ASPECTS OF LOW TEMPERATURE GEOTHERMAL RESOURCE ASSESSMENT WITH EXAMPLES FROM KANSAS AND OREGON

David D. Blackwell

Department of Geological Sciences
Southern Methodist University
Dallas, Texas 75275

ABSTRACT

The geological factors which need to be taken into account in the assessment of low temperature geothermal resources are those which control subsurface temperature and fluid flow. The primary quantities of interest are surface temperature, heat flow (thermal conductivity and geothermal gradient), geology, and hydrology. Factors causing variation in heat flow (geothermal gradient) are discussed. Local heat flow is effected by mantle heat flow, the heat production of the crust, local anomalous heat sources, local vertical and lateral variations in thermal conductivity and in some cases by water flow. Because of the many, complex, and geographically variable factors which are involved in determining temperature at a specific point, a uniform country-wide approach to low temperature geothermal assessment is not possible. The difficiencies of existing analyses of only "geothermal gradient" as a basic approach and as illustrated by three different gradient maps are demonstrated. Examples of approaches to evaluation in two different settings (Kansas and Oregon) are illustrated. Kansas is in the geologically stable Midcontinent where horizontal sedimentary rocks overlie a granitic and metamorphic basement. Oregon is in the geologically complex and young Cordillera. The significant data in Knasas include relatively widely spaced (10-20 km), vertically detailed, temperature-depth and thermal conductivity measurements and aquifer analyses (location, water quality and flow characteristics). It is possible to interpolate temperature values at intermediate sites using well samples and logs to determine basement radioactivity and thermal conductivity and thus local heat flow and temperature. In Oregon, in the more complex geological provinces, assessment must proceed statistically or on a case by case basis. The important parameters are local geothermal gradients and heat flow, and aquifer conditions. Interpolated values between data points are not reliable because an individual data point has a lateral zone of significance of only about 1-5 km.

INTRODUCTION

The utilization of low temperature geothermal resources (below 150°C and especially below 100°C) is quite different than the utilization of higher temperature geothermal resources. Similarly the exploration techniques are quite different so a customized exploration approach is necessary in low temperature resource assessment. Description of such a customized approach is the object of this paper.

In exploring for high temperature geothermal systems for power generation the main objective is to locate anomalous areas which can then be evaluated. On the other hand, utilization of low temperature resources is possible in many areas of "normal" temperatures. Similarly, there is a major difference in the transportability of the energy. It is as economically effective to transport electrical power produced from geothermal energy for long distances. However, it is not economical to transmit low temperature geothermal fludis over the same long distances. In Iceland, pipelines carry geothermal waters over 20 kilometers to heat most of the town of Reykjavik. However, pipeline costs probably prohibit such transportation distances for any but the largest of geothermal space heating systems. Northwest Geothermal Corporation in Portland, Oregon has studied the concept of transporting low temperature resources a distance of perhaps 50 kilometers from the vicinity of Mt. Hood to the outskirts of Portland, Oregon (John Hook, personal communciation, 1978). However, only feasibility studies have been initiated. Therefore, in general, utilization of low temperature geothermal resources depends on knowledge of the local geothermal conditions.

Finally, the economics of low temperature geothermal development do not allow for extensive exploration. The drilling of a single production well may be the only exploration costs that can be justified. Therefore the main objective of a low temperature resoruce assessment program should be to identify and to evaluate low temperature geothermal resources near likely areas of utilization so the user can then take the information, drill his production well, and install the surface equipment necessary. Because of the nature of the resource, only the most cost effective exploration techniques can be utilized. These techniques will consist generally of geological studies, geochemical studies, scrounge (freehole) temperature gradient and heat flow studies, and hydrology studies. Heat flow and geothermal gradient drilling would be the most expensive techniques justified in the general case. The difficulty with other geophysical techniques is that they cannot directly identify the actual temperatures present and drilling will be required in any case to evaluate the resource. Techniques such as resistivity profiling, gravity studies, and seismic studies may be important for specific sites.

