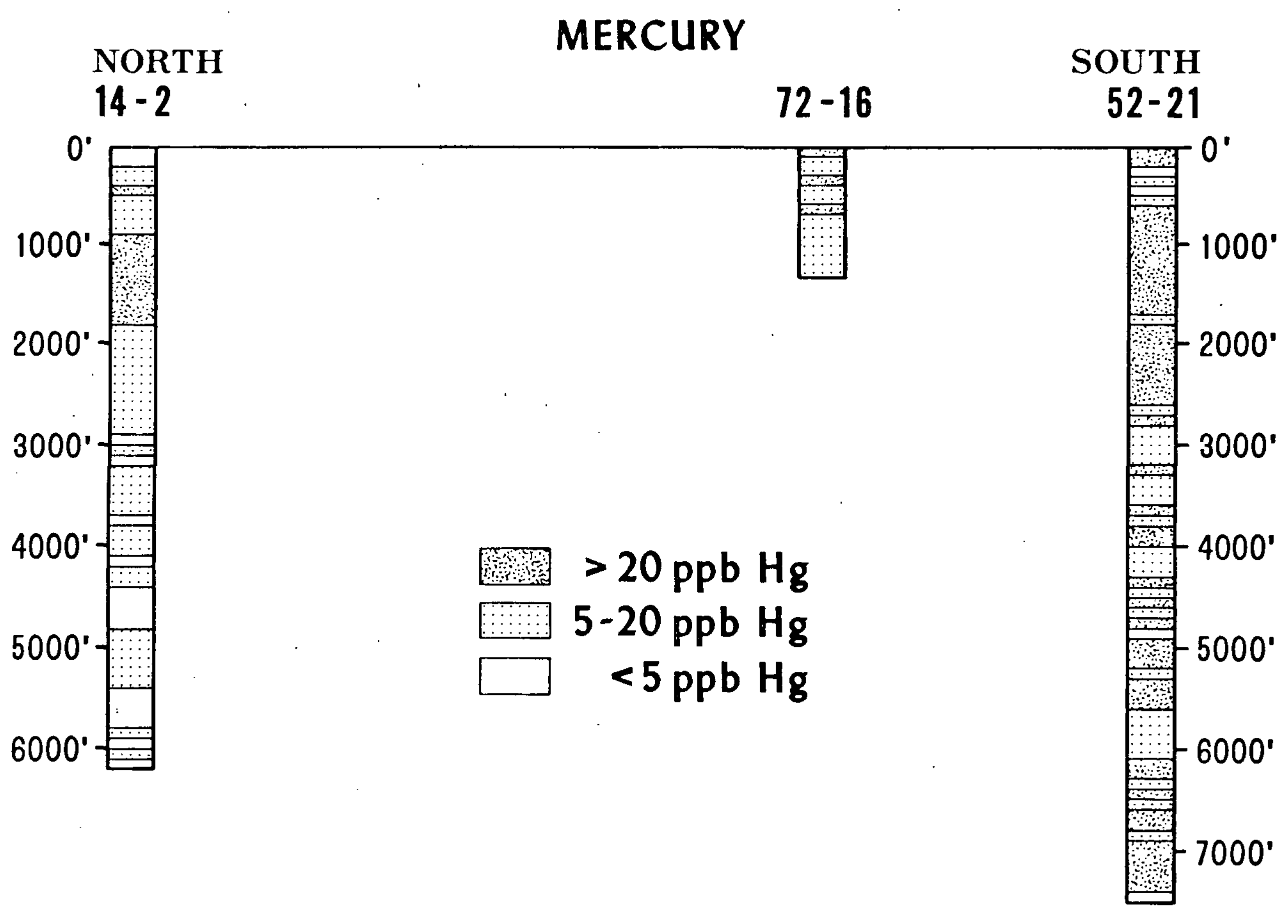


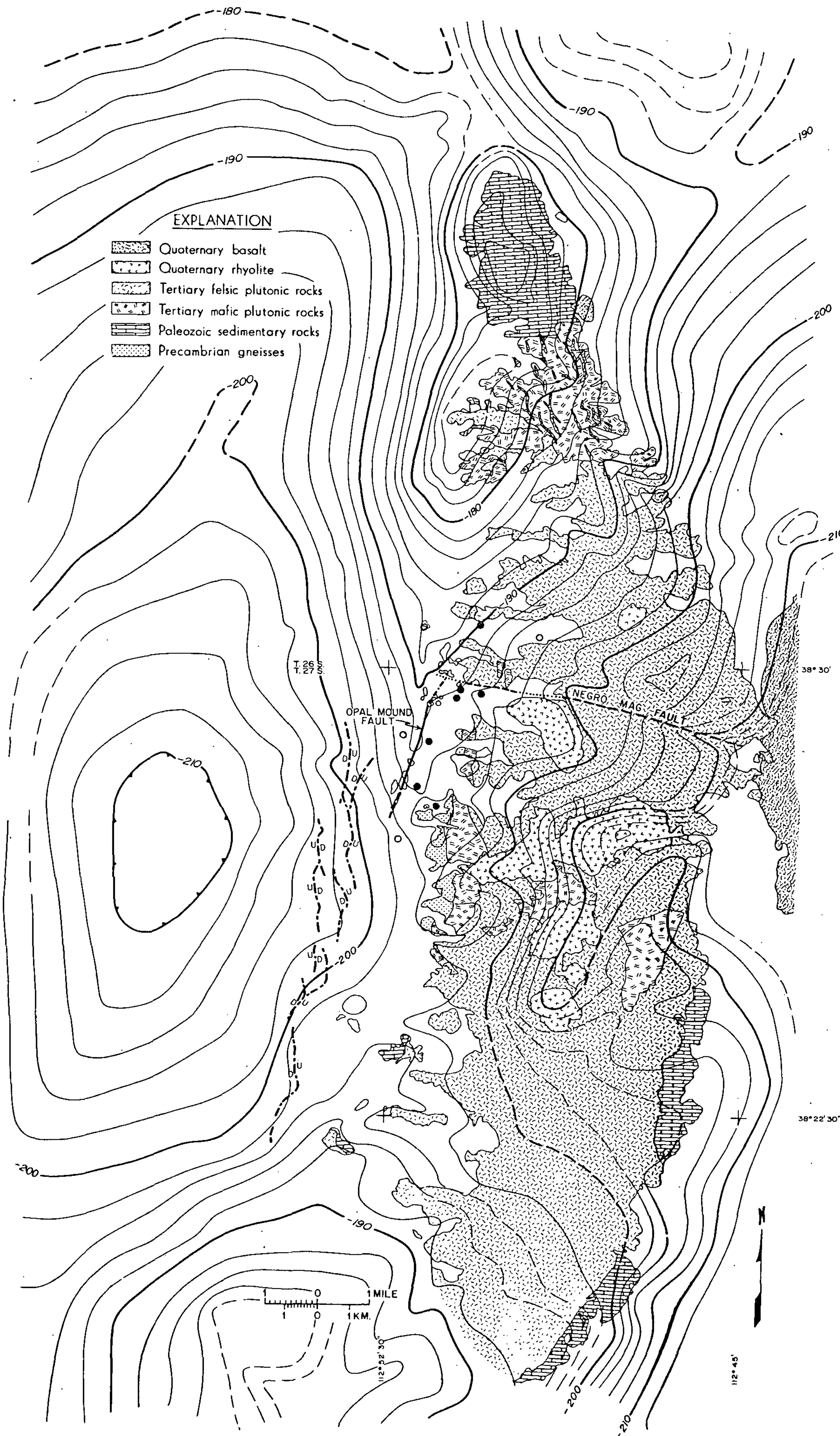
WELLS TAPPING NON-THERMAL ALLUVIAL AQUIFER
WATER-LEVEL CONTOURS (in feet above sea level)
113°
112°45'
38°30'
38°15'
MILLARD CO.
BEAVER CO.
KGRA
SITE OF ROOSEVELT HOT SPRINGS
NEGRO MAG FAULT
OPAL MOUND FAULT
BEAVER BOTTOMS
MILFORD VALLEY
UNION PACIFIC RR
BEAVER RIVER
MILFORD
MINERSVILLE
MINERAL MOUNTAINS
BEAVER VALLEY
BEAVER
RESERVOIR
4850
4900
4950
5000
5100
5200
5600
5800
6000
6200
6400
257
21
130
I15
Mi.
Km.
True N.
Mag. N.

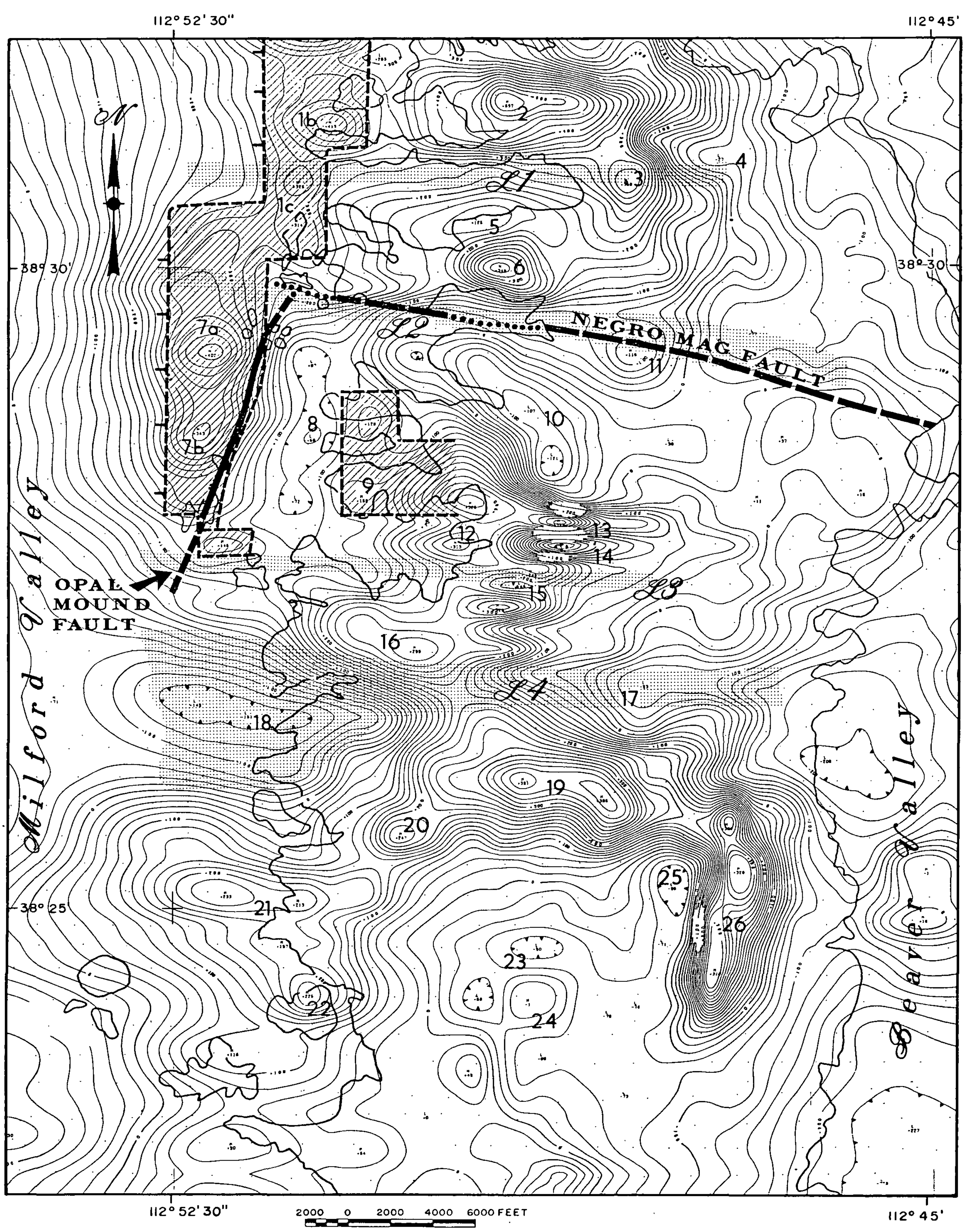

MILLARD CO.
BEAVER CO.
NEGRO MAG WASH
ROOSEVELT HOT SPRINGS AREA
MILFORD
0.25
0.5
5.0
20
1 0 1 2 3 MILES
N
BORON


MILLARD CO.
BEAVER CO.
NEGRO MAG WASH
ROOSEVELT HOT SPRINGS AREA
MILFORD
100
250
500
1000
2000
3000
1 0 1 2 3 MILES
N
CHLORIDE

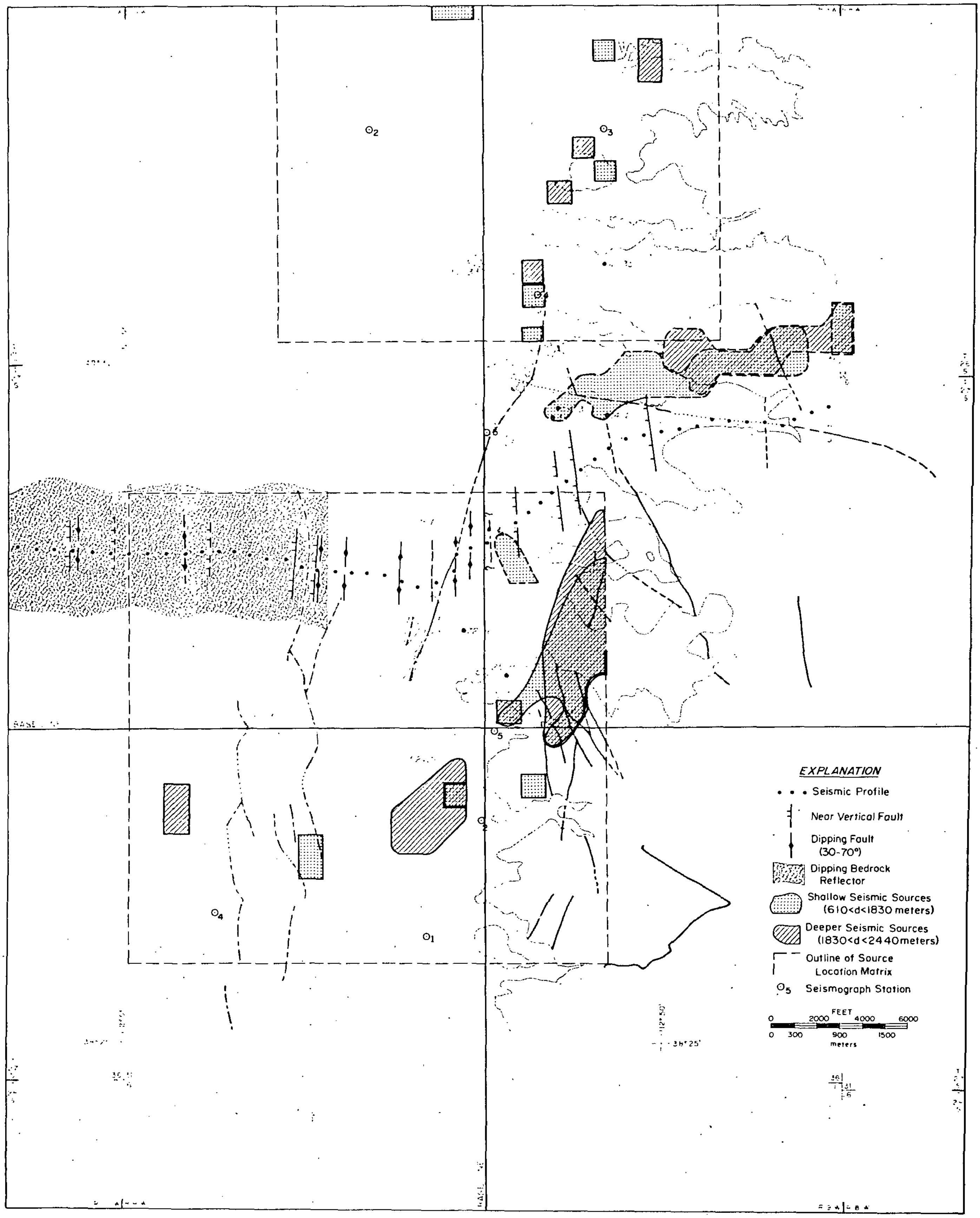

TEMPERATURE
NORTH
14-2
72-16
SOUTH
52-21
200°F
(93°C)
300°F
(149°C)
400°F
(204°C)
500°F
(260°C)
?
0
5000 Feet
0
1500 Meters
0'
1000'
2000'
3000'
4000'
5000'
6000'
7000'


MERCURY
NORTH
14-2
72-16
SOUTH
52-21
0'
1000'
2000'
3000'
4000'
5000'
6000'
7000'
> 20 ppb Hg
5-20 ppb Hg
< 5 ppb Hg

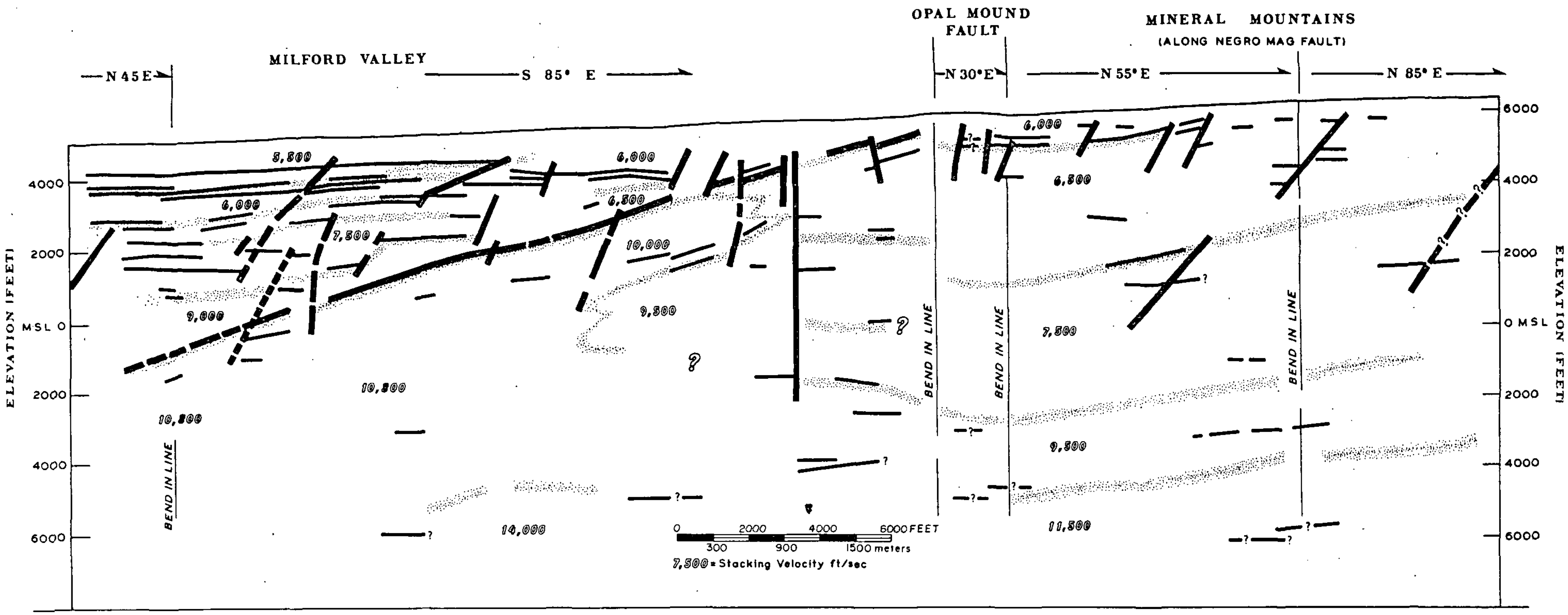

EXPLANATION
Quaternary basalt
Quaternary rhyolite
Tertiary felsic plutonic rocks
Tertiary mafic plutonic rocks
Paleozoic sedimentary rocks
Precambrian gneisses
180
190
200
210
T. 26 S.
T. 27 S.
OPAL MOUND FAULT
NEGRO MAG FAULT
38° 30'
38° 22' 30"
112° 52' 30"
112° 45'
N
1 0 1 MILE
1 0 1 KM.
U
D



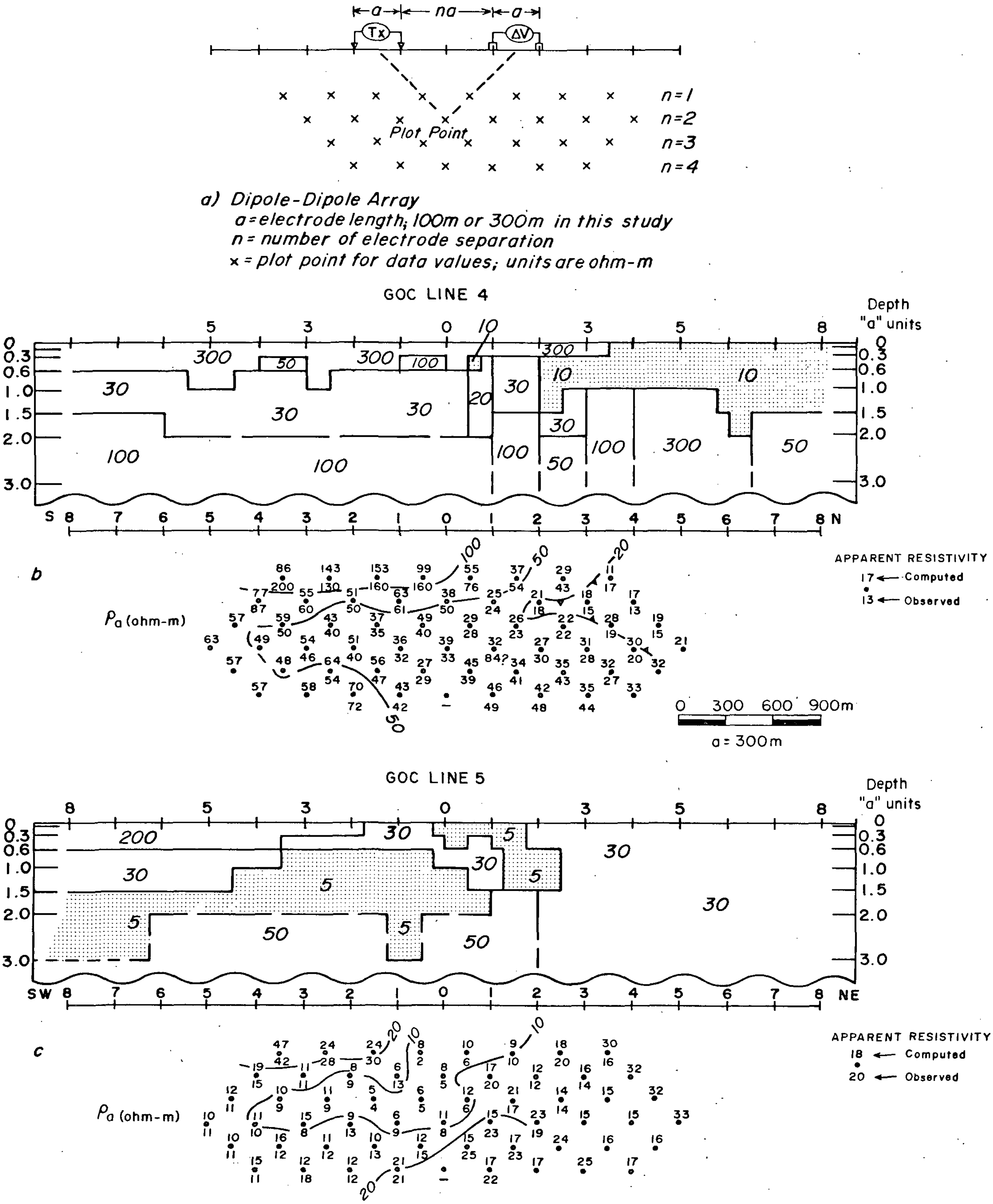
112° 52' 30"
112° 45'
38° 30'
38° 25'
N
NEGRO MAG FAULT
OPAL MOUND FAULT
Milford Valley
Beaver Valley
L1
L2
L3
L4
1b
1c
2
3
4
5
6
7a
7b
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
2000 0 2000 4000 6000 FEET



EXPLANATION
Seismic Profile
Near Vertical Fault
Dipping Fault (30-70°)
Dipping Bedrock Reflector
Shallow Seismic Sources (610<d<1830 meters)
Deeper Seismic Sources (1830<d<2440meters)
Outline of Source Location Matrix
Seismograph Station
FEET
0 2000 4000 6000
0 300 900 1500
meters



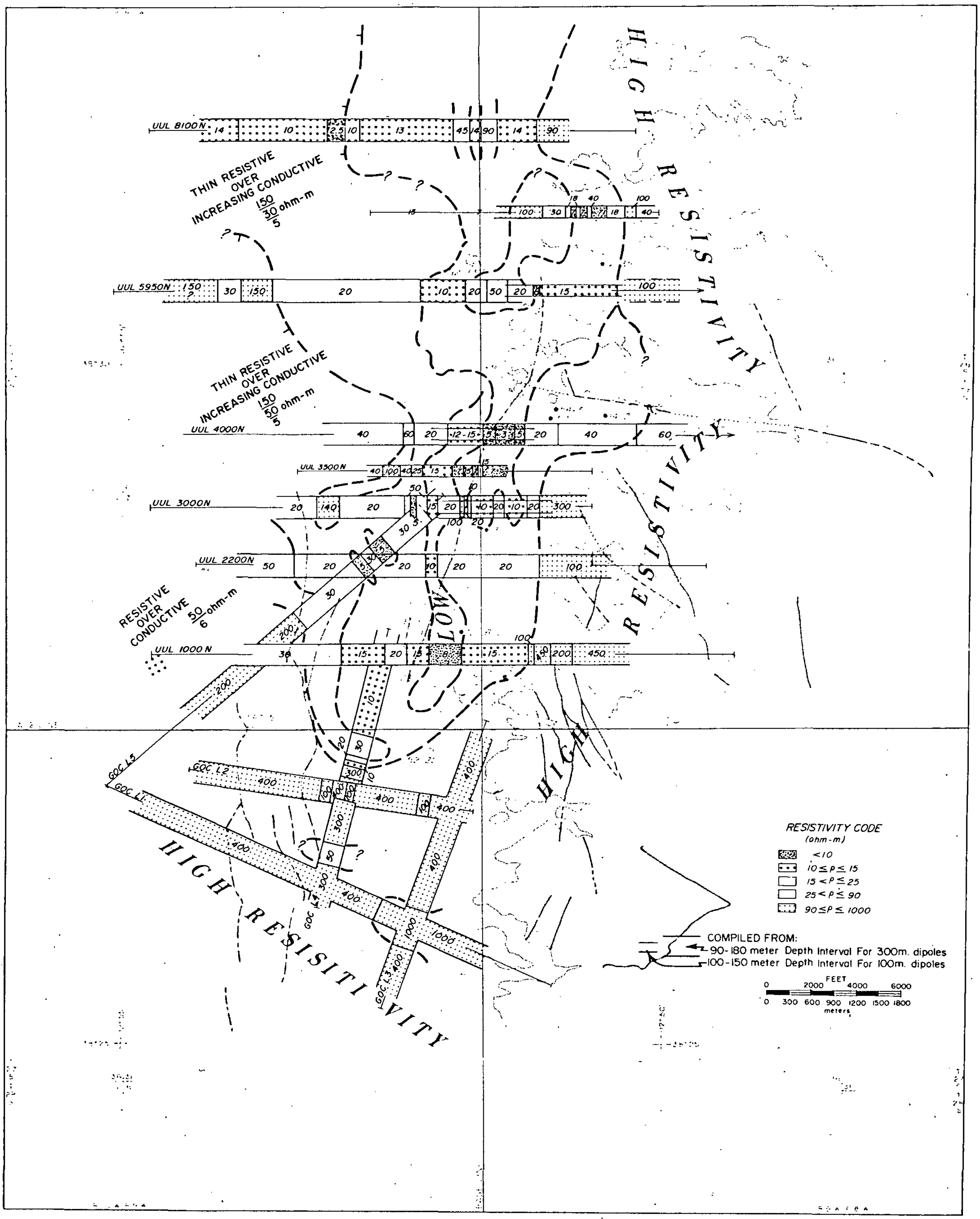
OPAL MOUND FAULT
MINERAL MOUNTAINS
(ALONG NEGRO MAG FAULT)
N 45 E
MILFORD VALLEY
S 85° E
N 30° E
N 55° E
N 85° E
ELEVATION (FEET)
4000
2000
MSL 0
2000
4000
6000
6000
4000
2000
0 MSL
2000
4000
6000
5,500
6,000
6,000
6,500
7,500
10,000
9,000
9,500
10,500
10,500
14,000
6,000
6,500
7,500
9,500
11,500
BEND IN LINE
BEND IN LINE
BEND IN LINE
BEND IN LINE
0 2000 4000 6000 FEET
300 900 1500 meters
7,500 = Stacking Velocity ft/sec



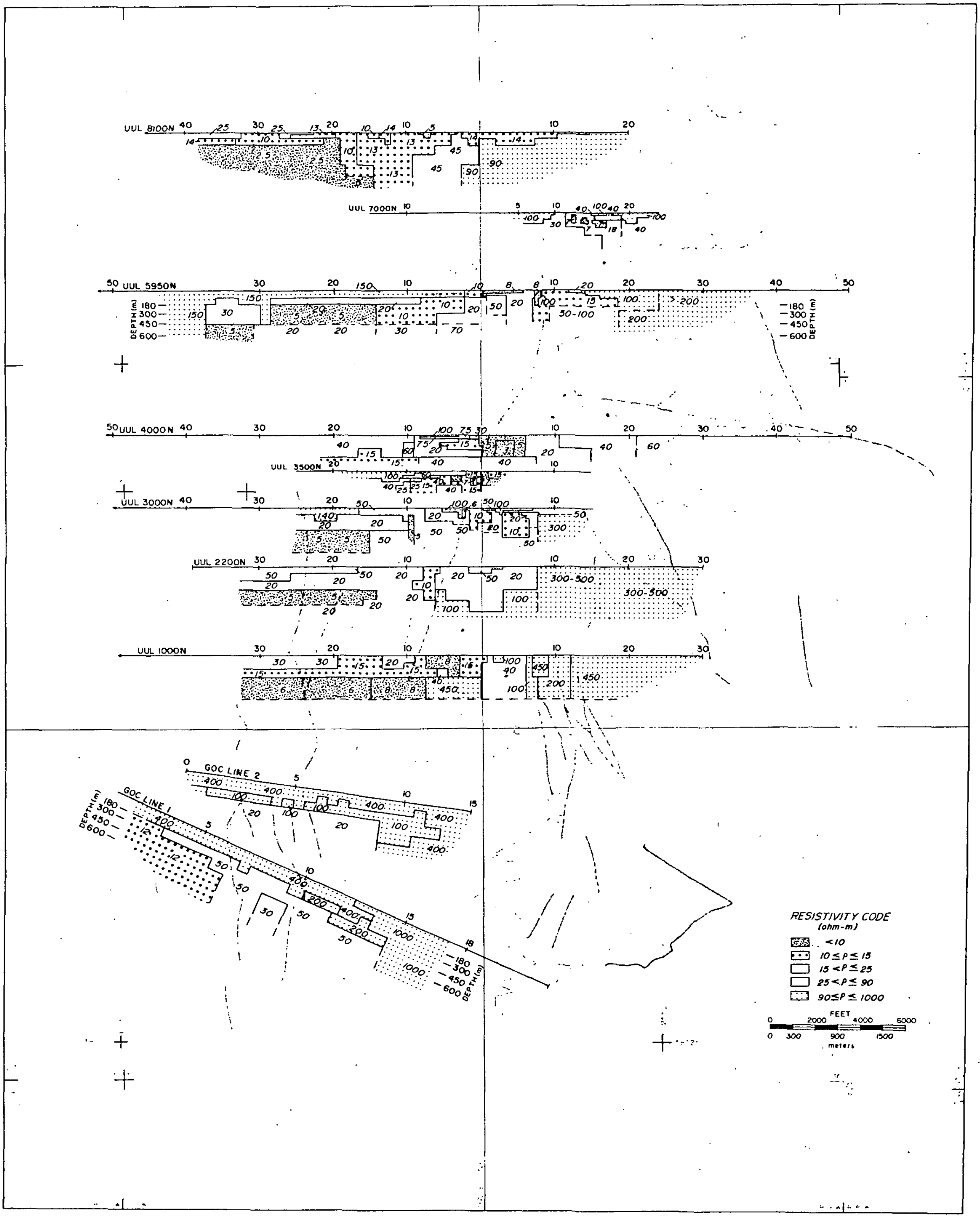
a) Dipole-dipole array geometry and plotting convention

b) GOC line 4: Numerical model

c) GOC line 5: Numerical model



HIGH RESISTIVITY
UUL 8100N
THIN RESISTIVE OVER INCREASING CONDUCTIVE 150/30/5 ohm-m
UUL 5950N
THIN RESISTIVE OVER INCREASING CONDUCTIVE 150/50/5 ohm-m
UUL 4000N
UUL 3500N
UUL 3000N
UUL 2200N
RESISTIVE OVER CONDUCTIVE 50/6 ohm-m
UUL 1000 N
LOW
HIGH
GOC L5
GOC L2
GOC L1
GOC L4
GOC L3
HIGH RESISITIVITY
RESISTIVITY CODE (ohm-m)
<10
10 ≤ ρ ≤ 15
15 < ρ ≤ 25
25 < ρ ≤ 90
90 ≤ ρ ≤ 1000
COMPILED FROM:
90-180 meter Depth Interval For 300m. dipoles
100-150 meter Depth Interval For 100m. dipoles
FEET
0 2000 4000 6000
0 300 600 900 1200 1500 1800
meters


UUL 8100N
UUL 7000N
UUL 5950N
UUL 4000N
UUL 3500N
UUL 3000N
UUL 2200N
UUL 1000N
GOC LINE 2
GOC LINE 1
DEPTH (m)
180
300
450
600
RESISTIVITY CODE
(ohm-m)
<10
10 ≤ ρ ≤ 15
15 < ρ ≤ 25
25 < ρ ≤ 90
90 ≤ ρ ≤ 1000
FEET
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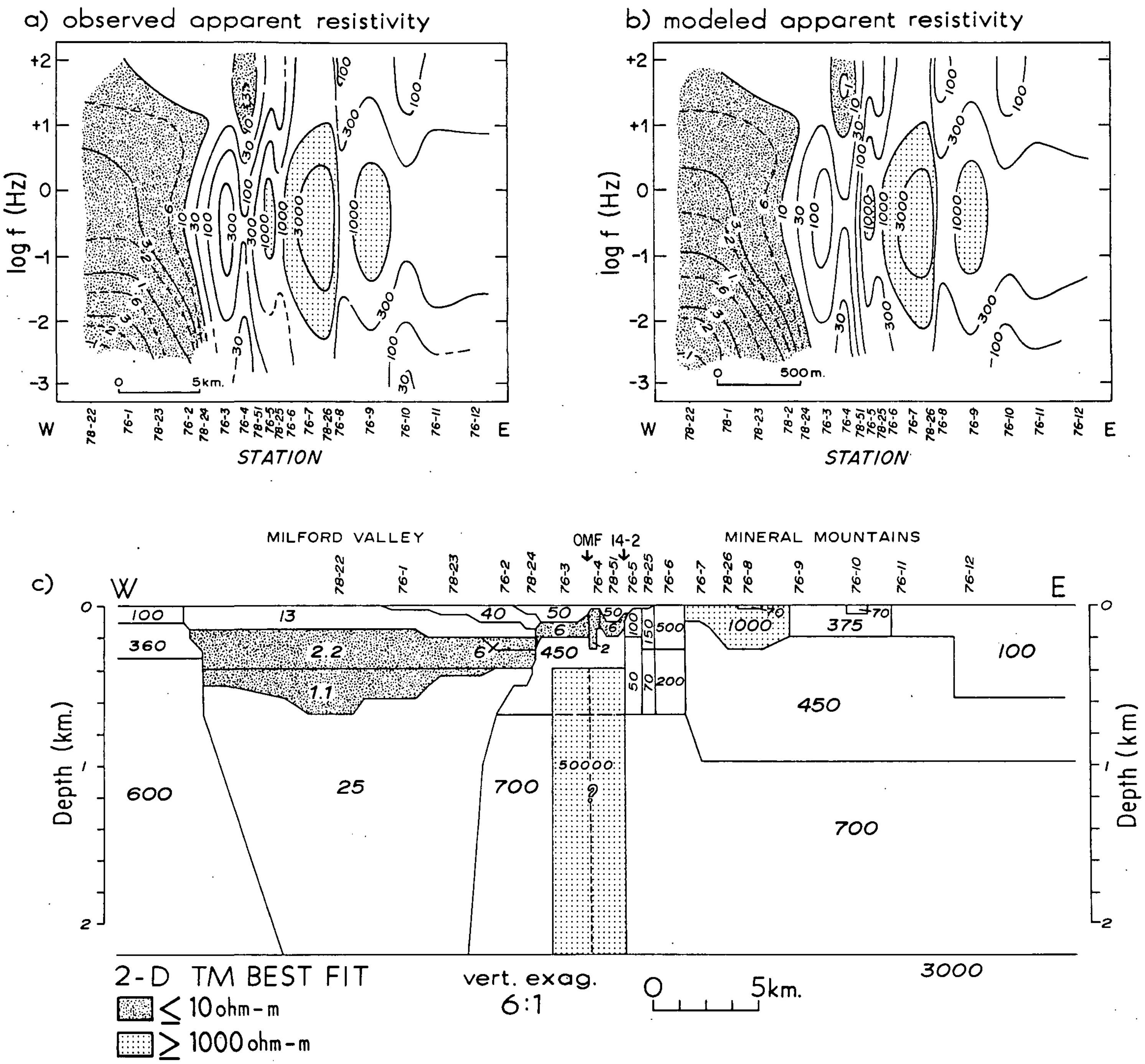
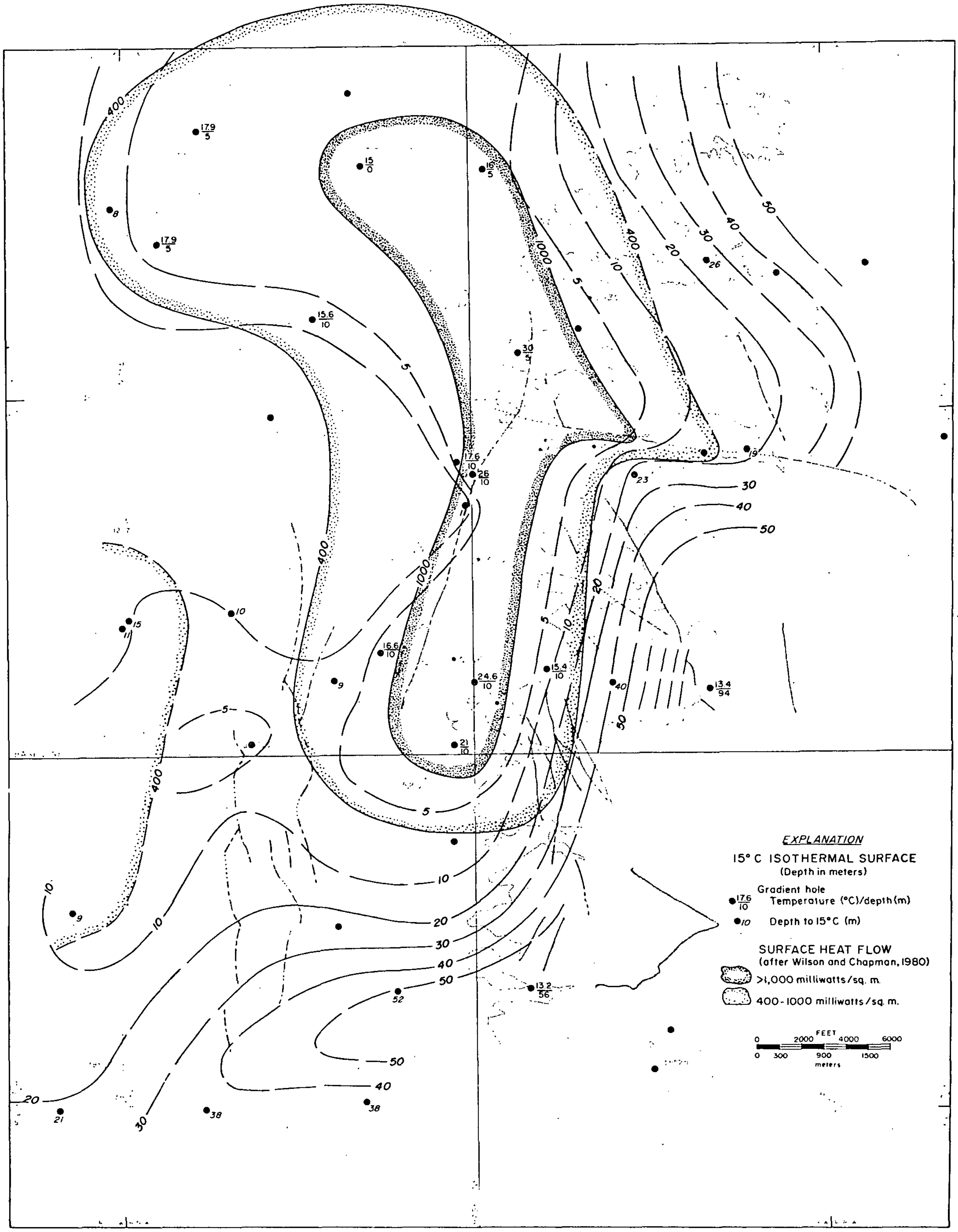
0 300 900 1500
meters



a) observed apparent resistivity
b) modeled apparent resistivity
log f (Hz)
+2
+1
0
-1
-2
-3
0 5km.
0 500m.
W
E
78-22
76-1
78-23
76-2
78-24
76-3
76-4
78-51
76-5
78-25
76-6
76-7
78-26
76-8
76-9
76-10
76-11
76-12
STATION
MILFORD VALLEY
OMF 14-2
MINERAL MOUNTAINS
c)
Depth (km.)
Depth (km)
2-D TM BEST FIT
≤ 10 ohm-m
≥ 1000 ohm-m
vert. exag.
6:1
0 5km.