The emphasis in this paper is on geothermal gradient and heat flow studies but the interrelation of these techniques with other techniques will also be discussed. In the application of heat flow and geothermal gradient techniques to the assessment of low temperature geothermal systems there are two strategies; the first strategy is to make temperature measurements in existing wells in order to directly measure the main parameter of interest, temperature, at various depths; the second strategy is to drill holes specifically for heat flow-geothermal gradient studies in areas which appear geologically and hydrologically favorable and where holes do not otherwise exist or are too shallow. One advantage of the heat flow-geothermal gradient studies is that they are non specific, i.e. they furnish information for evaluation of the area both for high temperature and dry hot rock resources as well as low temperature resources. Therefore, the data collected for the purpose of evaluation of any of the three resources can apply just as well to the other two.

The approach in this paper is to describe the different types of evaluation techniques which can be applied in two completely different geological settings; one, the state of Kansas in the stable interior part of the North American continent; two, the state of Oregon in the geologically young Cordillera of the western United States In Kansas, Pre-Cambrian igneous and metamorphic basement rocks are overlain by near horizontal Paleozoic, Mesozoic and Cenozoic sedimentary rocks. The main geothermal resource is warm water in porous sedimentary layers above the basement. Lateral and vertical variations in the temperature in the aquifers will be small and a relatively wide spacing of holes will suffice for the geothermal evaluation. On the other hand, in Oregon, the geological setting is complex and there is a wide variety of heat flow settings as well. A much denser measurement spacing is necessary to evaluate such a complicated area where there are many different controls on the heat flow. As a preface to the discussions of these two areas, the various general controls on subsurface temperature will be discussed and previous attempts to evaluate the low temperature geothermal potential on a country-wide basis will be briefly discussed.

CONTROLS ON SUBSURFACE TEMPERATURES

The factors which control subsurface temperature include regional heat flow, geology, hydrology and surface temperature. The regional heat flow is the heat which comes from the mantle. Its ultimate origin is uncertain but it probably comes from either the internal heat of the earth left over from planetary formation, deeply buried radioactivity, or viscous friction as the plates slip over the interior of the earth. In the absence of disturbing factors such as hydrologic circulation, lateral variation in thermal properties and so forth, the surface heat flow is a function of this regional heat flow and the heat production of the basement rocks, i.e. the amount of heat generated from the decay of uranium, thorium, and potassium in the outermost 10-20 km of the crust. Studies of the correlation between heat flow and heat generation in plutons demonstrate a very systematic relationship (Birch *et al*., 1968; Roy *et al*., 1968; Lachenbruch, 1968). This correlation between heat flow and heat production can be used to infer properties of the crust and to identify the regional background heat flow for a given heat flow province (Roy *et al*., 1972).

In a horizontally layered sequence of sedimentary rock such as that overlying the basement of the Mid-continent region, the vertical heat flow above the basement will be constant. Since the heat flow is constant, the temperature gradient will be inversely proportional to the thermal conductivity. Therefore, the temperature at a given depth is a function of the mean thermal conductivity above that point, the surface temperature and the regional heat flow. So in order to evaluate the temperature in the acquifer at a given depth three things must be known: the surface temperatue, the surface heat flow, and the total thermal resistance (inverse of the mean thermal conductivity) between the surface and the depth to the aquifer. The thermal resistance of the sedimentary section can be determined from direct measurements, from knowledge of lithology or from indirect measurements such as well logs, seismic data, etc. These various techniques will be discussed in a following section.

Lateral variations in heat flow within a given heat flow province not caused by variations in basement radioactivity are due to variations in the geology which result in lateral variations of thermal conductivity (thermal refraction) and in hydrologic effects on heat flow data. The effects of refraction on heat flow have been discussed in the literature (Lee and Henyey, 1974, and others). These refraction effects will be common in areas of complex geology. In general, basement variations of thermal conductivity will have a minor effect of the temperatures in the overlying sedimentary rocks.

The final major control on heat flow and subsurface temperature is the hydrologic setting. Although a large amount of data have been collected in the past few years so that we can quantify the effects of hydrology on heat flow, there is still a great amount of confusion in the literature about the exact nature of these effects. Enough data has been collected to know when hydrology is likely to effect heat flow values and when it is not. The areas where heat flow values are effected by hydrology are more easily recognizable than the impression generally given in literature. In areas of high heat flow and extensive fracturing, geothermal systems obviously represent severe hydrologic distortions of the regional conductive heat flow pattern. Such a geothermal system close to a heavily populated area represents a very cost effective energy source. In some situations the circulation is so fast that the heat flow is essentially washed away. These effects are confined, however, to extremely porous and permeable shallow rocks, such as very young volcanic rocks, typical of the Snake Plain aquifer of eastern Idaho (Brott et al., 1981), the Bend area in central Oregon (Blackwell and Steele, 1979) and in very porous gravels and alluvial deposits. Aquifer locations and characteristics in the United States have been extensively studied and even in most major aquifers, flow of water is slow enough that the conductive heat flow is preserved. In the basalt aquifers of the Columbia Plateau regional heat flow is only slightly distorted by the water circulation, although large head differences exist in aquifers cut by a single well. Intrabore communication may cause extremely complicated temperature-depth curves but if the holes are grouted the observed temperature-depth curves are linear and indicate that the primary effect on subsurface temperature is conductive heat flow and not fluid motion.

In the Midcontinent aquifer flow may cause small variations in heat flow and temperature, but these will be recongizable on a regional scale and will not effect small areas (see Gosnold, 1981). An exception to this comment may be found along the Missouri River in South Dakota and Nebraska where the hot water occurs in shallow wells in the Dakota Sandstone. Adolphson and LeRoux (1968) suggested that the water is leaking up from a deeper aquifer (Madison Limestone) into the Dakota Sandstone. This is the only known such geothermal gradient and heat flow anomaly in the eastern United States, although others may well exist. Therefore in the eastern United States the effects of hydrology on heat flow in general are relatively minor except in a few isolated cases.

EXISTING HEAT FLOW AND GEOTHERMAL GRADIENT MAPS

There are two extensive evaluations of the geothermal potential of the United States published by the U.S. Geological Survey (White and Williams, 1975; Muffler, 1979). The emphasis in these reports is on evaluation of high temperature hydrothermal (>90°C) convective systems. Thus these studies have not attempted to evaluate "background" areas where much, if not most utilizable, low temperature potential exists.

Various heat flow maps have been published for the United States (Sass _et al._, 1971; Roy _et al._, 1972; Sass _et al._, 1981). Heat flow maps by the nature of their preparation take into account the thermal conductivity of rocks so that they show less scatter due to geologic variations than would a geothermal gradient map. However, for the low temperature geothermal utilization the important quality is the _local_ mean geothermal gradient and the temperature in aquifers where the hot fluid can be brought to the surface relatively easily. These data are not readily apparent from heat flow maps.

Partially in an intent to go back more closely to temperature several different geothermal gradient maps of the United States have been prepared. However, geothermal gradient maps will be more susceptable to variations in approach and ways of treating the data than will the heat flow maps and this variation shows up in the gradient maps. Three rather extensive data bases have been utilized to prepare maps. A map was prepared and published by A.A.P.G.-U.S.G.S. (1976)

based on bottom hole temperatures measured during drilling of hydrocarbon exploration wells. Kron and Heiken (1980) published a geothermal gradient map based on the extensive heat flow data base available in the literature and in preprint form. Guffanti and Nathenson (1980) published a map based on a data set obtained during the 1920's and 30's of gradients from hydrocarbon exploration wells, generally deeper than 500m, obtained from logging with maximum reading thermometers (Van Ostrand, 1951). The maps prepared using these three different data sets show very little similarity of geothermal gradient contours and if the contours were plotted without the map base, it would not be apparent that the same areas are in fact involved in the studies. Part of the reason for the difference is that three totally different data sets have been evaluated and presented independently. These data are interrelated but do not a priori have the same information content. Thus, there is a great variaion in the contours for the same areas between the different maps and the fact is that none of the three maps is of much use in the general evaluation of geothermal potential. This point will be illustrated in the discussion. A major reason for variations on the A.A.P.G.-U.S.G.S. map is that the map is based on the single-point bottom-hole temperature values and is afflicted with problems of temperature non-equilibrium and data uncertainty. Nevertheless, this map has been used extensively and its relationship to more detailed data has been discussed fairly completely (for example, see Gosnold, 1981). The contours from the Guffanti and Nathenson (1980) map and the Kron and Heiken map (1980) in Kansas are shown in Fig. 1 as is some deep well control which has become available since these maps were produced. The Guffanti and Nathenson (1980) map has an area of high gradient in the center part of the state with lower gradient to the east and to the west (based on one data point). The Kron and Heiken (1980) map has a localized area of high gradient near the extreme eastern edge of the state and very low gradient in the center part of the state. Not only are these maps very different, but it will be demonstrated in the next section that neither of these contourings is an accurate and useful depiction of the geothermal gradient in Kansas from the point of view of low temperature resource assessment.