EXPLANATION
15° C ISOTHERMAL SURFACE
(Depth in meters)
Gradient hole
Temperature (°C)/depth(m)
Depth to 15°C (m)
SURFACE HEAT FLOW
(after Wilson and Chapman, 1980)
>1,000 milliwatts/sq. m.
400-1000 milliwatts/sq. m.
FEET
0 2000 4000 6000
0 300 900 1500
meters

in 17th U.S. Symp. on Rock Mechanics (only in Abstract form)

GEOTHERMAL EXPLORATION OF ROOSEVELT KGRA, UTAH

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ABSTRACT The Roosevelt geothermal field is situated at the boundary between the Mineral Range and Milford graben in eastern Beaver County, Utah. Valley fill sediments in the graben are approximately 1,500 meters thick in the center of the valley. Bedrock is stepped up along several normal faults to the west flank of the Range, where the westernmost exposures consist of Precambrian (?) gneissic rocks. The metamorphic rocks are invaded in a zone of injection by Late Cenozoic granite and related silicic differentiates. Paleozoic sedimentary rocks, exposed on the west side of the valley, terminate somewhere in the graben having been removed by erosion. The westernmost exposures of Precambiran crystallines are in an apparent horst block which is bounded on the east by the Dome Fault. Rhyolitic flows from seven or more eruptive centers cap much of the granite east of the KGRA. Magma additions to the chamber feeding these eruptive centers are thought to be supplying the heat beneath the Roosevelt prospect. Recent faulting in the vicinity of the prospect is indicated by fresh scarps in alluvium and the cutting and displacement of hot spring deposits. Faults appear to be major controlling structures in the subsurface hydrologic regime.

The investigations include hydrochemistry, chemical geothermometry, petrography, gravity, magnetics, resistivity (dipole mapping and MT), groundnoise and thermal gradient drilling. A model assimilating the data was formulated. Additional resistivity, thermal gradient drilling and deep drilling has resulted in modifications of the earlier model. The thermal anomaly is underlain by intermediate and silicic crystalline rocks, at the surface or at shallow depths. The reservoir is a fracture system. The depth to its top is less than 900 meters over a significant portion of the anomaly. Locally, the fracture zones have extraordinarily high effective permeability yielding up to 113,000 kg/hr flashed steam from a reservoir with temperatures in excess of 250°C, with pressures near hydrostatic, and with fluids of less than 8,000 ppm total dissolved solids. Terminating quartz crystals and drilling breaks indicate that open fractures are still maintained in granite at depths of 1625 m and temperatures above 240°C. Open fractures are fairly common at significantly shallower depths and lower temperatures and locally constitute difficult drilling problems above the reservoir.

INTRODUCTION Geothermal energy is the natural heat of the earth extractable through the media of subsurface fluids. Magma in the earth's crust is usually too deep for its energy to be directly exploited. In some areas, however, magma works itself close to the surface and heat is slowly transmitted to the adjacent rock. Any water present is also heated and may leak to the surface producing hot springs, geysers, or fumeroles. Wells drilled into potentially productive zones bring large volumes of the geothermal fluids to the surface.

Our main interest in developing the energy is to provide steam to generate electrical power. The reservoir fluids may be in the form of steam or pressurized hot water. Increased pressure raises the boiling point of water. For example, at 1,000 m the boiling point of water is approximately 310°C, about three times the boilng temperature for atmospheric conditions at sea level. As the fluids in a geothermal system are brought to the surface, the pressure will be relieved and the hot water may flash into steam. The higher the temperature, the higher the ratio of flashed steam to residual water. In some cases, where the temperature is very high or where the pressures are very low, a vapor-dominated system is formed rather than a water-dominated system.

Areas where thermal anomalies occur are usually associated with bold, young mountains where active tectonism is taking place, or has taken place in the recent post. In addition to the hot springs, recent volcanics, geysers and fumaroles, sinter, tufa, and rock alterations also serve as surface indicators of geothermal activity.

Figure 1. Utah index map. Geologic Setting of the Roosevelt Geothermal Field.

Regionally, Roosevelt Hot Springs is located at the east edge of the Basin and Range Province in southwestern Utah (Fig. 1). It lies approximately at the north edge of the vast southwest Utah-southeast Nevada ignimbrite belt, (Cook, 1965) and in the east-west mineralized belt that includes such well-known mining distrcts as Frisco and Marysvale (Hilpert and Roberts, 1969). The field is in the "pile-up" zone of the Cordilleran fold and thrust belt (Crosby, 1973), although at Roosevelt Hot Springs the local structural motif is dominated by Cenozoic normal faulting and plutonic intrusion.

Previous investigations include water studies by Lee (1908), Mundorff (1970), and Mower and Cordova (1974). Earll (1957) geologically mapped portions of the Mineral Range. Condie (1960) investigated the petrogenesis of the Mineral Range Pluton. Recently, Petersen (1974) focused attention on the geology and geothermal potential of the Roosevelt Hot Springs area. In 1975 the University of Utah launched an in-depth research program which is now in progress.

Locally, the Roosevelt Hot Springs area is situated at the boundary between the Mineral Range and Milford graben (Figs. 2 & 3). Valley fill sediments in the graben are approximately 1500 m (5,000 ft) thick in the center of the valley. Bedrock is stepped up along several normal faults to the west flank of the Range, where the westernmost exposures consist of Precambrian (?) gnessic crystalline rocks. The metamorphic rocks are invaded in a zone of injection by Late Cenozoic granites and related intrusives. Paleozoic sedimentary rocks, exposed on the west side of the valley, terminate in the graben, having been removed by erosion.

Figure 2. Geologic Map of the Roosevelt Hot Springs area. Modified after Petersen, 1974; Liese, 1957; Earll, 1957; Hintze, 1963.

The Precambrian rocks crop out in what appears to be a structurally uplifted block that was in existance prior to intrusion of the granites. The east side of the horst block is obscured in the zone of injection. Rhyolitic flows from seven or more eruptive centers cap much of the granitic intrusive east of the hot springs area. These flows have recently been age dated by the University of Utah group and range in age between 0.4 and 2.8 m.y. Magma additions to the chamber feeding these eruptive centers are thought to be supplying the heat beneath Roosevelt Hot Springs.

Recent faulting in the vicinity of Roosevelt Hot Springs is indicated by fresh scarps in alluvium and by the cutting and displacement of hot spring deposits. The Dome fault, relatively unimportant in terms of displacement, is a major controlling structure in the hydrologic regime in the subsurface. The Negro Mag Wash fault may be equally significant. Many additional faults are indicated in the array of geophysical data. Production is controlled by heat and fluid availability and fluid availability is specifically controlled by the presence of fracture zones.

The thermal anomaly is associated with silicic crystalline rocks which are outcropping at the surface or occur at shallow depths. The reservoir is a fracture system. The depth to the reservoir top is less than 900 m (3,000 ft) over a portion of the anomaly. Locally, the fracture zones have high effective permeability yielding high flow rates of fluids in excess of 260°C, at pressures exceeding hydrostatic, and containing less than 8,000 mg/l total dissolved solids. The data strongly favors a water-dominated reservoir.

EXPLORATION PROGRAM Overview - Exploration activities began in late 1972 with reconnaissance water sampling and geophysical surveys. Reconnaissance surveys continued in 1973 in parts of Iron, Beaver, and Millard Counties. By midyear, exploration activities began to focus at Roosevelt Hot Springs, as well as other areas in the region. Evaluation based on the results of several geophysical surveys led to the conclusion that the Roosevelt area showed exceptional promise. Phillips Petroleum Company successfully bid at the July, 1974, competitive lease sale for approximately 7300 Hectares (18,000 acres) of the Roosevelt KGRA. With the approval of Plans of Operation we commenced drilling in February, 1975, and subsequently drilled nine wells: two observation holes to depths of approximately 610 m (2,000 ft) and seven designed to explore for a reservoir. After we first drilled a nonproductive hole, we drilled our discovery well which came in near the end of April. This well could not be tested or produced for mechanical reasons. Since then, three wells have been drilled that are capable of production and one that is not. Two of the productive wells require additional testing before their level of productivity is known. Our model of the subsurface is generalized in figure 3.

Water Chemistry Survey - The main structural elements of the hydrologic regime are the intermontane valley west of the field and the mountains on the east. The bedrock floor of the graben is let down by several faults to a depth of approximately 1500 m (5,000 ft) below surface near the center of the valley. Clastics, shed from the granite pluton, underlie the eastern slope of the valley adjacent to the Mineral Mountains. These give way to fine-grained lacustrine and fluviatile silts and clays in the central valley.

The Mineral Range, held up by granitic rocks, channels groundwater through fracture systems and the mantle of alluvium that occurs in the larger drainage channels in thicknesses of tens to hundreds of meters.

Figure 3. Generalized structure section through Well No. 9-1.

The potentiometric surface (Fig. 4), where it is known, generally conforms to surface configuration, sloping toward the center of the valley which, in turn, slopes north toward Blackrock at about the same rate as does the ground surface. Shallow groundwater flows down these slopes and out the north end of the valley (Mower and Cordova, 1974).

Figure 4. Map of the Milford area showing potentiometric surface contours for the principal ground water reservoir (Modified after Mower and Cordova, 1974).

Subsurface water has been sampled at virtually all existing springs, seeps and wells on and around the Roosevelt Hot Spring (Fig. 5). Water types identified chemically are a) sodium chloride, b) sodium sulfate, c) sodium bicarbonate, d) calcium chloride, e) calcium sulfate and f) calcium bicarbonate. It is clear that sodium chloride type water is leaking from the geothermal reservoir into the shallow groundwater flow system and moving down the potentiometric surface into the central portion of the valley.

Figure 5. Water characterization map showing sample locations, water types, and water quality.

Waters from an active seep on the KGRA, from Roosevelt Hot Spring and formation water from boreholes (Table 1), are characterized by high ammonia, boron, chloride (±3,800 mg/l), and total dissolved solids between 6,000 and 8,000 mg/l. It can be concluded that the seep and hot spring waters are reservoir waters on the basis of gross chemical similarity and the close agreement between reservoir temperatures and geothermometer temperatures calculated from composition of the seep and spring waters. The Na-K-Ca and the silica geothermometers are on the order of 280°C and 215°C, respectively.

Passive Seismic - Initial seismic monitoring by Phillips and the University of Utah Group indicated that the Roosevelt KGRA was aeseismic. In April, 1976, microearthquake swarm activity commenced in the KGRA area and continued to at least August, 1976, during which time seismic activity was monitored.

Table 1. Selected Water Analyses of Roosevelt KGRA

There appears to be a groundnoise anomaly associated with the reservoir (Fig. 6). The results of the survey are not straightforward; however, by selecting a frequency window between 9 and 11 hz, a discrete integrated power anomaly is approximately coincident with the extent of the reservoir as it is presently known. The selection of this window appears to minimize the amplification effects of the alluvium on distantly generated microseismic noise and local effect of wind and cultural activities affecting a wider spectrum of the groundnoise.

Figure 6. Groundnoise anomaly map showing contoured integrated power in frequently window 9-11 hz. C.I. - 4db.

Gravity - Reconnaissance gravity surveys (Fig. 7) indicate a thickness of valley fill of approximately 1500 m (5,000 ft) west of the KGRA, a positive anomaly centered over the Precambrian gneiss outcrops on the west side of the range, and an irregular low over the geothermal field. Linear gradient zones express the gross fault pattern which agrees well with fault positions determined independently by other data.

Figure 7. Bouguer anomaly map of the Roosevelt Hot Spring area. Contour interval is 1 mgal.

Ground Magnetics - A ground level total field magnetic survey (Fig. 8) was carried out over the KGRA with station density of two per km^2 —— something less than a detailed survey. A quantitative solution is made virtually intractable by 1) the complex structure, 2) composite nature of the Cenozoic pluton, 3) seven or more individual rhyolite flows and domes, 4) injection of Cenozoic magmas into Precambrian gneiss, 5) late dike intrusion of more basic rocks, 6) local alteration, and 7) magnetic field reversals.

To date, wells drilled that are capable of commercial production are situated within magnetic lows. Magnetic trends are coincident with the Dome and Negro Mag Wash faults. Further, right-lateral offsetting of magnetic highs in the vicinity of the intersection of the Dome and Negro Mag Wash faults suggest possible strike-slip faulting.

Figure 8. Total intensity ground level magnetics. Contour interval is 20 gammas.

Electrical Surveys - An early bipole-dipole survey was undertaken before exploration had sharply focused on the Roosevelt area and a highly conductive zone characterized by conductivities in excess of 2000 mhos was located in the valley center. Subsequently, the cause of this anomaly has been determined to be a combination of thick porous valley fill sediments, a thick clay section, and saline reservoir water leaking downslope from Roosevelt, but lacking abnormal heat. A resisitivity map, based on measurements made only from a source dipole near the south end of the KGRA, shows an anomaly of less than 25 ohm-m obtaining along the trace of the Dome fault and trending toward Roosevelt Hot Spring (Fig. 9). This anomaly, however, lies in a low resistivity embayment about 50 ohm-m.

An MT survey was carried out using a 4.8 km (3 mi) spacing interval between stations. With this reconnaissance spacing only four stations fell on the KGRA. However, these four stations gave highly significant results, namely a low resistivity interval with top varying between 900 and 1370 m (3,000 and 4,500 ft) below surface. These values, depths and resistivities are readily reconciled with the results of subsequent drilling.

A detailed MT survey is presently undergoing analysis.

Figure 9. Resistivity map based on bipole-dipole data from a single source. Contoured in ohm-m.

Temperature Gradient Survey - The temperature gradient map (Fig 10) is based on a total of thirty-nine holes and combines Phillips results with data from Petersen (1975) and Whelan (University of Utah, personal communication). Depths of gradient holes vary from 60 to 610 m (200 to 2,000 ft). Five holes in the center of the thermal anomaly exceed 40°C/100 m. The anomaly is elongate north-south with a change in trend to the northwest in the northern third of the anomaly. The north-south trace of the Dome fault centers on the anomaly, and the east-west Negro Mag Wash fault is coincident with the zone of the change in trend. Gradients are generally conductive, that is, they are linear. The production wells demonstrate that this condition holds until a fluid producing zone is intersected.

Exploration Program Summary - 1. There are discrete and strong temperature gradient, chemical, gravity, MT and groundnoise anomalies that are mutually coincident and lie on that portion of field so far proven by drilling.

2. Weaker roving dipole and magnetic anomalies are coincident with a portion of the thermal anomaly.

3. Microearthquake swarms are apparently associated with the Roosevelt geothermal field.

4. No single geological, geochemical, or geophysical survey, including temperature gradients, is adequate to correctly identify the position and quality of the Roosevelt geothermal field.

5. Drilling ± 610 m (2,000 ft) deep slim observation holes provide information on deep temperature gradients, absolute temperatures, stratigraphy, hydrologic distortion of temperature gradients, and potential drilling problems, and is a cost effective part of a geothermal exploration program on good prospects.

6. The reconciliation of the several surveys, measuring independent parameters, into an integrated concept of the subsurface, adequately defined the geothermal system and provided a rational basis for investing in deep exploratory boreholes.

7. Drilling seven deep tests has generally borne out the subsurface model conceived earlier by exploratory survey results.

8. The reservoir consists of fracture systems within granitic and gneissic rocks.

9. The resource is a liquid-dominated, high temperature, low salinity fluid.

Figure 10. Temperature gradients in the Roosevelt Hot Springs area. Contour Interval is $10^{\circ}C/100$ m. (Includes data from Petersen, 1975, and Whelan, personal communication, 1976)

DRILLING PROGRAM Well Control - The Roosevelt drilling program has gone through continuous evolution since its inception in January, 1975. The casing program calls for three strings of casing set to a depth of approximately 550 m (1,800 ft) with the remainder of the well completed as open hole (Fig. 11).

The detailed program is ± 18 m (60 ft) of 50.8 cm (20 in) OD conductor pipe, ± 183 m (600 ft) of 35 cm (13 3/8 in) OD surface pipe and ± 550 m (1,800 ft) of 24.5 cm (9 5/8 in) OD long string. The surface pipe is set a minimum of 45 m (150 ft) into consolidated rocks to anchor the casing firmly. The long string is set above the depths at which the geothermal reservoir has been encountered in other wells. A massive blowout preventer stack is then installed before drilling into the reservoir.

Figure 11. Schematic diagram of casing program used at Roosevelt Hot Springs.

The blowout preventer program has undergone the most change with time. A 50.8 cm Hydril annular BOP is mounted on the conductor pipe for security in the unlikely event a very shallow geothermal reservoir is encountered. A double preventer, providing one set of pipe rams and one set of blind rams, an annular BOP and a Grant rotating head are attached to the surface pipe. Maximum protection is provided on the long string which is required to withstand the heat and pressures of the geothermal reservoir fluids. Well control is provided here by three sets of pipe rams, a single set of blind rams, a Hydril annular BOP which will close on anything and a Grant Rotating Head which diverts gas or fluids from the rig floor (Fig. 12). The Grand Head and the second double ram preventer are elements added to the BOP stack used on the earlier Roosevelt wells. These additions necessitated the construction of ramp and base to raise the rig floor an additional 2.5 m (8 ft).

The completed wells deliver large volumes of fluids at elevated temperatures. As a result, like the BOP stack, the completion tree is also a massive assembly including a Braden head, expansion spool, double gates and various wing valves.

Figure 12. Schematic diagram of the BOP stack used after setting 24.5 cm (9 5/8 in) casing.

Figure 13. Schematic diagram of the completion tree.

Drilling Problems - Drilling difficulties fall into three categories, namely problems caused by 1) heat, 2) pressure, and 3) rock conditions. Excessive heat is the most serious obstacle to successfully complete a well at Roosevelt Hot Springs; of least concern is control of pressure.

Bit costs, because of heat, are much higher than for oil or gas tests drilled to similar depths. Heat affects bits by accelerating deterioration of the O-ring seals long before the cutting structures are dulled. Continued use of a bit after loss of the seals might result in destruction of the bearings and eventual loss of a cone. We have learned by experience to pull the bits prior to destruction of the O-ring seals.

Even more serious is the effect heat has on the rubber elements in the BOP stack. Heat causes the ram rubbers to melt and the rubber element in the Grant rotating head to fail. The drilling medium below the 24.5 cm (9 5/8 in) casing is clear water. In the event of a well kick, one or two sets of rams depending upon the working temperatures are closed. To change the ram rubbers, the well is then killed with a saturated solution of salt water. Subsequently the salt water is returned to the holding tank and drilling continued with clear water.

The water used in drilling, heated by contact with hot rock and mixing with reservoir fluids, is cooled by circulating through the reserve pit, instead of through the flow line to the mud pits. Temperature of the returning fluids are kept within acceptable limits by varying pressure through the choke manifold.

r example, a lowering of back pressure increases flow and temperature. Thus essure is used to control fluid flow and temperature while allowing for continued drilling in a hostile environment.

The only major down hole problem encountered is the intersection of open ractures with total loss of returns. To pass these zones safely, it is sometimes ecessary to switch from clear water to mud plus lost circulation material, or o use compressed air and water to form an air-water mist that reestablishes eturns. Sloughing into or bridging off of the hole by collapse of the walls as not been a problem.

OCK BEHAVIOR Mechanical properties of Roosevelt rocks are partially revealed in drilling rates. The lithologies of rocks encountered in drilling thus far, in order of decreasing abundance, are granodiorite, granite, injection gneiss, quartz diorite, quartz monzonite, and lamprophyre. Drilling rates average 3.7 m/hr (12.15 ft/hr) in unfractured sections of the bore. There is no statistical difference in drilling rates that can be attributed to lithology. All rocks except lamprophyre are medium to coarse grained, and with the exception of injection gneiss, are adequately described as hypidiomorphic granular textures. The weak to strongly foliated gneisses are characterized by the greatest variation in drilling rates.

The variation in rates is greatest at the top of the hole, but decreases with depth. With drilling undertaken thus far there are no established trends toward higher or lower drilling rates in unfractured sections as a function of depth; therefore, there are no apparent correlations among rates, increasing temperatures, and increasing confining pressures. Temperatures to 265°C (507°F) and confining pressures to at least 550 bars have been encountered. The lack of a correlation between drilling rates and increasing temperature with depth appears to be borne out by the absence of a penetration rate change at times when specific operations are undertaken to cool the hole such as abruptly circulating cold drilling fluid concomitant with pumping at higher volumes.

Drilling rates clearly increase in fracture zones and indeed, this is the immediate indication along with the tendency of the rig to vibrate that fractures have been encountered. Average drill rates of 23 m/hr (75 ft/hr) have been maintained through a 15 m (50 ft) thick fractured zone. Cuttings from most fracture ones show increased alteration which doubtlessly aids the penetration rate; however, alteration is never excessive in the fracture zones in the deep subsurface and the rock still behaves in a brittle manner as it interacts with the bit. Instantaneously the drilling rate approached infinity as the bit breaks into open fractures. The free fall of the drill string, on several occasions, was over distances measured in centimeters and meters. One fall in Well No. 13-10 was on the order of one meter.

Terminating quartz crystals have been noted in the cuttings from several wells during the passage through a fractured zone. There is little doubt that these come from drusy linings in open fractures. The greater than hydrostatic pressures in the productive wells enhance the rock's ability to support the stress differential associated with open fractures. The fact remains, however, that that rock behaves in a brittle fashion at the temperatures and pressures extant in the subsurface at the Roosevelt Geothermal Field.