In the state of Oregon, the Kron and Heiken (1980) map is based on an extensive set of heat flow data presented by Blackwell _et al_. (1978). Because the thermal conductivity of most of the surface rocks in Oregon does not vary greatly, the gradient map approximates the published heat flow map. The Guffanti and Nathenson (1980) map contours are based on four data points. Their map indicates that the area east of the Cascade Range has a gradient between 35 and 39°C/km whereas, in fact, except for the Blue Mountains and the Columbia Plateau regions, the whole area has a gradient in excess of 50°C/km (see below). The Guffanti and Nathenson (1980) map is of little use in the state of Oregon because of the small data base and because it has essentailly ignored a very extensive set of data which was available at the time the map was prepared.

Thus the existing gradient maps do not approach adequacy for the evaluation of low temperature geothermal potential. In part this is because of lack of data, but in part it is because of an incomplete approach to the problem. Since the nature of the data which must be used in the evaluation varies from place to place based on geologic setting, a single systematic countrywide approach is not possible. The object of the remaining sections of this paper is to discuss in more detail the way low temperature geothermal evaluation might proceed and a possible types and sources of data which should be used in low temperature geothermal evaluation.

KANSAS

Until quite recently, there was only one area in Kansas for which relatively accurate and detailed temperature measurements were available. This was near the Lyons area in central Kansas (Sass _et al_., 1971). Recently, a series of wells drilled as aquifer tests by the U.S. Geological Survey and the Kansas Geological Survey (Steeples and Bickford, 1981) allowed a much more detailed set of measurements in several holes. In addition, some abandoned or shut-in hydrocarbon wells were logged. The locations of these wells are shown in Fig. 1. Details of the drilling and the studies which have been carried out in the wells are given by Steeples and Bickford (1981). In this discussion the holes are identified by their township /range and section location and the township identifications are given in Fig. 1. The basic section encountered in the hole consists of Permian sands, shales and evaporite deposits. Below the Permian is a Pennsylvanian section which is predominately limestone and shale with minor amounts of sandstone and coal. The Paleozoic section below the Pennyslvanian consists primarily of carbonate rocks (limestone and dolomite) with one or two units of shale (the Chattanooga shale and the Silvan shale). The geologic section in each hole varies greatly in thermal conductivity both between geologic systems and of course within the systems. As an example mean temperature gradients between 100 and 200 meters and between 400 and 500 meters for the various holes are shown in Fig. 1. These gradients can be compared to the gradient contours taken from the two maps previously discussed. The complications are clear. They are also clearly indicated by the data in the hole 31S/20E-22cc (Fig. 2). This hole is a mean gradient of 53^{o}C/km between 100 and 200 meters while mean gradient between 400 and 500 meters is 14^{o}C/km, the main difference being the variation in conductivity between the shale rich Pennsylvanian section and dolomite rich lower Paleozoic section. The thermal conductivity contrast for these two sections of rock is 4-5 with the similar resulting ratio of geothermal gradients.

Thus it is clear that a useful evaluation of low temperature geothermal potential of Kansas is not possible without taking into account the local geologic section (thermal conductivity) and the depths to aquifers (in the case of Kansas these are primarily the lower Paleozoic carbonate rocks). Since detailed temperature and thermal conductivity measurements are not available for many holes, the mean thermal conductivity (or thermal resistivity) of the sedimentary section must be obtained by some other technique. Fig. 3 shows a comparison of geothermal gradient, gamma ray and P-wave velocity logs for hole 13S/2W-32cc. The gamma ray and P-wave data are based on commercially run well logs, digitized at 0.5 intervals and averaged over 3.5 m. The correlation of the gradient with other geophysical logs and lithology is obvious. Based on the data from Kansas, a linear correlation between geothermal gradient, and gamma ray activity and travel time has been established for a predominately carbonate-shale section (Blackwell and Steele, 1981). Thus, within the state of Kansas the mean thermal conductivity of the sedimentary section can be established from well log parameters so that, given the heat flow from the basement, the temperatures at depth in the sedimentary section can be predicted at points between the control wells.

A result of this study was the discovery that shale thermal conductivity values have been overestimated by as much as 50% in literature and that the Paleozoic shale conductivities are as low as 1.0-1.2 $Wm^{-1}K^{-1}$. Thus thick shale sections in the Midcontinent are as thermally "insulating" as the basins in the Basin-Range province or as the Tertiary and Cretaceous section along the East Coast and so the concept of hot spots associated with radiogenic plutons (see Costain, 1977) certainly applies to the Midcontinent as well as to the East Coast. The recent discovery of a radiogenic basement pluton in Illinois with a heat generation of approximately 16μWm^{-3} (Rahman and Roy, 1981) suggests that very high temperatures may occur locally in the Midcontinent and that drilling based on certain types of basement geophysical anomalies might find much higher temperatures than has previously been expected.

Another parameter which must be obtained for low temperature evaluation by interpolation between control wells is the background heat flow. In the Mid-continent this heat flow will be related to basement radioactivity because in Kansas the basement heat flow seems to be related to the Q-A relationship discussed by Birch et al. (1968) and Roy et al. (1968). Thus if the radioactivity of the basement rocks is known, the heat flow can be predicted at any locality. Initial studies have begun but much remains to be done.

A final factor which could significantly effect the temperatures would be lateral motion in the aquifers. The objective of drilling the test wells was to evaluate flow in these aquifers. Detailed temperature measurements and comparisons of ratios of gradients in various lithologies, both above and below the carbonate aquifers, clearly indicates that lateral aquifer flow does not affect temperature measurements (heat flow) in Kansas. The precision with which this conclusion can be established is approximately ±5% so that water flow effects below this value would not be detected.

OREGON

The geologic setting in Oregon is much more complicated than that in Kansas. From the geothermal point of view the state can be divided into approximately four different regions: the northeast, the southeast, the High Cascade Range, and the provinces west of the High Cascade Range. The assessment of the geothermal potential of each of these areas must be addressed on an individual basis. Extensive measurements are available for the state of Oregon (Blackwell et al., 1978; Blackwell and Steele, 1979, unpublished data). A heat flow map for the state is shown in Fig. 4 based on over 200 data points. The division of the state into the four previously mentioned areas is illustrated. Histograms of geothermal gradient are shown in Fig. 5 (Blackwell et al., 1978). In the case of Oregon the distribution of gradient values is approximately the same as the distribution of heat flow because of the relatively small variation in thermal conductivity in the basic volcanic, intrusive and volcanoclastic rocks which make up most of the surface exposure.

The assessment of the coastal provinces is relatively straightforward. Because the gradient and heat flow variations are quite small and because the rocks in general are relatively impermeable so that there do not appear to be many hydrothermal convective systems, the primary resource is merely water from aquifers. The potential of such aquifers can be assessed, given their location and depth, with available data.

The Cascade Range is generally a high temperature province because of its relatively inaccessible nature (distance from low temperature market) with the exception perhaps of the Mt. Hood region, and its assessment represents problems which cannot be addressed in the space available.

The northeastern corner of the state in general has a relatively small scatter in gradient. However there are a significant number of hot springs (Bowen et al., 1978) and so there is some potential for local hydrothermal systems as well as for deep hot water in aquifers. It is clear that the scatter of gradients is somewhat larger than that observed in the coastal provinces.

By far the most complicated region geothermally is the southeastern part of the state. The histogram shown in Fig. 5 illustrates the extreme scatter in the gradient values, ranging from 0 to over 200°C/km as opposed to the coastal provinces where the range is only from 20 to 50°C/km maximum. There are many geothermal systems and many local and regional aquifers in southeastern Oregon. The complicated nature of the heat flow contours (Fig. 4) is related to the local variations caused by the many different factors discussed above. In contrast to Kansas most of the heat flow and gradient variations are related to regional hydrologic circulation, lateral variations in heat flow related to perhaps intrusive centers, or some other source of crustal heat, rather than lateral variations in basement radioactivity and thermal conductivity of the geological section. In this area, the contribution of the crustal radioactivity contrasts to heat flow is essentially negligble. Hence evaluation of the hydrology, the structure and the water geochemistry are much more significant than is the study of basement rock radioactivity.

Because of the complex geology, young volcanism, and high relief, deterministic heat flow measurements were not possible and the average lateral significance of individual measurements is no more than 1 or 2 km (as opposed to 10-20 km in Kansas). In a setting such as this, low temperature geothermal assessment must be either on a specific area basis or on a statistical basis. We have approached the assessment in this area by obtaining as many heat flow and geothermal gradient measurements as possible in existing wells and holes drilled specifically for geothermal evaluation. The assessment is not as grim as it might appear, however, because near most of the populated areas, where low temperature geothermal resources utilization is most likely, there is generally a significant amount of drilling. Therefore it is possible to make scrounge heat flow, gradient, and temperature measurements directly in the areas of interest and thus one does not have to resort to statistics in evaluating many areas of high population density.