Acknowledgments

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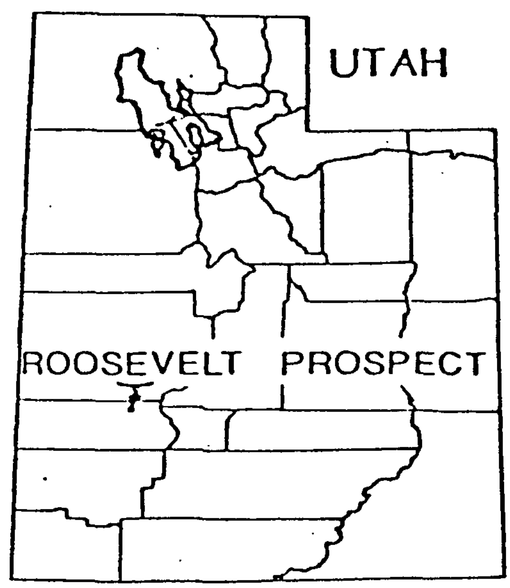
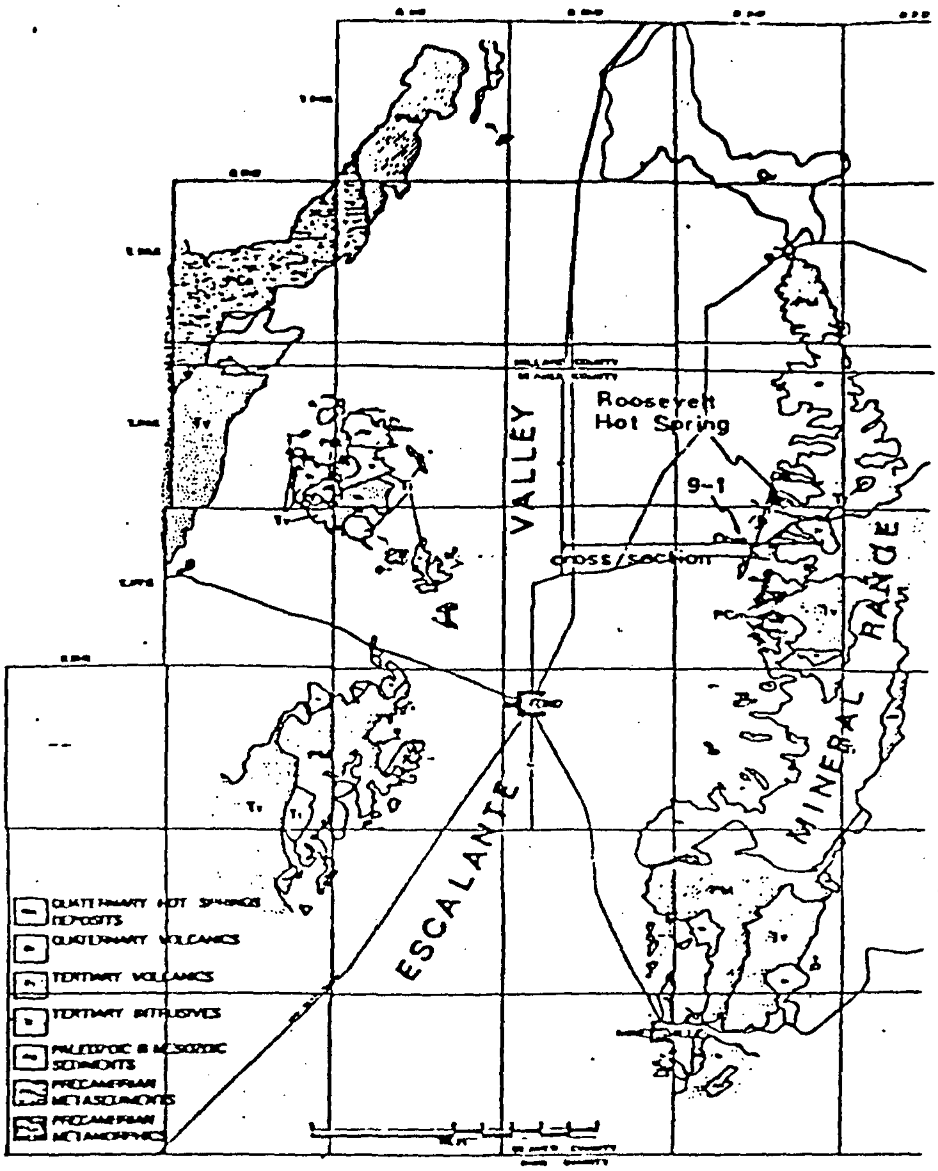


fig. 1



Roosevelt
Hot Spring
9-1
cross section
VALLEY
ESCALANTE
MINERAL RANGE
QUATERNARY HOT SPRINGS DEPOSITS
QUATERNARY VOLCANICS
TERTIARY VOLCANICS
TERTIARY INTRUSIVES
PALEOZOIC & MESOZOIC SEDIMENTS
PRECAMBRIAN METASEDIMENTS
PRECAMBRIAN METAMORPHICS

fig. 2

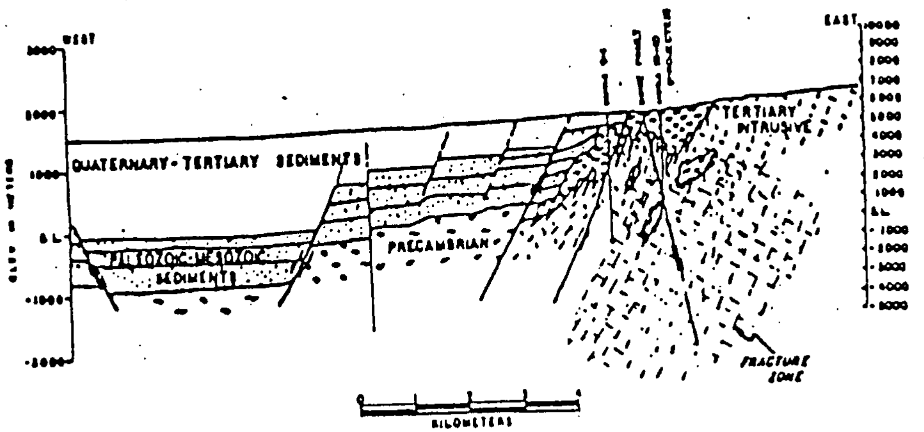
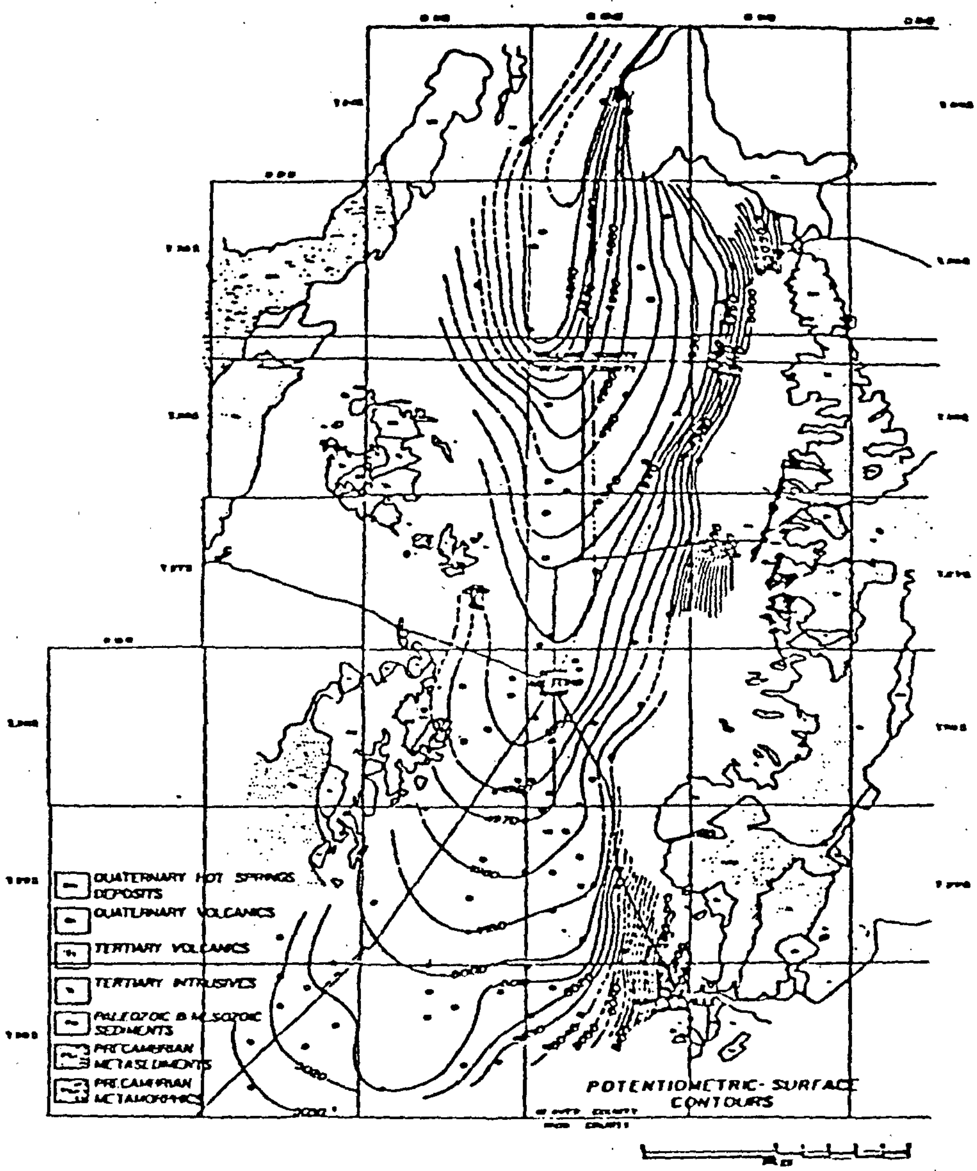

WEST
EAST
QUATERNARY-TERTIARY SEDIMENTS
PRECAMBRIAN
SEDIMENTS
TERTIARY INTRUSIVE
FRACTURE ZONE
KILOMETERS
0
1
2
3
4
S.L.
-1000
-2000

fig.3


QUATERNARY HOT SPRINGS DEPOSITS
QUATERNARY VOLCANICS
TERTIARY VOLCANICS
TERTIARY INTRUSIVES
PALEOZOIC & MESOZOIC SEDIMENTS
PRECAMBRIAN METASEDIMENTS
PRECAMBRIAN METAMORPHICS
POTENTIOMETRIC-SURFACE CONTOURS

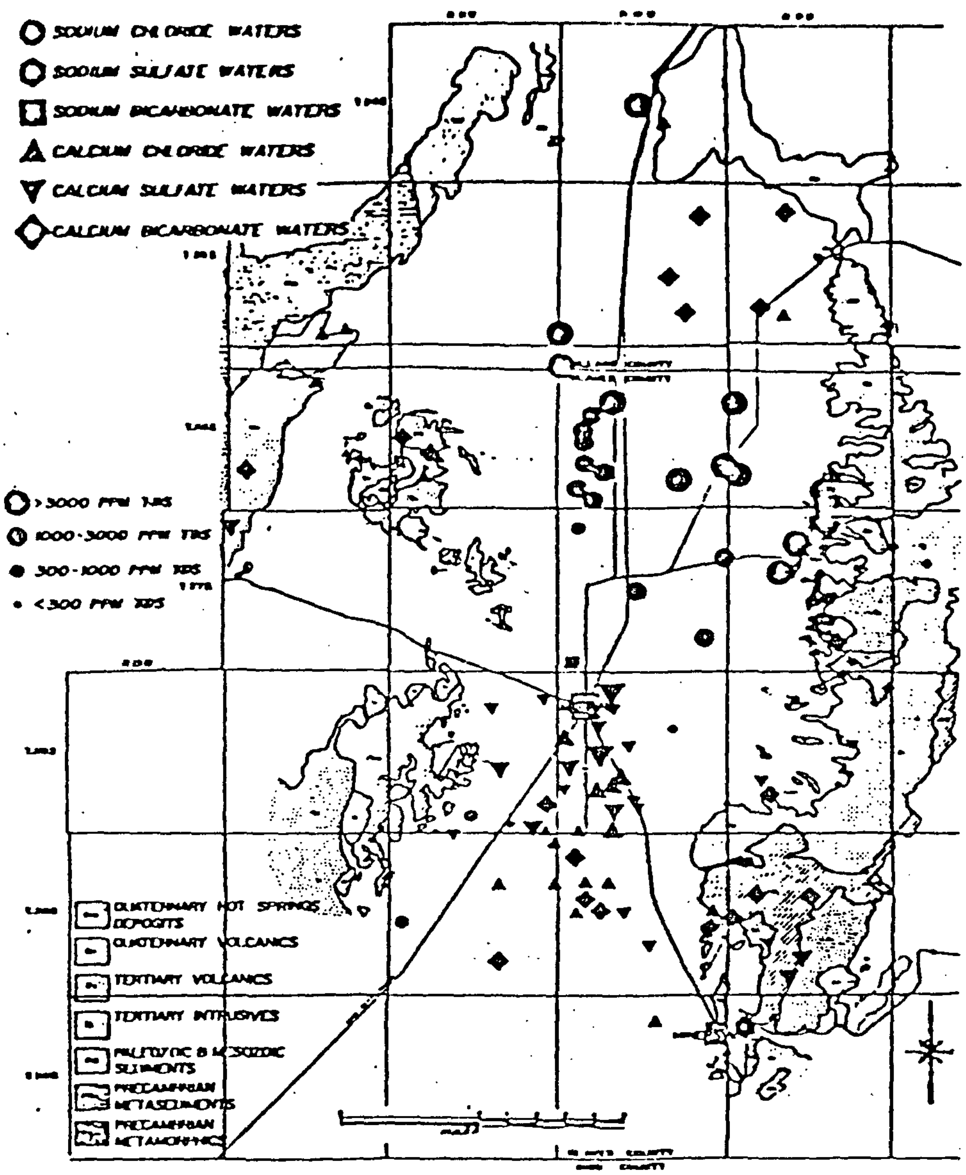
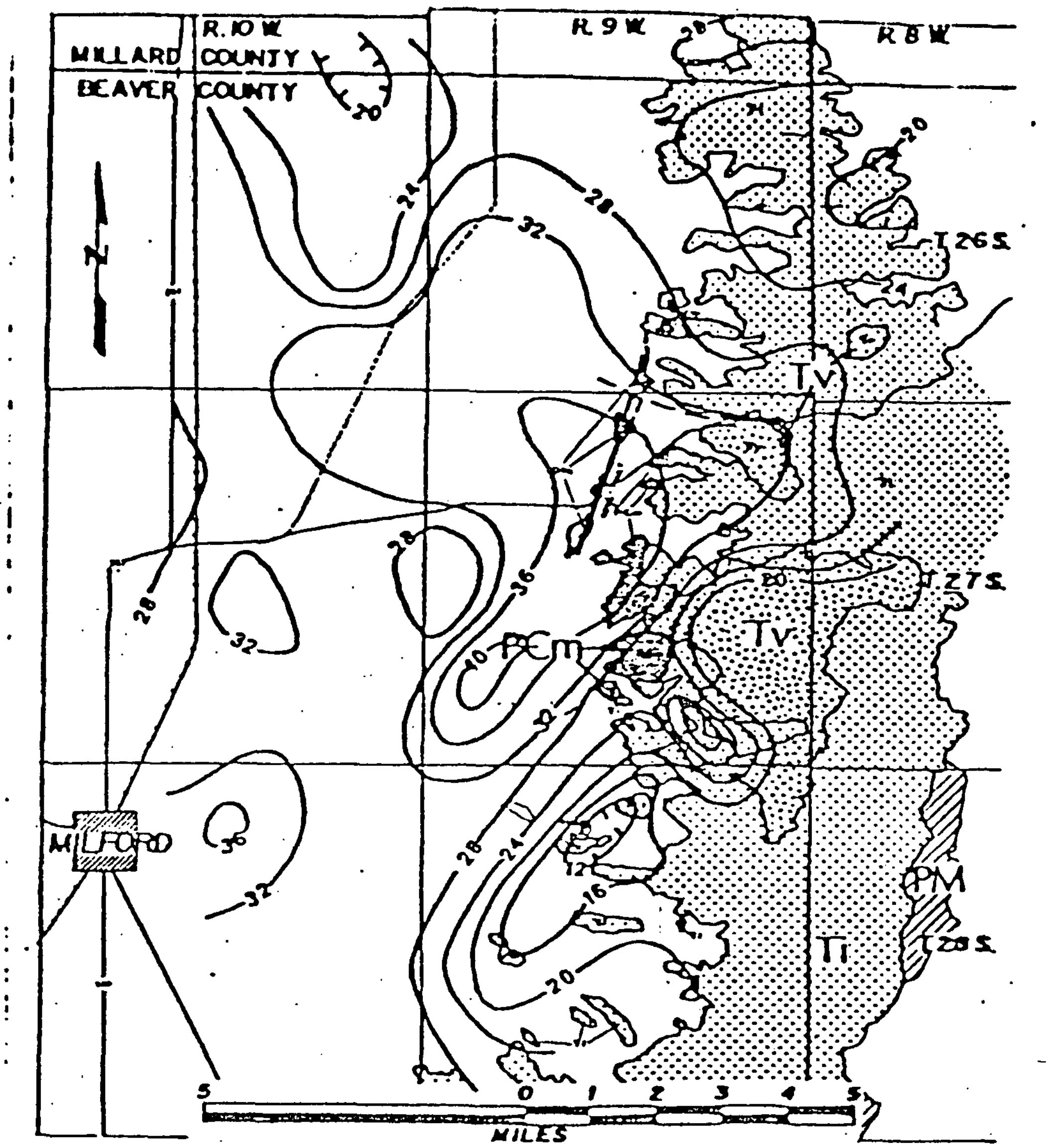

SODIUM CHLORIDE WATERS
SODIUM SULFATE WATERS
SODIUM BICARBONATE WATERS
CALCIUM CHLORIDE WATERS
CALCIUM SULFATE WATERS
CALCIUM BICARBONATE WATERS
>3000 PPM TDS
1000-3000 PPM TDS
300-1000 PPM TDS
<300 PPM TDS
QUATERNARY HOT SPRINGS DEPOSITS
QUATERNARY VOLCANICS
TERTIARY VOLCANICS
TERTIARY INTRUSIVES
PALEOZOIC & MESOZOIC SEDIMENTS
PRECAMBRIAN METASEDIMENTS
PRECAMBRIAN METAMORPHICS

fig. 5



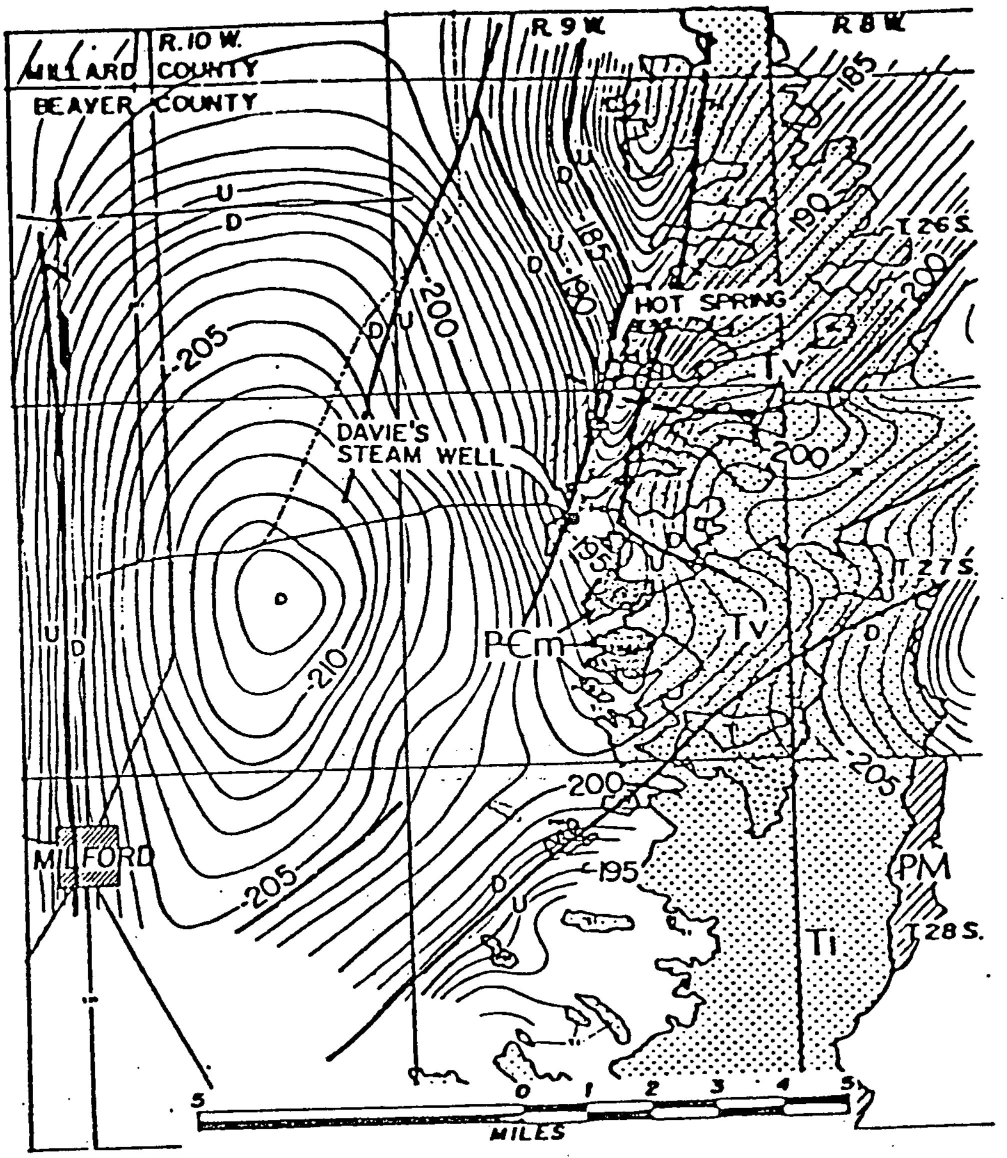
ground noise anomaly

fig. 6

SELECTED WATER ANALYSES of ROOSEVELT KGRA

	Roosevelt Hot Springs	Roosevelt Hot Springs	Roosevelt Seep	Roosevelt Seep	54-3	3-1
Date	11-4-50	9-11-57	5-9-73	8-15-75	8-26-75	5-25-75
Temperature (°C)	85	55	17	28	> 260	> 205
Silica (ppm)	405	313	76	107	560	560
Calcium (ppm)	19	22	113	107	101	8.0
Magnesium (ppm)	3.3	0	17	23.6	0.24	0.01
Sodium (ppm)	2080	2500	2400	1800	2000	2437
Potassium (ppm)	472	488	378	280	410	448
Bicarbonate (ppm)	156	156	536	300	200	180
Sulfate (ppm)	65	73	142	70	54	59
Chloride (ppm)	3810	4240	3800	3200	3400	4090
Fluoride (ppm)	7.1	7.5	5.2	3.3	5.0	5.0
Nitrate (ppm)	1.9	11	TR	TR	TR	0.1
Boron (ppm)	—	38	37	29	29	25
Lithium (ppm)	—	0.27	—	17	19.0	20.0
TDS (ppm)	7040	7800	7506	5948	6442	7067
pH	—	7.9	8.2	6.43	6.5	6.3
1. Na-K-Ca [illegible]	295	285	247	239	294	273
Na-K [illegible]	307	282	250	248	290	294

1. [illegible] 2. [illegible]


R.10 W.
MILLARD COUNTY
BEAVER COUNTY
R 9 W.
R 8 W.
U
D
185
190
T 26 S.
200
205
HOT SPRING
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DAVIE'S
STEAM WELL
195
T 27 S.
210
PЄm
200
MILFORD
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PM
Ti
T 28 S.
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MILES

fig. 1

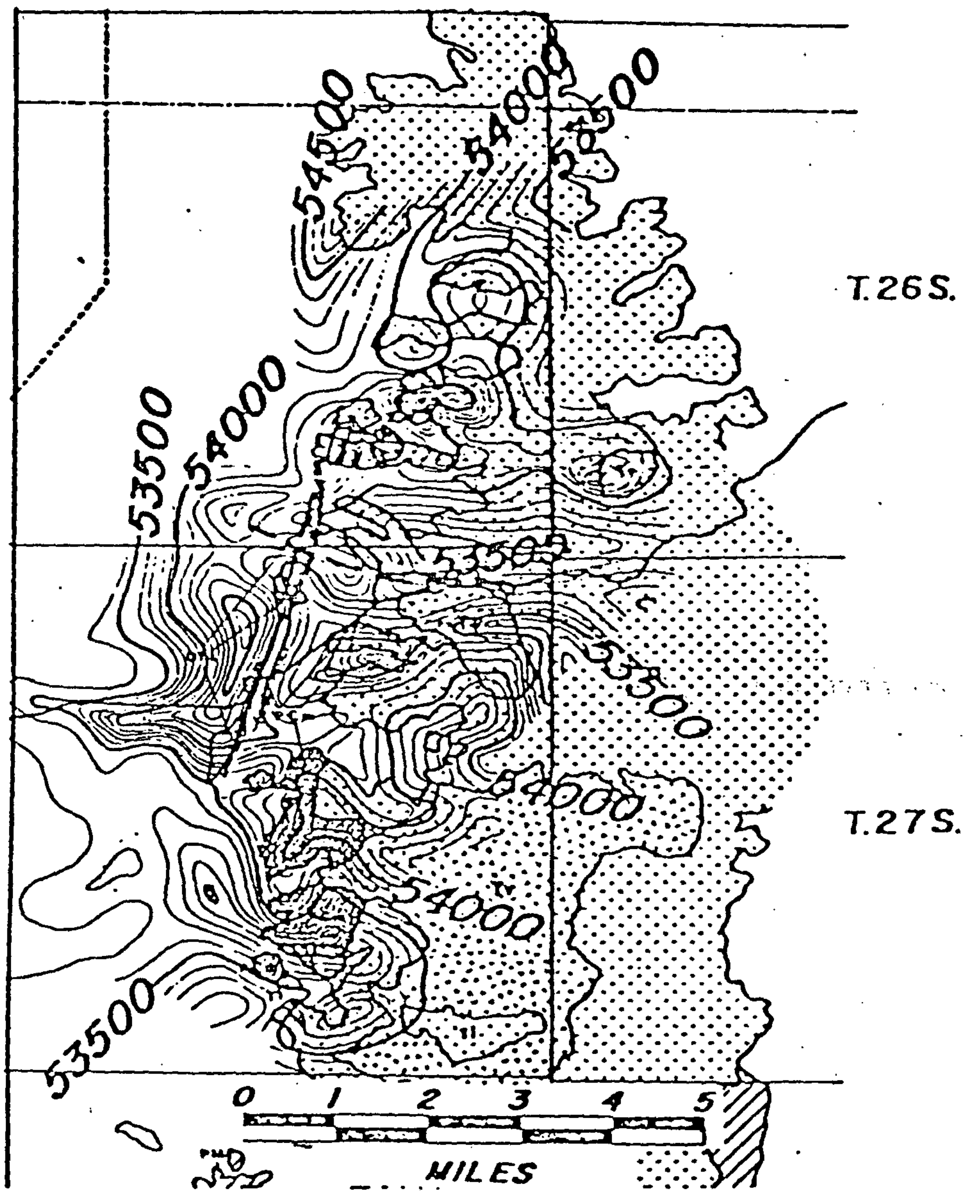
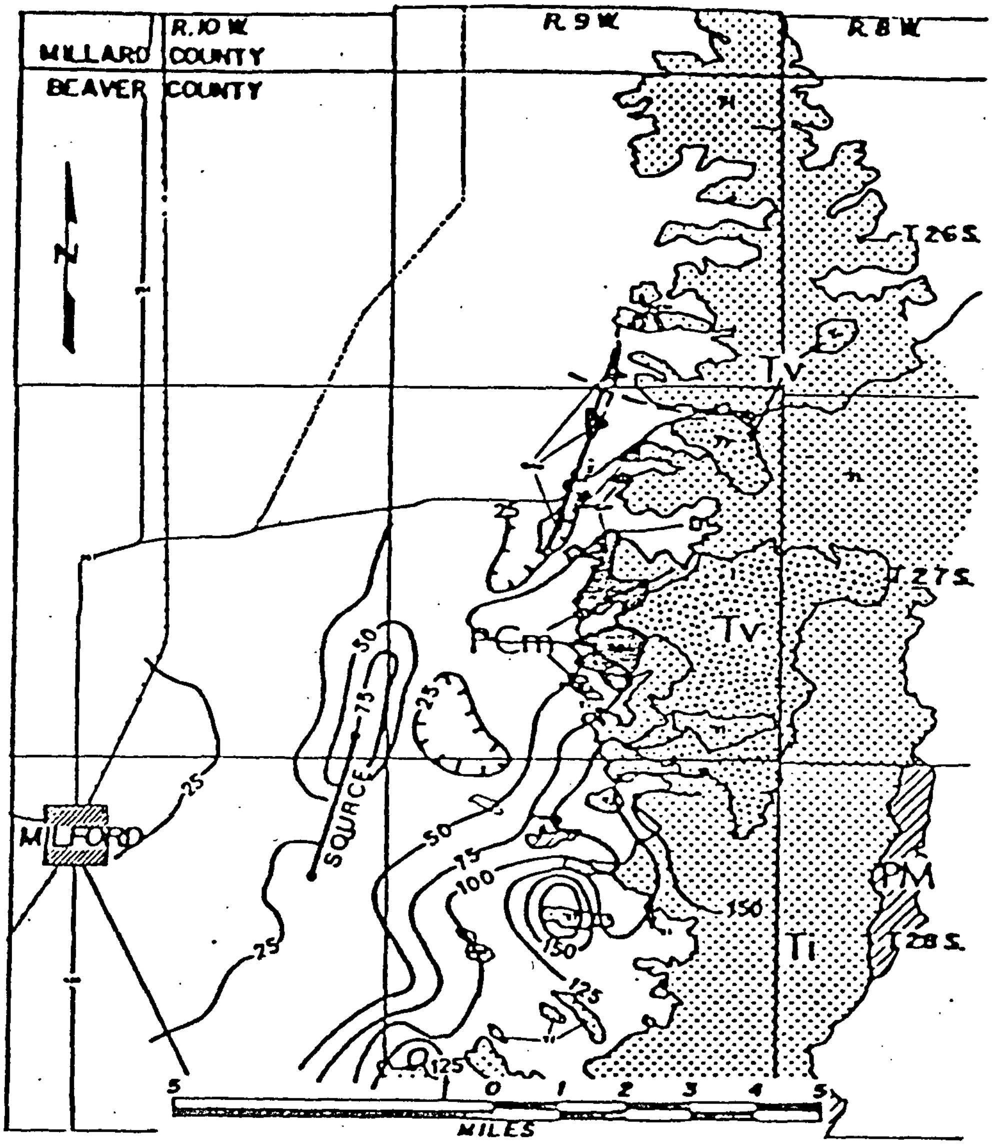


Fig. 8. Total Magnetic Intensity


R.10 W.
R.9 W.
R.8 W.
MILLARD COUNTY
BEAVER COUNTY
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T.26 S.
T.27 S.
T.28 S.
Tv
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Ti
PM
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SOURCE
25
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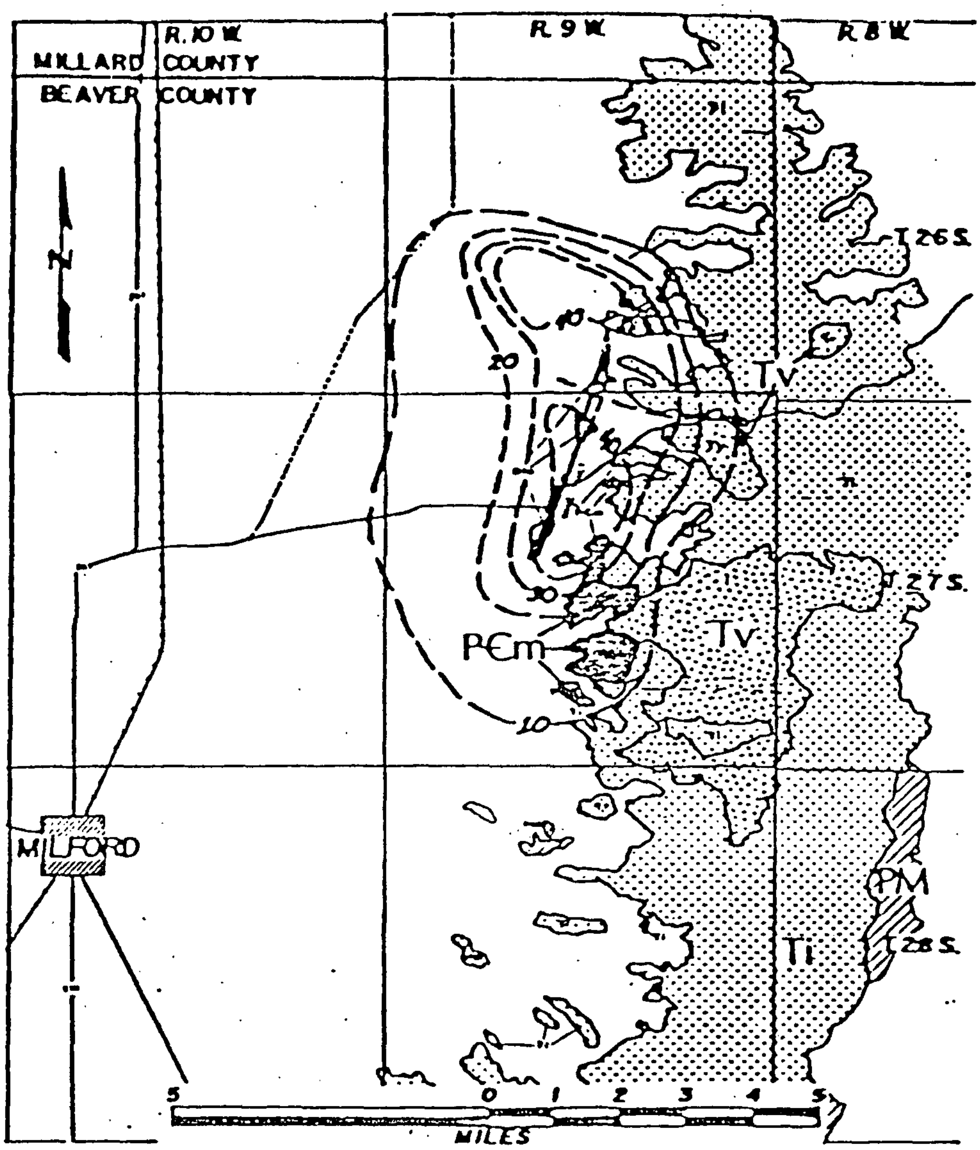
Fig ~~1~~ 9


R. 10 W.
R. 9 W.
R. 8 W.
MILLARD COUNTY
BEAVER COUNTY
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T. 26 S.
T. 27 S.
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40
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30
10
Tv
Tv
PЄm
Ti
PM
MILFORD
5
0
1
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3
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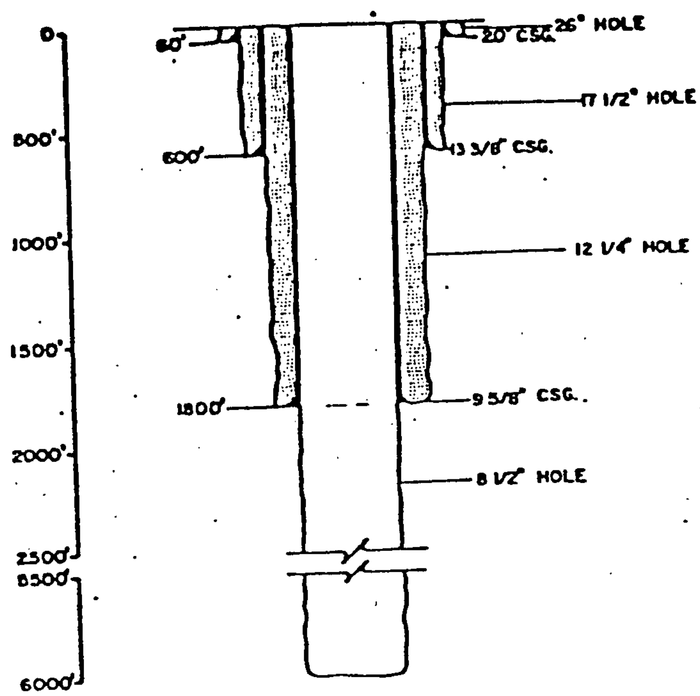
MILES

fig. XI 10



SCHEMATIC OF TEST WELL
PERMANENT CASINGS

... (JULY, 1968), P. 1279-1290, 9 FIGS.

LOWER CAMBRIAN AND UPPER PRECAMBRIAN STRATA OF BEAVER MOUNTAINS, UTAH[1]

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ABSTRACT

Cambrian Prospect Mountain Quartzite and seven underlying units, tentatively considered Precambrian, are mapped in the Beaver Mountains. This concordant section begins where alluvium covers the base of the oldest unit and ends at the contact between the Prospect Mountain Quartzite and the overlying Pioche Shale.

Assigned to the Precambrian is 5,145 ft of strata. These rocks are slate, argillite, marble, metasiltstone, quartzite, and conglomerate, which have undergone very low-grade regional metamorphism.

The uppermost Precambrian unit, a purple conglomeratic quartzite, is correlated with the Mutual Formation of north-central Utah. The overlying tan and pink Prospect Mountain Quartzite is at least 4,000 ft thick.

Lithologic correlation of the Beaver Mountains section with the strata described by Christiansen (1952) from the Canyon Range in central Utah is established. Tentative correlations with the late Precambrian McCoy Creek Group of eastern Nevada and western Utah (Misch and Hazzard, 1962) are suggested.

INTRODUCTION

Lower Cambrian and upper Precambrian rata in the Beaver Mountains, Millard County, tah, are of regional interest because they form e principal Precambrian outcrops midway be- een the late Precambrian McCoy Creek Group eastern Nevada (Misch and Hazzard, 1962) d unnamed Precambrian strata described by ıristiansen (1952) in the Canyon Range of cen- ıl Utah (Fig. 1). The stratigraphic and litho- ʒic similarities of the Precambrian rocks of the ınyon Range and the Beaver Mountains are of uctural significance, because they support the ggestions by Misch (1960) and Miller (1966) ıt the Precambrian and Lower Cambrian klippe the Canyon Range (Christiansen, 1952) is a nnant of a once continuous thrust plate that luded the Canyon, Beaver, San Francisco, and ıh Wah Ranges (Fig. 1).

The geologic map of southwestern Utah intze, 1963) shows undifferentiated metasedi- ntary rocks, chiefly quartzite and argillite, that assigned to the upper Precambrian and over- ıg Cambrian Prospect Mountain Quartzite in Beaver Mountains. This is based on unpublished reconnaissance work by W. C. Jeffs, D. M. Lemmon, and others (Hintze, 1963).

In the writer's study, the Precambrian sequence was divided into seven formations. These formations were mapped at a scale of 1:20,000 on a topographic base (Fig. 2). Thicknesses of the formations were scaled from structure sections. Approximately 200 hand specimens were collected and 80 of them were thin-sectioned. Fifteen additional thin sections of specimens collected from the Mutual Formation at other localities were examined.

The Precambrian and Cambrian rocks of the Beaver Mountains show the effects of regional metamorphism in the lower greenschist facies. In eastern Nevada (Fig. 1, locs. 2, 3), Misch (1960) and Misch and Hazzard (1962) showed that the regional metamorphism of upper Precambrian and Cambrian rocks occurred during the first stage of a mid-Mesozoic orogeny prior to thrust faulting. Christiansen (1952) suggested that large-scale thrusting probably took place during Late Jurassic to Early Cretaceous time in the Canyon Range (Fig. 1, loc. 7). Thus, regional metamorphism of rocks of the Beaver Mountains probably occurred during the same Mesozoic orogeny.

STRATIGRAPHY

Cambrian Prospect Mountain Quartzite and seven underlying formations, tentatively considered upper Precambrian, have been mapped near Beaver Mountain (Fig. 2). This concordant section begins where alluvium covers the base of the

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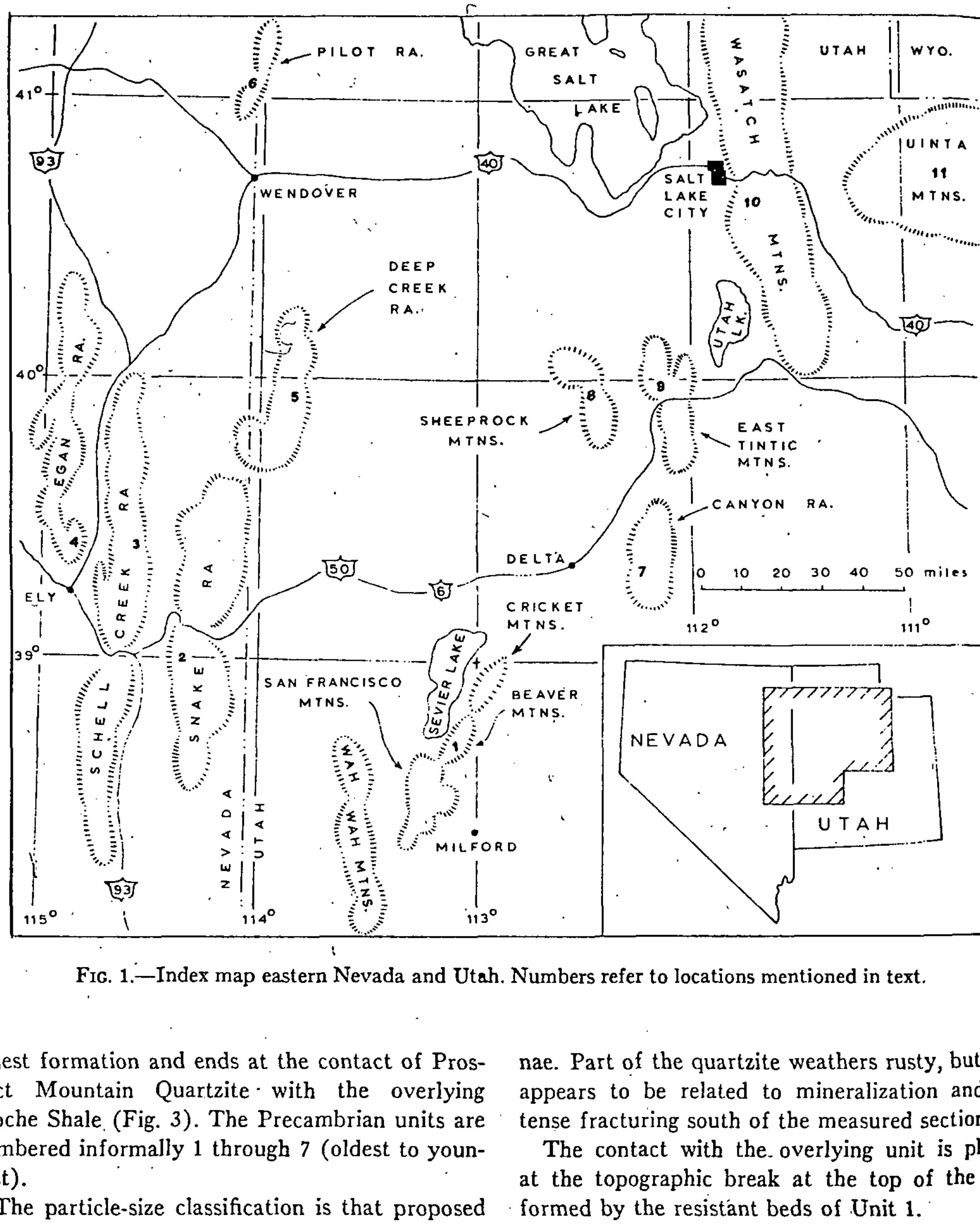

FIG. 1.—Index map eastern Nevada and Utah. Numbers refer to locations mentioned in text.

oldest formation and ends at the contact of Prospect Mountain Quartzite with the overlying Pioche Shale (Fig. 3). The Precambrian units are numbered informally 1 through 7 (oldest to youngest).