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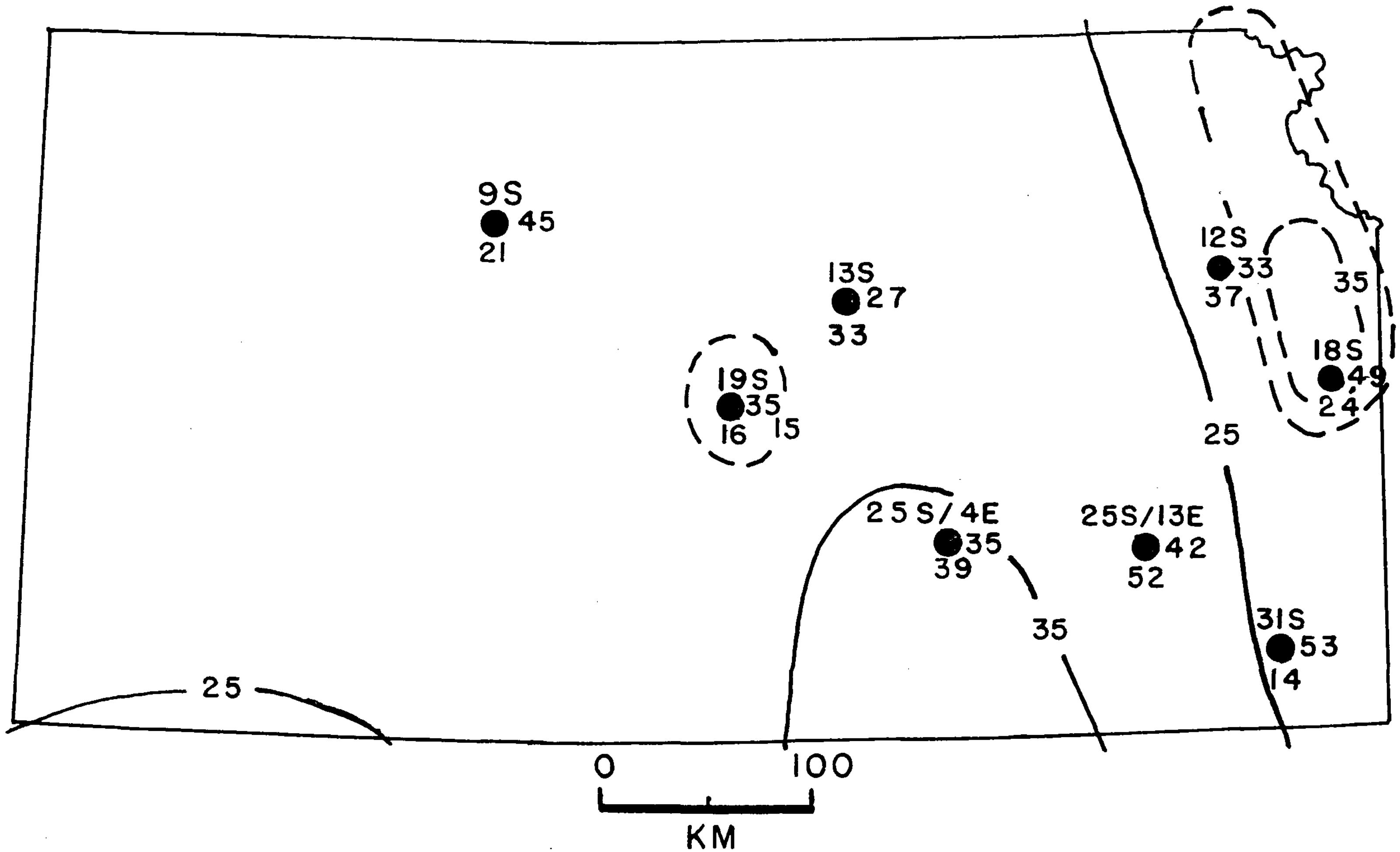
Figure 1. Geothermal gradients in Kansas. Geothermal-gradient-to-2-km contours (Guffanti and Nathenson, 1980) shown with solid lines. Geothermal gradient contours shown with dashed lines from Kron and Heiken (1980). Hole locations shown are from Blackwell and Steele (1981). Hole identification is Township number. Gradients between 100-200 and 400-500 meters shown at right and below location dot respectively.