The particle-size classification is that proposed by Wentworth (1922) and terms used to describe thickness of stratification are those of McKee and Weir (1953).

Unit 1.—The total thickness of this unit is not known, because the base is covered by alluvium; however, 970 ft was measured.

Unit 1 consists of thick-bedded, light-gray, cross-bedded quartzite. It is mostly medium to coarse grained, but locally is very coarse. Near the top of the unit are a few hematitic slaty laminae. Part of the quartzite weathers rusty, but this appears to be related to mineralization and intense fracturing south of the measured section.

The contact with the overlying unit is placed at the topographic break at the top of the cliff formed by the resistant beds of Unit 1.

In thin sections the quartzite is seen to be composed almost entirely of quartz clasts with minor interstitial sericite and chlorite and traces of feldspar, hematite, chert, zircon, tourmaline, stilpnomelane, and apatite. The quartz clasts are cemented by quartz outgrowths, the original grain margins being outlined by very thin hematite coatings. Stilpnomelane is rare and appears to be related to postdiagenetic introduction of iron.

Unit 2.—Unit 2 forms a bench or slope on in-

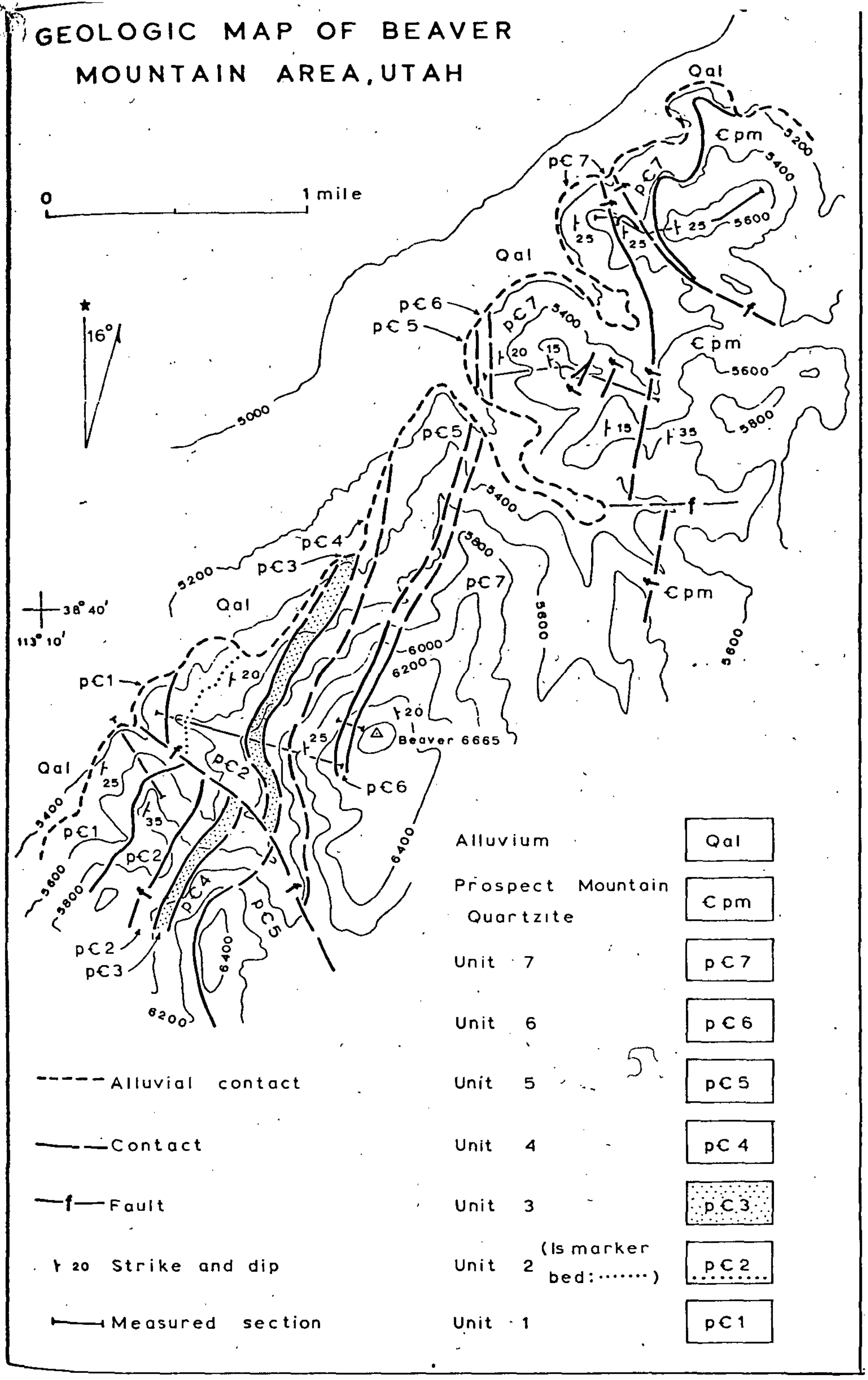

FIG. 2.—Generalized geologic map of Beaver Mountain area, Millard County, Utah.

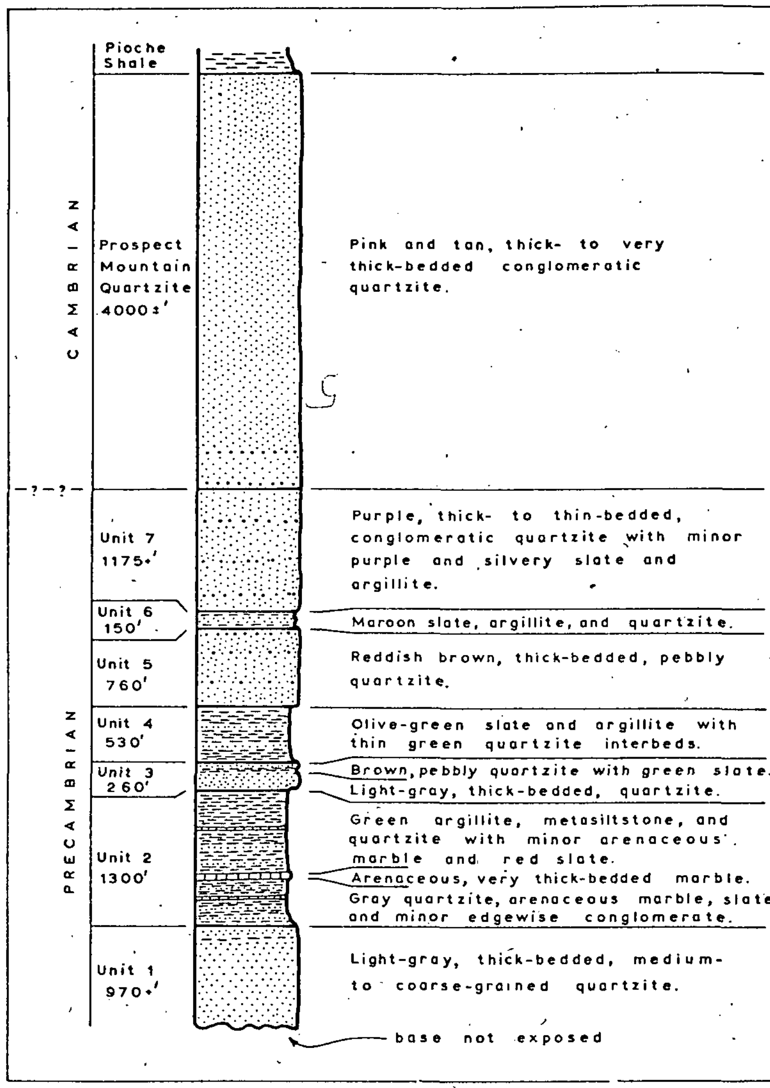

FIG. 3.—Composite columnar section of upper Precambrian and Lower Cambrian strata of Beaver Mountain Utah. Thicknesses in feet.

terbedded quartzite, slate, marble, and argillite. The topographic break at the lower contact reflects a lithologic change; bedding becomes thinner and the proportion of slate and argillite increases upward. Unit 2 measured 1,300 ft, but this is probably not the true stratigraphic thickness because there is considerable internal deformation, including shearing and small-scale disharmonic folding (Fig. 4).

The lower 500 ft of the unit consists of thin-bedded, very fine- to coarse-grained, gray quartzite with intercalated red and olive-green s[illegible] laminae and thin-bedded arenaceous marble. [illegible] marble is made of finely crystalline sheared [illegible] cite containing well-rounded quartz grains. N[illegible] the top of this sequence is a minor zone of in[illegible] formational edgewise conglomerate consisting [illegible] clasts up to 6 cm long of fine-grained quartzite [illegible] a coarse-grained quartzite matrix.

Above this sequence and 800 ft below the [illegible] of Unit 2 is a 25-ft cliff of very thick-bed[illegible] sheared, quartzose marble (Fig. 5). Very [illegible]

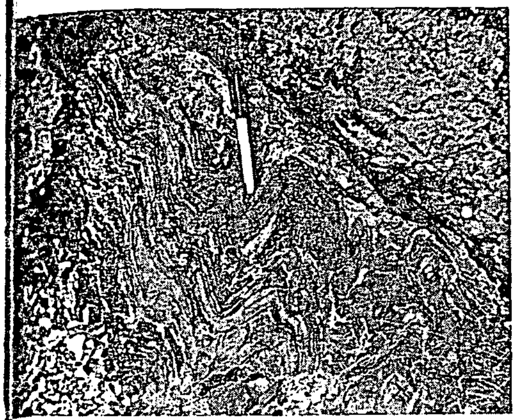

FIG. 4.—Small-scale disharmonic folds within Unit 2, upper Precambrian of Beaver Mountains, Utah. Thick bed of arenaceous marble in upper right and folded slate and calcareous sandstone laminae below. Scale shown by marking pen.

grained, angular quartz in thin lenses probably represents chert lenses that were sheared and recrystallized during regional metamorphism.

Above the cliff-making marble is a bench formed on the upper 800 ft of Unit 2. This sequence consists chiefly of thinly interbedded slate, argillite, metasiltstone, quartzite, and marble.

The slate is weakly schistose parallel with bedding and is deep red because of abundant hematite. Argillite and metasiltstone are not schistose and are mostly tan, olive green, and greenish gray. Most of the marble contains quartz and locally grades into calcareous sandstone. Tan to olive-green quartzite predominates and ranges from very fine to coarse grained.

The upper contact of Unit 2 generally is covered with quartzite float from the overlying unit. The highest occurrence of greenish argillite float characteristic of Unit 2 coincides with the base of the cliff formed by Unit 3; this topographic break is mapped as the contact.

Petrographic study of the rocks of Unit 2 shows that the quartzite is composed principally of quartz clasts with minor interstitial sericite and chlorite or rarely calcite. Clasts of sodic plagioclase, potassium feldspar, hematite, chert, muscovite, biotite, zircon, and tourmaline are present in trace amounts. Some fine-grained clots of intergrown sericite and chlorite appear to be recrystallized shale or clay fragments. Stilpnomelane is rare and is related to late introduction of iron along fractures.

The marbles are calcitic and typically contain minor quartz. Siderite and dolomite rhombs are common in some specimens. Calcite lamellae are curved and crystals are elongate parallel with shearing in the more deformed marble. Evenly distributed, well-rounded, coarse quartz grains are probably of clastic origin. Other specimens contain small lenses of angular quartz, suggesting recrystallized chert lenses.

Unit 3.—This cliff-making unit is made up of 260 ft of quartzite with minor slate and metasiltstone near the top. The lower part of Unit 3 consists entirely of thick-bedded, medium- to coarse-grained, whitish and light-gray quartzite.

Approximately 50 ft below the top of the unit is an abrupt upward change from light-gray quartzite to dark-brown quartzite containing quartz granules and pebbles up to 2 cm in diameter. These larger clasts are subrounded and are composed of white quartz with minor jasper.

There is a 15-ft bench formed on olive-green slate and metasiltstone near the top of the unit. The slate is overlain by 10 ft of very thick-bedded, medium- to coarse-grained, dark-gray quartzite. The top of this resistant quartzite is assigned as the upper contact of Unit 3.

Quartzites of Unit 3 consist principally of quartz with traces of chert, feldspar, zircon, tourmaline, and apatite. Interstitial hematite, sericite, and chlorite are present. The dark quartzite is colored by the hematite coating of clasts; the detrital texture is preserved by this coating although quartz outgrowth has resulted in a granoblastic fabric.

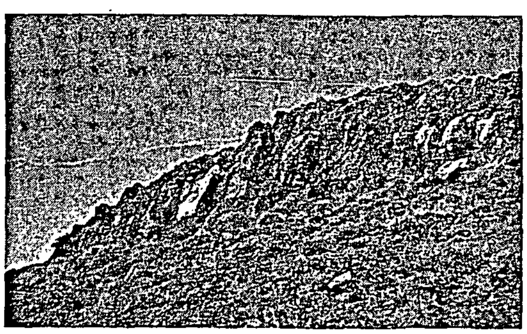

FIG. 5.—Marker bed of cliff-making marble 25 ft thick in Unit 2, upper Precambrian of Beaver Mountains, Utah.

In the slate, a schistosity (s_2), parallel with bedding (s_1), is marked by moderately well-aligned sericite and chlorite. Minor clastic biotite is present, as well as very small porphyroblasts of syn- and late-kinematic chlorite and stilpnomelane.

Unit 4.—Unit 4 consists of 530 ft of bench-forming, olive-green slate and argillite with minor thin quartzite intercalations. The basal contact has been described. At most places the upper contact is talus covered; however, where exposures are good the contact is marked by an abrupt upward change to thick-bedded, reddish quartzite.

Schistosity in the slate, defined by sericite and chlorite, is parallel with bedding and is moderately to strongly developed. Argillite is nonschistose and commonly contains abundant quartz sand and silt; as the quartz content increases, these rocks grade into argillitic quartzite. Minor clastic feldspar, chlorite, and biotite are present. Small chlorite porphyroblasts are common. Round aggregates up to 0.25 mm in diameter are composed of finely intergrown sericite and chlorite; these intergrowths are probably recrystallized shale or clay fragments.

The quartzite is very fine to medium grained and contains abundant chlorite and sericite matrix. This matrix preserves the clastic texture. Clasts of feldspar, biotite, muscovite, chlorite, zircon, tourmaline, and apatite are present in trace amounts. Hematitic staining along fractures is accompanied by development of stilpnomelane.

Unit 5.—This 760-ft unit is composed entirely of resistant, cliff-forming quartzite. The quartzite is thickly to very thickly bedded, reddish brown, and medium grained to conglomeratic. Well-rounded, white quartz pebbles up to 3 cm in diameter are sparse in the lower part of the unit, but are conspicuous in thin layers near the top of the formation (Fig. 6).

In thin sections the rocks are seen to be composed almost entirely of quartz with minor chert, sericite, and hematite. Traces of zircon and tourmaline are present. Quartz outgrowth has resulted in granoblastic fabrics; however, the original grain boundaries are seen as thin hematite coatings. Contiguous quartz clasts are incipiently sutured.

Unit 6.—Approximately 150 ft of maroon slate, argillite, and thick-bedded, coarse-grained to gritty quartzite makes up this bench-forming unit.

FIG. 6.—Conglomeratic quartzite, upper part of Unit 5, Precambrian of Beaver Mountains, Utah.

These rocks are intercalated in nearly equal amounts.

The slate and argillite consist mainly of sericite and chlorite with abundant quartz silt and very fine-grained sand. Finely disseminated hematite is present in amounts of up to 15 percent. Minor clastic muscovite and feldspar also are present. The slate is weakly schistose parallel with bedding and the argillite is nonschistose. The quartzite is composed almost entirely of quartz with minor interstitial white mica and traces of hematite, chlorite, zircon, and tourmaline. Outgrowth of quartz grains is common although contiguous clast may be incipiently sutured.

Unit 7.—This purple conglomeratic quartzite conformably overlies and forms a cliff above Unit 6. Unit 7 is mostly in fault contact with the stratigraphically overlying Prospect Mountain Quartzite. However, the sedimentary contact is sharp and well exposed near the north end of the map area (Fig. 2) where pink and tan Prospect Mountain Quartzite lies on the purple quartzite. A composite section of Unit 7 is 1,175 ft thick.

The quartzite is medium to very coarse grained and contains abundant gritty and conglomeratic zones. The conglomerate zones are thin and contain pebbles up to 1 cm in diameter of white or clear quartz and minor jasper. Beds are thin near the base of the unit but become thicker upward. There are a few silver and purple laminae of slate and argillite.

Quartz is the predominant mineral and minor clastic white mica and interstitial chlorite and sericite are present. There are traces of feldspar, chert, zircon, and tourmaline. Hematite makes up

FIG. 4.—Small-scale disharmonic folds within Unit 2, upper Precambrian of Beaver Mountains, Utah. Thick bed of arenaceous marble in upper right and banded slate and calcareous sandstone laminae below. Scale shown by marking pen.

grained, angular quartz in thin lenses probably represents chert lenses that were sheared and recrystallized during regional metamorphism.

Above the cliff-making marble is a bench formed on the upper 800 ft of Unit 2. This sequence consists chiefly of thinly interbedded slate, argillite, metasiltstone, quartzite, and marble.

The slate is weakly schistose parallel with bedding and is deep red because of abundant hematite. Argillite and metasiltstone are not schistose and are mostly tan, olive green, and greenish gray. Most of the marble contains quartz and locally grades into calcareous sandstone. Tan to olive-green quartzite predominates and ranges from very fine to coarse grained.

The upper contact of Unit 2 generally is covered with quartzite float from the overlying unit. The highest occurrence of greenish argillite float characteristic of Unit 2 coincides with the base of the cliff formed by Unit 3; this topographic break is mapped as the contact.

Petrographic study of the rocks of Unit 2 shows that the quartzite is composed principally of quartz clasts with minor interstitial sericite and chlorite or rarely calcite. Clasts of sodic plagioclase, potassium feldspar, hematite, chert, muscovite, biotite, zircon, and tourmaline are present in trace amounts. Some fine-grained clots of intergrown sericite and chlorite appear to be recrystallized shale or clay fragments. Stilpnomelane is rare and is related to late introduction of iron along fractures.

The marbles are calcitic and typically contain minor quartz. Siderite and dolomite rhombs are common in some specimens. Calcite lamellae are curved and crystals are elongate parallel with shearing in the more deformed marble. Evenly distributed, well-rounded, coarse quartz grains are probably of clastic origin. Other specimens contain small lenses of angular quartz, suggesting recrystallized chert lenses.

Unit 3.—This cliff-making unit is made up of 260 ft of quartzite with minor slate and metasiltstone near the top. The lower part of Unit 3 consists entirely of thick-bedded, medium- to coarse-grained, whitish and light-gray quartzite.

Approximately 50 ft below the top of the unit is an abrupt upward change from light-gray quartzite to dark-brown quartzite containing quartz granules and pebbles up to 2 cm in diameter. These larger clasts are subrounded and are composed of white quartz with minor jasper.

There is a 15-ft bench formed on olive-green slate and metasiltstone near the top of the unit. The slate is overlain by 10 ft of very thick-bedded, medium- to coarse-grained, dark-gray quartzite. The top of this resistant quartzite is assigned as the upper contact of Unit 3.

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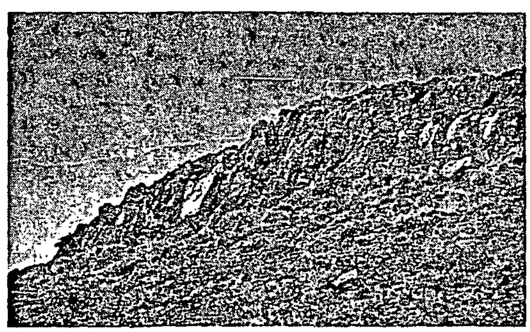

FIG. 5.—Marker bed of cliff-making marble 25 ft thick in Unit 2, upper Precambrian of Beaver Mountains, Utah.

1-3 percent of most specimens and is present as thin films on quartz clasts and as interstitial filling. The dark-purple color is apparently the result of the abundant but finely disseminated hematite. Quartz outgrowth is common and minor incipient suturing is also present.

The slate contains abundant white mica clasts parallel with bedding in a sericite and chlorite matrix. Quartz silt is a minor component. The argillite is similar in composition but is nonschistose. Hematite is present in amounts of up to 10 percent.

Unit 7 tentatively is correlated lithologically with the Mutual Formation of north-central Utah on the basis of its stratigraphic position, general lithology, and unusual color. At the type area in the Wasatch Mountains the Mutual Formation is overlain with angular unconformity by the Cambrian Tintic Quartzite (Crittenden *et al.*, 1952). In the Beaver Mountains the purple quartzite is concordant with the overlying Prospect Mountain Quartzite, but the contact may be a disconformity. This problem is discussed in the section on correlation.

Prospect Mountain Quartzite.—The sharp basal contact between the Prospect Mountain and the underlying Unit 7 has been mentioned. It is likely that this contact approximates the Cambrian-Precambrian boundary, although conclusive evidence is not available; a more detailed consideration of this problem follows in the discussion of correlation.

The Prospect Mountain Quartzite is thickly to very thickly bedded, mostly coarse grained to conglomeratic, and is predominantly pinkish tan. There are minor local zones of red, purple, and light-gray colors.

The lowest beds are poorly sorted and gritty, becoming conglomeratic about 20 ft above the base of the formation. This conglomerate zone is 20 ft thick and contains pink and tan, subrounded quartz pebbles up to 2 cm across. Upsection the rocks become pebbly quartzite. Locally the pebbles are sufficient to form thin conglomerate beds. Approximately 200 ft above the base of the formation the pebbles become less abundant and the rock is more uniformly coarse grained or gritty.

Pink feldspar clasts, mostly of microcline, up to 8 mm in diameter, are common in the lower part of the unit; some specimens contain as much as 10 percent feldspar.

Quartz outgrowth and minor suturing are common, although some rocks contain abundant interstitial sericite and chlorite which have prevented suturing of quartz clasts.

There are high-angle strike faults in the Prospect Mountain Quartzite and, because of the absence of distinctive marker beds within the formation, the stratigraphic separation on these faults is not known. Therefore, the true stratigraphic thickness cannot be determined precisely. It appears, however, that the thickness is at least 4,000 ft.

The contact with the overlying Pioche Shale is shown on the geologic map of southwestern Utah (Hintze, 1963).

There are no reported fossils in the Prospect Mountain Quartzite, but its conformable relation with the fossiliferous overlying Lower and lower Middle Cambrian Pioche Shale led Wheeler (1943) to assign the Prospect Mountain to the Lower Cambrian.

Correlation

The Beaver Mountains section is critical in Precambrian correlations in the eastern Great Basin because it is the principal Precambrian outcrop between the late Precambrian McCoy Creek Group (Misch and Hazzard, 1962) of eastern Nevada (Fig. 1, locs. 2, 3) and the Precambrian rocks described by Christiansen (1952) in the Canyon Range of central Utah (loc. 7).

Lithologic correlations are suggested for units in the lower part of the Beaver Mountains section and the lower part of the McCoy Creek Group (Fig. 7); however, there are considerable differences between the upper parts of these two sections.

Metamorphism of rocks of the McCoy Creek Group in their type section (Fig. 1, loc. 3) ranges from chlorite zone in the upper part of the group to biotite and garnet zone in the lower units. Thus, the type McCoy Creek Group is of a higher metamorphic grade than the rocks of the Beaver Mountains. If the effects of metamorphism are subtracted, the original sedimentary characteristics of the rocks can be compared.

Unit 1 of the Beaver Mountains is lithologically similar to the lower part of Misch and Hazzard's (1962) Unit A_2 which consists predominantly of light-colored quartzite. Marble and quartzite in the upper part of Unit A_2 are corre-

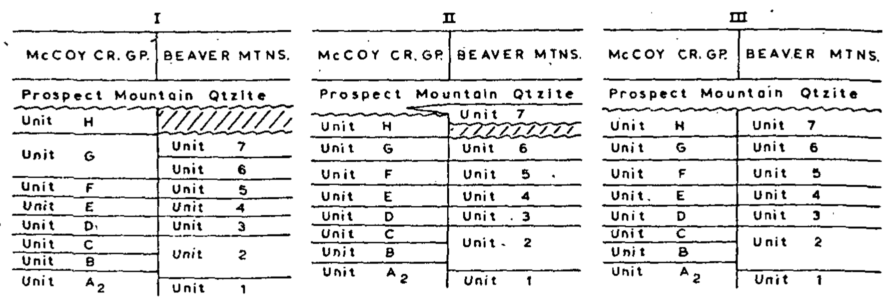

FIG. 7.—Possible correlations of rocks of Beaver Mountains, Utah, with McCoy Creek Group (Misch and Hazzard, 1962) of Schell Creek Range, Nevada (Fig. 1, loc. 3). Alternative I is favored.

lated with similar rocks in the lower part of Unit 2 (Beaver Mountains).

Subordinate carbonate interbeds in pelitic rocks in the lower part of Unit 2 and in Unit B (McCoy Creek) and intercalated quartzite and slate in the upper part of each unit suggest the equivalence of these formations.

Light-gray to whitish quartzite becoming pebbly upward and having green slaty interbeds near the top characterizes Unit 3 and Unit D (McCoy Creek).

Unit 4 and Unit E both are composed of greenish slate, argillite, and minor quartzite. Unit 5 and Unit F are composed entirely of thick-bedded quartzite with conspicuous conglomeratic zones near the top; Unit F is mostly gray with some purple zones, whereas Unit 5 is reddish brown.

The strata above Units 5 and F and below the Prospect Mountain Quartzite are markedly different in the two sections. The McCoy Creek rocks consist of gray, purple, and green argillite and slate (Unit G) overlain by coarse-grained white quartzite with minor, fine-grained, green quartzite interbeds (Unit H); in the Beaver Mountains this interval consists of maroon slate and quartzite (Unit 6) overlain by purple conglomeratic quartzite (Unit 7).

Under alternative I (Fig. 7), believed to be the correct correlation, Unit G represents the westward continuation of Units 6 and 7, and Unit 7 is a coarser and thicker lateral equivalent of the upper part of Unit G. Misch (personal commun., 1967) notes that the lithologic description of Unit 6 is very similar to that of parts of the Osceola Argillite of the Snake Range (Fig. 1, loc. 2), the eastward partial equivalent of Unit G of the Schell Creek Range (loc. 3). A further implication of this correlation is that a major disconformity is present at the top of Unit 7 in the Beaver Mountains. It seems probable, however, that the absence in the Beaver Mountains of strata equivalent to Unit H is caused by nondeposition rather than by erosion.

Alternative II does not seem likely; however, this possibility has been suggested in part in a cross section by Cohenour (1959a, p. 37). Unit 7, correlated with the Mutual Formation of north-central Utah, may intertongue westward with the Prospect Mountain Quartzite. The presence of purple zones in the lower part of the Prospect Mountain in eastern Nevada and western Utah may provide some support to this argument, but purple zones also are found in the Prospect Mountain Quartzite of the Beaver Mountains. Thus, there is no positive evidence in favor of alternative II. Furthermore, if this correlation were accepted, it would follow that there is probably a disconformity beneath Unit 7 in the Beaver Mountains.

Correlation of Unit H with Unit 7 (alternative III) seems improbable, although it is possible that the lithologic differences represent facies changes. However, these two units do not appear to be laterally equivalent for two reasons. First, the Mutual Formation (=Unit 7) grades from arkose in the Uinta Mountains into nonfeldspathic quartzite in the Beaver Mountains; Unit H on the west contains up to 4 percent detrital feldspar. Thus, a marked reversal in mineralogic trend toward the west must be accounted for if the units are considered correlative. Second, abundant hematite in Unit 7 (= Mutual Formation), at least in part detrital, is absent in Unit H. In view of the widespread and distinguishing

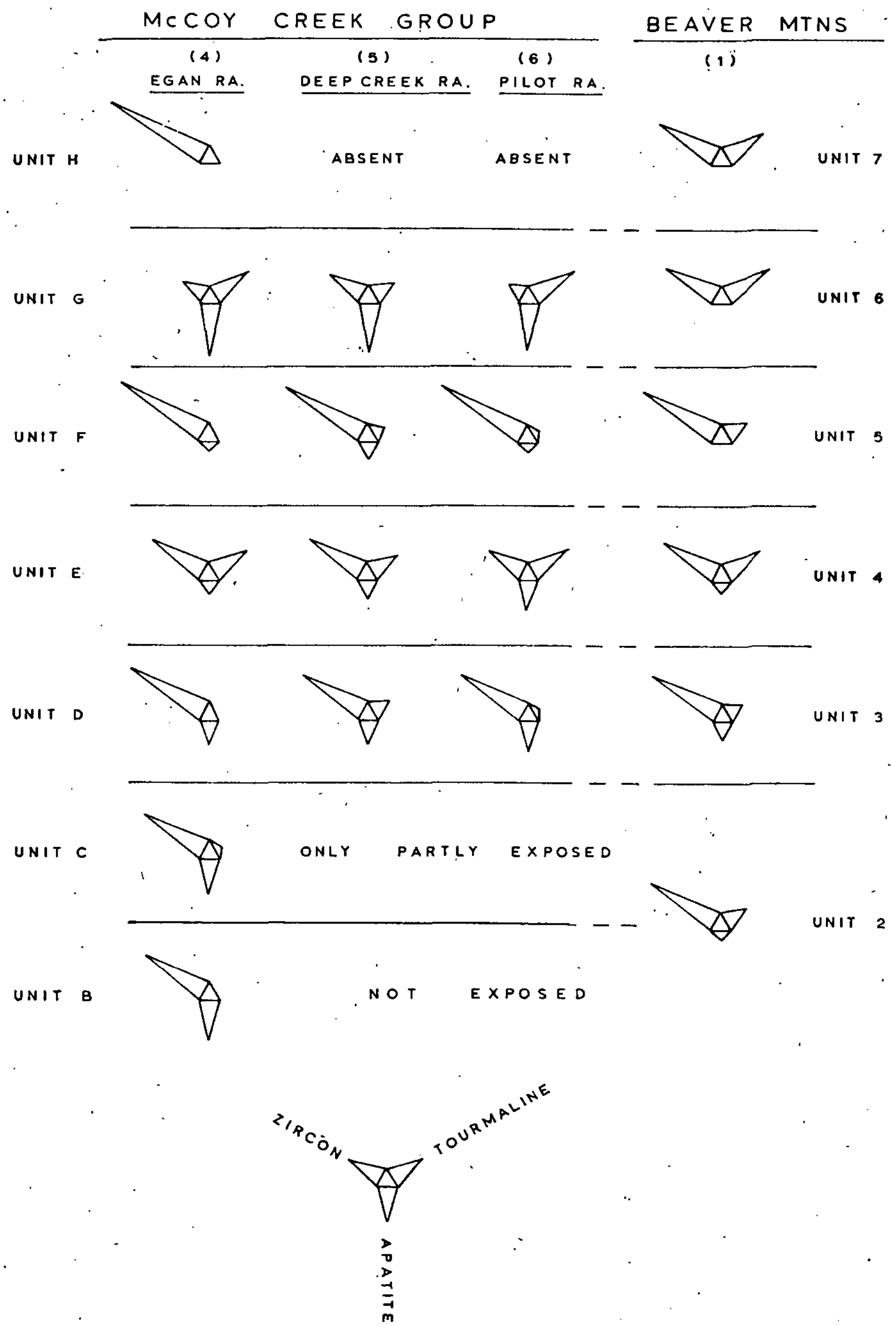

FIG. 8.—Comparison of nonopaque heavy-mineral assemblages of upper Precambrian rocks from Beaver, Pilot, northern Deep Creek, and Egan Ranges. Approximately 250 thin sections were examined. Locations are shown in Figure 1. Total minerals counted for each unit = 100 percent.

presence of hematite in the Mutual Formation, such a drastic westward change seems unlikely.

Nonopaque heavy minerals in thin sections of Beaver Mountains specimens were counted and compared with the heavy-mineral assemblages of the McCoy Creek Group (Woodward, 1967). It is seen that there are similarities in the assemblages in the lower units, but that the upper units contrast markedly (Fig. 8).

The Mutual Formation, the lithologic correlative of Unit 7, has been reported from the Sheeprock Mountains (Fig. 1, loc. 8) by Cohenour (1959b). The other Precambrian rocks described by Cohenour cannot be related firmly to the Beaver Mountains section even though they occupy the same general stratigraphic position below the Mutual Formation.

In the East Tintic Mountains (Fig. 1, loc. 9) the Cambrian Tintic Quartzite lies on strata that are correlated with the Precambrian Big Cottonwood Series of the Wasatch Mountains (Morris and Lovering, 1961).

Detailed petrographic and stratigraphic data concerning the Precambrian rocks of the Sheeprock and East Tintic Mountains are needed before their relation to the Beaver Mountains section can be determined.

The Precambrian strata of the Beaver Mountains and Canyon Range (Christiansen, 1952) correlate well lithologically (Fig. 9), although there are minor differences.

Unit 1 in both sections consists of light-gray, medium-grained quartzite. Arenaceous carbonate, quartzite, and pelitic beds in the lower part of

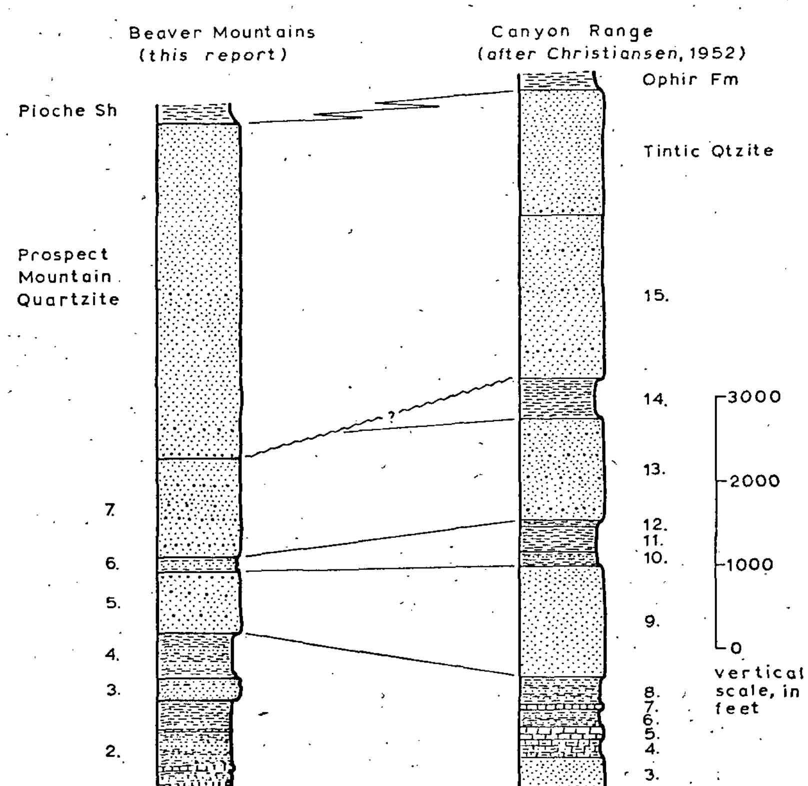

FIG. 9.—Tentative correlation of Precambrian and Lower Cambrian strata of Beaver Mountains and Canyon Range. Numbers beside columnar sections refer to informal unit designations used in this report and by Christiansen (1952). Locations are shown in Figure 1.

each section cannot be correlated bed for bed, but the overall sequence and the lithologic characteristics are remarkably similar.

Unit 5 (Beaver Mountains) and unit 9 of the Canyon Range are both made up of brownish, medium- to coarse-grained quartzite.

Unit 7 of the Beaver Mountains and Christiansen's unit 13 are both purple conglomeratic quartzite and are correlated with the Mutual Formation described by Crittenden *et al.* (1952) in the Wasatch Mountains (Fig. 1, loc. 10). There are facies changes between the Mutual Formation sections in the Beaver, Canyon, Wasatch, and Uinta Mountains in that the eastern sections contain more feldspar (Condie, 1966), becoming arkosic in the Uinta Mountains (Williams, 1953). Examination of thin sections of specimens from these other areas shows that the feldspars increase from traces in the Beaver Mountains to 1.5 percent (by volume) in the Canyon Range to 5.5 percent in the Wasatch Mountains to 38.5 percent in the western Uinta Mountains. Microcline is the dominant feldspar; sodic plagioclase is less abundant and tends to be more altered to clay and sericite than is the potassium feldspar. These facies changes are consistent with a northeast source of the sediments as described by Hansen (1955).

Unit 14 of the Canyon Range, consisting of maroon to tan, micaceous shale, is not present in the Beaver Mountains (Fig. 9). This shale occupies the same stratigraphic interval as the Red Pine Shale (uppermost Precambrian) in the western Uinta Mountains. There the Red Pine is underlain by the Mutual Formation and is overlain by the Tintic Quartzite with angular unconformity (Williams. 1953).

Thus, the lithologic correlative of the Red Pine Shale may be present in the Canyon Range. If this correlation is correct, it is likely that there is an unconformity at the base of the overlying quartzite (Christiansen's unit 15). In support of this hypothesis, it is pointed out that Christiansen (1952, p. 720) noted that unit 14 changes thickness considerably.

Christiansen (1952) has placed the base of the Tintic Quartzite and the base of the Cambrian System at the top of his unit 15, a light-gray to tan, medium-grained to conglomeratic quartzite. It is suggested that this unit is considered better to be the lower part of the Tintic Quartzite because its lithology is similar to that of the Tintic in adjacent areas (Cohenour, 1959b; Morris and Lovering, 1961). Also, the total thickness of Christiansen's unit 15 and his Tintic Quartzite, 3,450 ft, is closer to the 2,300–3,200 ft described by Morris and Lovering (1961) in the East Tintic Mountains (Fig. 1, loc. 9). Furthermore, Morris and Lovering (1961, p. 16) suggested that the Tintic thickens toward the south (*i.e.*, toward the Canyon Range).

The Tintic Quartzite of central Utah is the formational equivalent and partly the time equivalent of the Prospect Mountain Quartzite of eastern Nevada and western Utah (Wheeler, 1943, p. 1808–1811).

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LEASEHOLD EVALUATION

Exploration History and Results

Exploration began with the accidental drilling of the 270-foot-deep hole on the opalite mound in 1968. That event yielded an eruption of 270°F water, and began a leasing rush that is still in progress; this in turn led to the discovery of the Roosevelt geothermal field.

Figure 7 shows the location of deep holes (greater than 400 meters) in the Roosevelt area, and table 2 gives their characteristics. These holes were drilled either by Phillips Petroleum Company or by Thermal Power Company (a subsidiary of the Natomas Company). In addition to this deep drilling, which has reached approximately 7,500 feet in maximum depth, there have been dozens of gradient holes drilled in the broad area from north of Cove Fort to south of Thermo. This has been accompanied by tens of line miles of resistivity and magnetotelluric soundings, several months of microseismic recording, and correspondingly great efforts in hydrogeochemistry, seismic profiling, passive noise recording, gravimetry, aeromagnetometry, photogeology and field geologic mapping, and other exploration techniques.

This work has been accomplished by private companies, members of the U. S. Geological Survey and Utah Geological and Mineralogical Survey, faculty and student researchers at the University of Utah, and private contractors. Relatively little data are in the public domain, although proprietary information does circulate informally by trade or privileged release. These efforts in the Roosevelt Hot Springs area rival those accomplished at The Geysers or in Imperial Valley, California. Phillips Petroleum Company alone acknowledges spending $6 million on leasing, exploration and drilling since beginning work in 1973; they estimate a total outlay of $30,000,000 by 1982, to develop 110 mW of electric power generation (Gary Crosby, oral presentation, September 1976).

Drilling results reveal a hot-water reservoir, with temperatures as great as 260°C (500°F) at depths as shallow as 2,800 feet. Production comes from highly sheared and fractured Precambrian(?) gneiss and schist, and possibly from fractured

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Tertiary granite. The extent of Precambrian(?) rocks within the range is unknown: it may have been terminated beneath the Mineral Mountains by the Pliocene granitic intrusion or by the younger generation of rhyolite magma represented by the rhyolite domes and flows. Precambrian(?) rocks do crop out farther to the south along the range front, extending at least to the vicinity of Cave Canyon in T. 29 S.

Figure 2 reveals the conceptualization by scientists at Phillips Petroleum Company that pervasively fractured rock, rather than individual fault planes, forms the reservoir. Those holes not cutting the "shear zone" of figure 2 reportedly exhibit low permeability. However, all youthful faults may serve to carry hot fluids toward the surface, thus generating high temperature gradients in shallow holes.

The fault trending east-west across Negro Mag Wash apparently does not terminate the field: high temperature (∿440°F) was reported from hole 12-35, nearly 1.5 miles north of this south-dipping feature (G. Crosby, oral presentation, September 1976). Similarly, the southern boundary of the field is unknown.

Wells drilled very close to the range front commonly exhibit low gradients in their upper positions, reflecting cold-water recharge through fractured rocks of the mountains. Those drilled west of Dome fault exhibit high gradients in their upper portions, because of outflow of thermal fluid up Dome fault or other faults shown in figures 2 and 7. However, deeper gradients may be disappointing, and permeability often is limited.

It is interesting to note that the high gradients (as shown in plate 5) are displaced slightly westward from the proven geothermal field (figure 7). This recharge-discharge effect (described above) imposes a constraint on the use of gradient holes in the siting deep exploratory wells at or beyond the field boundaries. Structural and petrologic geology, results from earlier holes, and gravimetry all should be considered carefully, along with data on seismicity or the electromagnetic field, as these become available in the future.

Reservoir fluid is water-dominated and chloride-rich (∿3,000 mg/ℓ), with TDS of 6,000 to just over 8,000 mg/ℓ, average enthalpy of 500 BTU/pound, and temperature at one kilometer of

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160°C (320°F) to 260°C (500°F). Mass flow in productive wells averages about one million pounds per hour of hot water, at pressures slightly above hydrostatic. Individual well tests up to 1.5 million pounds per hour have been reported, but stabilized flow may be expected to be less than this. An average steam flash of 18 to 20 percent is anticipated, although pressure, mass flow, and therefore percentage of steam flash and steam enthalpy, vary somewhat from hole to hole.

Using "average" conditions, a productive well may be expected initially to yield 200,000 pounds of steam per hour, after flash separation. This should serve to generate 10,000 kW per hour, or 10 mW. This is quite high, and therefore attractive commercially. Ultimately, stabilized flow may be somewhat lower.

It is probable that more than one thermal zone is present, one being a shallow system fed by leakage up faults, and a second being a deeper system of pervasively fractured rock. Differences may be expected in pressure, temperature, enthalpy, and possibly mass flow and well life in these two systems. Because field dimensions are unknown, and because no data are available on long-term well productivity, ultimate generating capacity is unknown. Well spacing probably will be determined empirically, using results of interference tests between existing wells to determine optimum locations.

Reinjection will be required for all waste fluids, and probably for any condensed steam that is not evaporated in the cooling process. Such wells may be located to the northwest and west of the producing field, and a ratio of one injector for two producers is anticipated (G. Crosby, oral presentation, September 1976).

Phillips has applied to drill 21 additional wells to define field boundaries and parameters. Still additional wells will be required for development purposes.

Competitive Interest

A Known Geothermal Resources Area (KGRA) was established at Roosevelt Hot Springs prior to the beginning of leasing in the public domain in January 1974. Therefore, no applications for non-competitive leases were accepted for a zone of 23,000

Table 2. Geothermal Wells Drilled at Roosevelt Hot Springs, Utah

Well Number	Location	Date Drilled	Depth, ft.	Casing	Results and Status
					Phillips Petroleum Company
OH-2 =13-10	SW/4 NW/4, 10-27S-9W	2/2/75-2/15/75	2,250	N.D.	Deep temperature-gradient hole; reportedly high gradient
OH-1 (also 17-1)	SE/4 NE/4, 17-27S-9W	3/3/75-3/12/75	2,321	N.D.	Deep temperature-gradient hole; "high" temperature; low permeability
9-1	NE/4 NW/4, 9-27S-9W	3/13/75-4/8/75	6,885 2099 m	N.D.	"High" temperature; poor permeability
3-1 (also 55-3)	NW/4 SE/4, 3-27S-9W	4/20/75-5/24/75	2,728 831 m	N.D.	Tested at 1.2 million #/hr of hot water
54-3	SW/4 NE/4, 3-27S-9W	7/5/75-8/28/75	2,882 878 m	N.D.	~jmillion #/hr of hot water at >500°F and >500 BTU/#; rated as "best" well
12-35	NW/4 NW/4, 35-26S-9W	8/6/75-10/1/75	7,324 2232 m	7" liner to 4,500'	Suspect shallow-zone cool-water contamination; ~440°F thermal aquifer now lined off; cannot test satisfactorily
13-10	SW/4 NW/4, 10-27S-9W	10/2/75-11/4/74	5,351 1631 m	N.D.	Tested above 1 million #/hr of hot water at 75-125 psig
82-33	NE/4 NE/4, 33-26S-9W	11/5/75-12/23/75	6,028 1837 m	13-3/8" to 575'	>300°F, <350°F; possible future injection well
25-15	NW/4 SW/4, 15-27S-9W	8/26/76-11/12/76	~7,500 2286 m	9-5/8" at ~2,500'	Shallow-zone cool-water contamination; less satisfactory than wells to north
(Note: All Phillips' wells are on Federal lease blocks)					
					Thermal Power Company (Natomas Company)
Utah State 14-2 (ML27536)	SW/4 NW/4, 2-27S-9W	9/11/76-10/21/76	6,108 1862 m	9-5/8" at 1,805'	Reported >400°F hot water at ~4,000'
Utah State 72-16 (ML25128)	NW/4 NE/4, 16-27S-9W	10/22/76-1/5/77	1,254 382 m	N.D.	Reported 1 million #/hr of hot water at 432°F and 355 psig

UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY
Conservation Division, MS-92
345 Middlefield Road
Menlo Park, CA 94025

MAR 30 1981

→ Dennis

MAR 27 1981

Memorandum

To: Interested Parties

From: Acting Deputy Conservation Manager, Geothermal

Subject: Unit Plan of Operation for Development, Phillips Petroleum Company, Roosevelt Hot Springs Unit, Federal Lease U-27386, Beaver County, Utah
Ref: 2404-02 RHSU (POO for CER #177-81)

In August 1980, Phillips Petroleum Company submitted Plans of Operation for Development, Injection, Utilization and Production for a 20 MW (net) power plant within the Roosevelt Hot Springs Unit. The Office of the Deputy Conservation Manager for Geothermal (DCM, Geothermal) is currently preparing an Environmental Assessment (EA) for the Plans of Injection, Utilization and Production. As stated in our Interested Parties letter of March 18, 1981, operations proposed under the Plan of Development are to be treated by a separate environmental review in order to expedite the drilling program. Under this Plan, Phillips is proposing the drilling of two wells and has selected three possible sites. Enclosed is an Addendum to the previously submitted Plan of Development, with a map, indicating the locations currently under consideration.

A Categorical Exclusion Review (CER #177-81) will be prepared by the Office of the DCM, Geothermal for the proposed site preparation, drilling operations and pipelines.

Because numerous field inspections have been conducted in the immediate vicinity of the proposed operations, a field inspection is not considered necessary at this time. However, you are invited to visit the sites at your own convenience. Any inquiries should be directed to Rich Hoops or Buford Holt of this office (Tel: [415] 323-8111, ext. 2848; FTS: 467-2948).

We urge you to send written commentary and will appreciate hearing from you even if you are of the opinion that the existing regulations, lease terms, and operational orders provide adequate environmental protection.

All comments concerning the proposed actions should be received no later than April 17, 1981 by:

DCM, Geothermal
U.S. Geological Survey
Conservation Division
345 Middlefield Road, MS-92
Menlo Park, CA 94025
Attn: Richard Hoops
(415) 323-8111, ext. 2848
FTS: 467-2848

All comments will be given serious consideration in the preparation of the Categorical Exclusion Review and any subsequent condition of approval.

The Deputy Conservation Manager for Geothermal will not send a draft CER to interested parties for review. Certain parties, however, such as the surface managing agency, the lessee, GEAP, and the U.S. Fish and Wildlife Service will receive a copy of the completed CER.

Copies of the CER and the complete Plan of Operation are available for inspection during normal business hours at the Office of the Deputy Conservation Manager for Geothermal, the Office of the District Geothermal Supervisor in Salt Lake City, and at the Cedar City District BLM Office. Copies are available upon request.

William F. Isherwood

William F. Isherwood

Enclosure

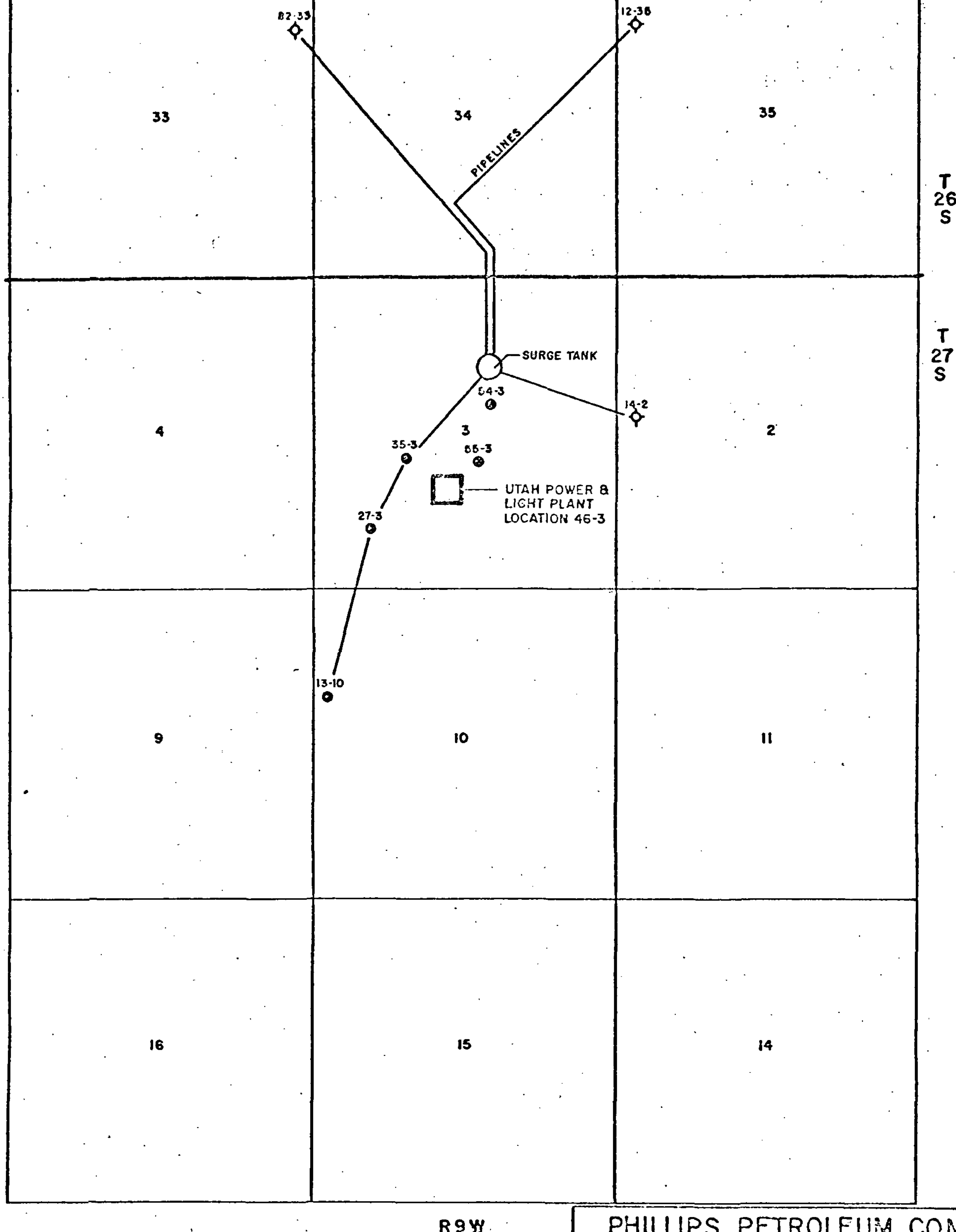
ADDENDUM NO. 1 TO PLAN OF OPERATION, UTILIZATION

This represents an addition to Section II, B. of Utah Power and Light's Plan of Operation, Utilization submitted to U.S.G.S. in January, 1981.

A proposed alternate well will be located in Section 3, T. 27 S., R. 9 W. in the Roosevelt Hot Springs geothermal area approximately 580' south of existing well 54-3 and 650' east of proposed well 35-3. The proposed location is shown as well 55-3 in the attached figure.

This well is an alternate producing well which will be drilled only if either of the proposed producing wells 35-3 or 27-3 are determined to be non-productive or unfit for production. The alternate well will be drilled as an offset to abandoned well 3-1 in a known producing zone of the field. This will provide a relatively low risk option for encountering production. The drilling and operation of this well will follow the guidelines presented in the Plan of Operations, Development.

The production and injection pipeline design to include the alternate well will depend largely on which of the proposed wells is found non-productive. In any case, the present pipeline design will be modified to optimize operational flexibility and cost. The attached figure shows the brine injection system as it is presently designed.



PRODUCTION WELL

INJECTION WELL

PHILLIPS PETROLEUM COMPANY
GEOTHERMAL OPERATIONS
431 SOUTH 300 EAST SALT LAKE CITY, UTAH 8411

ROOSEVELT HOT SPRINGS
WELL LOCATION MAP
BEAVER COUNTY, UTAH

0 1/2 1 2 MI.
SCALE

GEOLOGIST	JOHN BAZA	DATE	JAN. 22, 198
DRAFTSMAN	DOYLE OLSON	DATE	JAN. 23, 198
REVISED	DOYLE OLSON	DATE	MAR. 3, 198

AREA
UT
Beaver
Roosevelt
Chem

Table G-1. CHEMISTRY OF THERMAL WATERS, ROOSEVELT HOT SPRINGS KGRA UTAH: A SUMMARY[a]

Sample No.		1	2	3	4	5	6	7	8	9
Well: Utah State[b]		14-2[f,g]	14-2[f,h]	14-2	14-2	54-3	54-3	72-16	72-16	72-16[i]
Source/Analyst		ESL	ESL	USGS	USGS	Woodward-Clyde Consultants	Phillips Petroleum Company	USGS	USGS	Univ. of Utah
Reference		This Report	This Report	ESL, 1978a	ESL, 1978a	White, 1978	Lenzer et al., 1976	ESL, 1978a	ESL, 1978a	Ward et al., 1978
Collected		5/78	?/78	11/77	11/77	11/75	8/75			1/77
Analyzed		8/79	8/79	3/77	3/77			10/77	10/77	
Na	ppm	2070	2340	2150	2200	2000	2000	1800	2000	2072
K	ppm	384	419	390	410	400	410	380	400	403
Ca	ppm	11	6.8	9.2	6.9	7.0	10.1	12.4	12.20	31
Mg	ppm	0.28	<0.24	0.6	0.08	0.1	0.24	0.29	0.29	0.26
Fe	ppm	0.13	<0.02			0.2				0.016
Al	ppm	0.31	<0.28			0.5				
Si	ppm	226	31	299	383	140	>262	238	244	299
Sr	ppm	1.44	1.28			<5		1.36	1.20	
Ba	ppm	0.24	<0.24			<0.4				
Mn	ppm	<0.20	<0.20			<0.02				
Cu	ppm	<0.20	<0.20							
Pb	ppm	<0.20	<0.20			0.18				
Zn	ppm	<0.20	<0.20							
As	ppm	3.2	3.6	3.0	2.2	3.8				
Li	ppm	25	28			20	19.0	15.0	16.0	
Be	ppm	0.004	<0.004							
B	ppm	23	25	29	28	28	29	26.4	27.2	
Ce	ppm	<0.20	<0.20			3.1				
Br	ppm					<5				
I	ppm									
Rb	ppm					3.9				
F	ppm			5.2	4.8	6.0	5.0	5.2	5.3	
Cl	ppm			3650	3650	3600	3400	2110	3260	3532
HCO_3	ppm					200	200	181	181	25
SO_4	ppm			78	60	55	54	33	32	48
NO_3	ppm					<0.05	tr.			
NH_4	ppm					<1				
PO_4	ppm	<2	<2			<1				
T.D.S[c]	ppm					6700	6442	6074	6444	6752
pH				5.9	6.2	6.7 to 7	6.5	7.83	7.53	5.0
T	°C			14	9		>260			92
$\delta^{18}O$	‰			-13.46	-13.27					
Geothermometer										
T(Na-K-Ca)[d]	°C	284	291	256	293	295	293	289	288	274
T (Na-K-Ca-Mg)[e]	°C	-	-	-	-	-	-	-	-	-

FOOTNOTES

a. A blank indicates data not determined or information not available.
b. For well locations see Figure 1, or footnotes k and l.
c. Total dissolved solids.
d. Calculated using the method of Fournier and Truesdell (1973, 1974). $\beta = 1/3$.
e. Calculated using the method of Fournier and Potter (1978).
f. Elements analyzed for but present at concentrations less than ICPQ limits of quantitative detection (Appendix C) include: Ti, V, Cr, Co, Ni, Mo, Cd, Ag, Au, Sb, Bi, U, Te, Sn, W, Zr, La, Th.
g. Sample supplied by J.R. Bowman, Univ. of Utah, Salt Lake City, Utah. Precipitate was noted in the original sample container.
h. Sample supplied by A.H. Truesdell, U.S.G.S., Menlo Park, CA. Precipitation in the original bottle is suspected.
i. Surface leakage.
j. Flow test at 7170 to 7500 feet.
k. Location: Sec. 34 dcb, T26S, R9W.
l. Location: Sec. 34 bdd, T26S, R9W (Salt Spring). Flowing at the time of collection (personal communication, W.T. Parry, Univ. of Utah, Salt Lake City, Utah, 1979).

Table G-1 (cont.). CHEMISTRY OF THERMAL WATERS, ROOSEVELT HOT SPRINGS KGRA UTAH: A SUMMARY[a]

Sample No.		10	11	12	13	14	15	16	17	18	19
Well: Utah State[b]		3-1	52-21[j]	52-21[j]	52-21[j]	9-1	Roos. Hot[k] Spring	Roos. Hot[k] Spring	Roos. Seep[l]	Roos. Seep[l]	Roos. Seep[l]
Source/Analyst		Phillips Petroleum Company	Getty Oil Company	Getty Oil Company	Getty Oil Company	Univ. of Utah	USGS	USGS	Univ. of Utah	Phillips Petroleum Company	Phillips Petroleum Company
Reference		Lenzer et al., 1976	ESL, 1978b	ESL, 1978b	ESL, 1978b	Parry et al., 1976	Mundorff, 1970	Mundorff, 1970	Ward et al., 1978	Lenzer et al., 1976	Lenzer et al., 1976
Collected		5/75	11/78	11/78	11/78		11/50	9/57	6/75	5/73	8/75
Analyzed			12/78	12/78	12/78						
Na	ppm	2437	1845	1900	1900	2210	2080	2500	1840	2400	1800
K	ppm	448	237	218	216	425	472	488	274	378	280
Ca	ppm	8.0	106	114	107	83	19	22	122	113	107
Mg	ppm	0.01	5.2	3.9	4.0		3.3	0	25	17	23.6
Fe	ppm		0.31	6.9	6.3						
Al	ppm		<0.1	<0.1	<0.1			0.04			
Si	ppm	262(?)	234	67	65	170	189	146	81	36	50
Sr	ppm										
Ba	ppm										
Mn	ppm										
Cu	ppm										
Pb	ppm										
Zn	ppm										
As	ppm										
Li	ppm	20.0						0.27			17
Be	ppm										
B	ppm	25	24.4	27.0	27.0			38		37	29
Ce	ppm										
Br	ppm										
I	ppm										
Rb	ppm										
F	ppm	5.0	2.8	3.4	3.6		7.1	7.5		5.2	3.3
Cl	ppm	4090	2810.8	2885.1	2881.6	3800	3810	4240	3210	3800	3200
HCO_3	ppm	180	602.9	550.0	615.0		158	156	298	536	300
SO_4	ppm	59	78	86	85	122	65	73	120	142	70
NO_3	ppm	0.1	0.5	1.3	1.3		1.9	11		tr.	tr.
NH_4	ppm										
PO_4	ppm										
T.D.S[c]	ppm	7067	5940	5727	5677		7040	7800	6063	7506	5948
pH		6.3	6.4	7.3	6.8			7.9	6.5	8.2	6.43
T	^oC	>205					85	55	25	17	28
$\delta^{18}O$	$^o/oo$										
Geothermometer T(Na-K-Ca)[d]	^oC	292	227	219	219	262	293	284	235	246	239
T (Na-K-Ca-Mg)[e]	^oC	-	209	210	209	-	283	-	139	181	141

FOOTNOTES

a. A blank indicates data not determined or information not available.
b. For well locations see Figure 1, or footnotes k and l.
c. Total dissolved solids.
d. Calculated using the method of Fournier and Truesdell (1973, 1974). $\beta = 1/3$.
e. Calculated using the method of Fournier and Potter (1978).
f. Elements analyzed for but present at concentrations less than ICPQ limits of quantitative detection (Appendix C) include: Ti, V, Cr, Co, Ni, Mo, Cd, Ag, Au, Sb, Bi, U, Te, Sn, W, Zr, La, Th.
g. Sample supplied by J.R. Bowman, Univ. of Utah, Salt Lake City, Utah. Precipitate was noted in the original sample container.
h. Sample supplied by A.H. Truesdell, U.S.G.S., Menlo Park, CA. Precipitation in the original bottle is suspected.
i. Surface leakage.
j. Flow test at 7170 to 7500 feet.
k. Location: Sec. 34 dcb, T26S, R9W.
l. Location: Sec. 34 bdd, T26S, R9W (Salt Spring). Flowing at the time of collection (personal communication, W.T. Parry, Univ. of Utah, Salt Lake City, Utah, 1979).