Figure 2. Temperature-depth and gradient data for Kansas hole 31S/20E-22 cac (Blackwell and Steele, 1981).

Figure 3. Bar graphs of gradient, P-wave velocity, and natural gamma ray activity and a generalized geologic section for Kansas hole 13S/2W-32 ccc.

Figure 4. Heat flow and physiographic province map of Oregon. Heat flow contours are shown in 20mWm^{-2} intervals (0.5 μcal/cm^2 sec). Data from Blackwell _et al_ (1978).

Figure 5. Histograms of geothermal gradient for various combinations of physiographic provinces in Oregon (Blackwell _et al_, 1978). N is the total number of samples.



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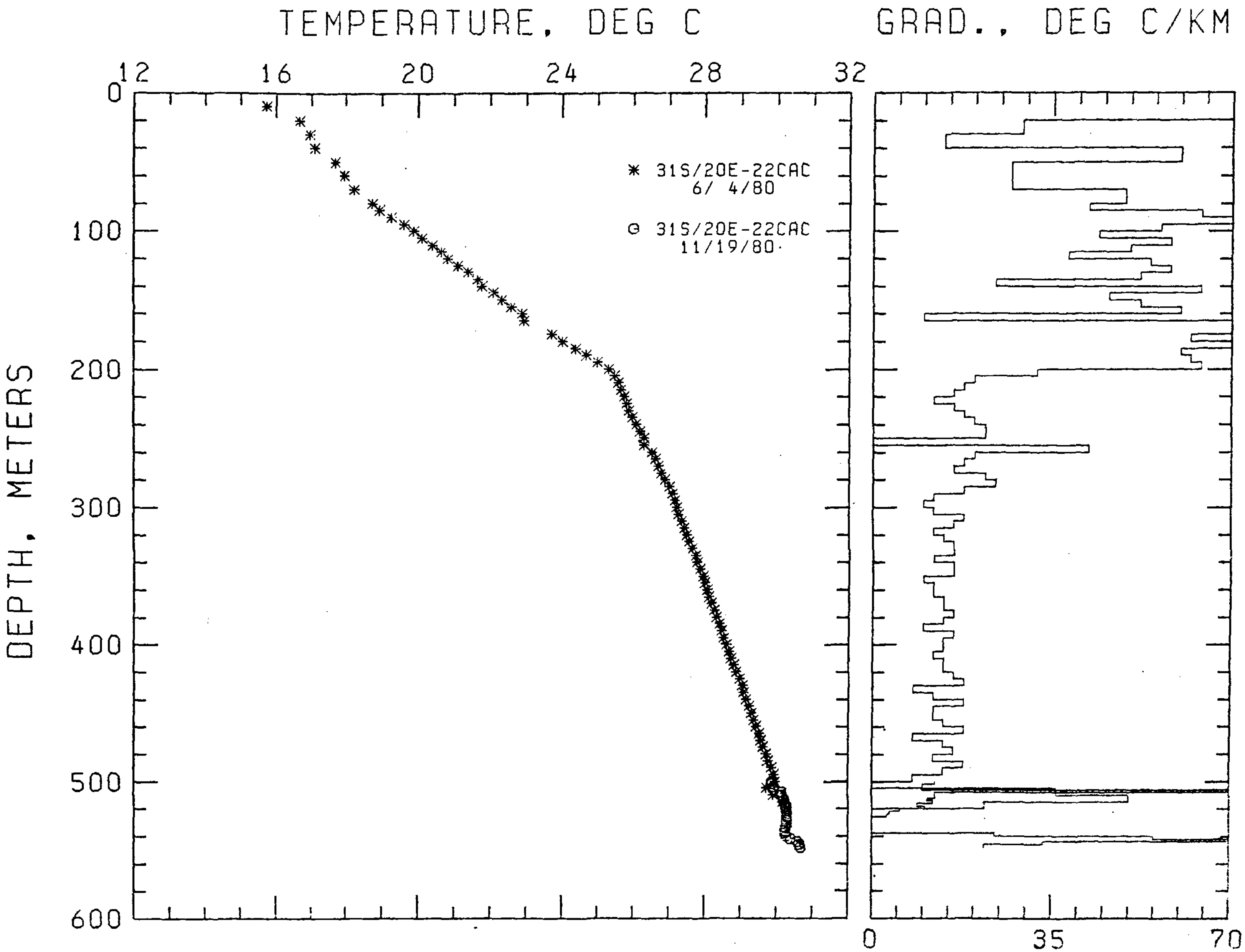
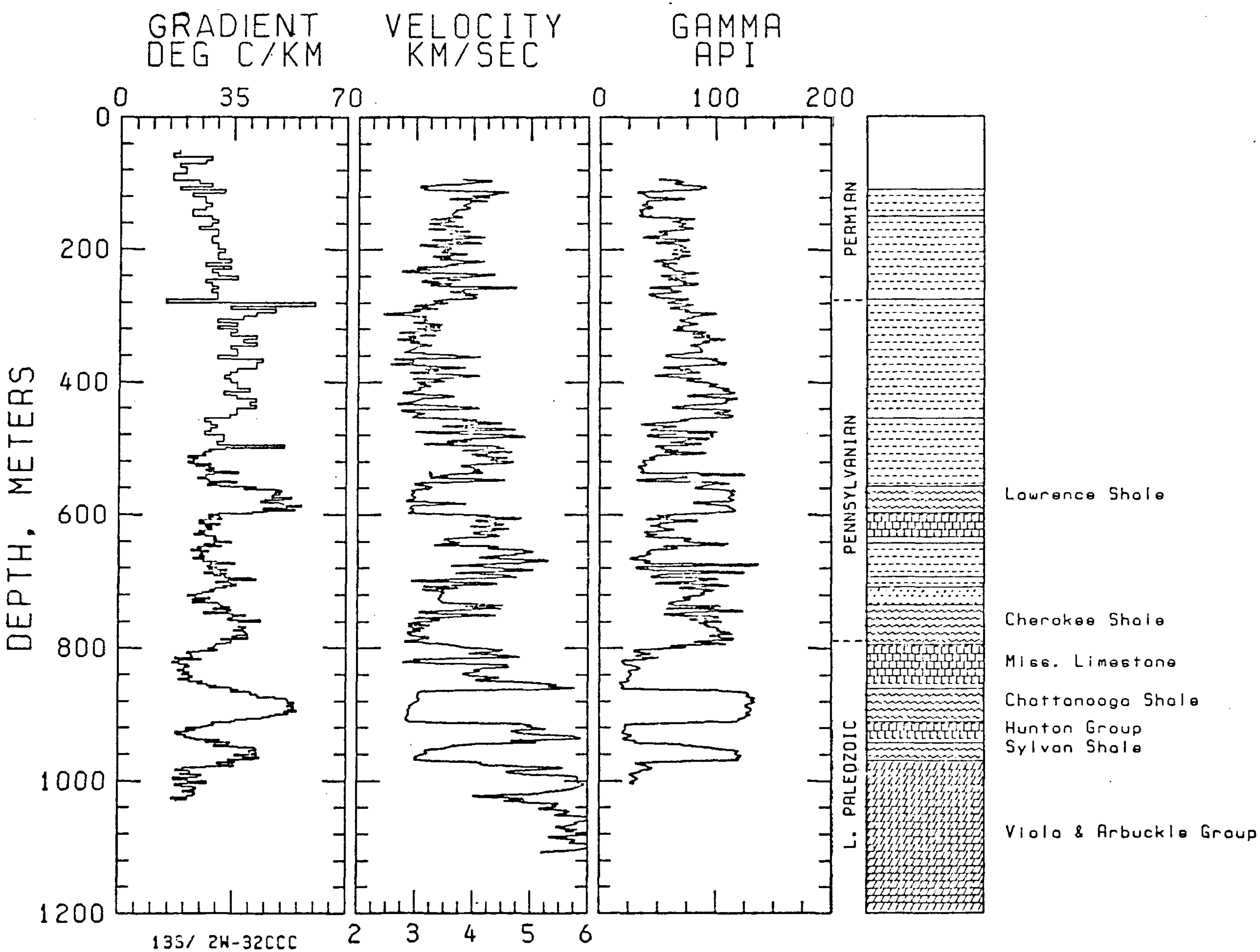


Figure 2. Temperature-depth and gradient data for Kansas hole 31S/20E-22 cac (Blackwell and Steele, 1981).



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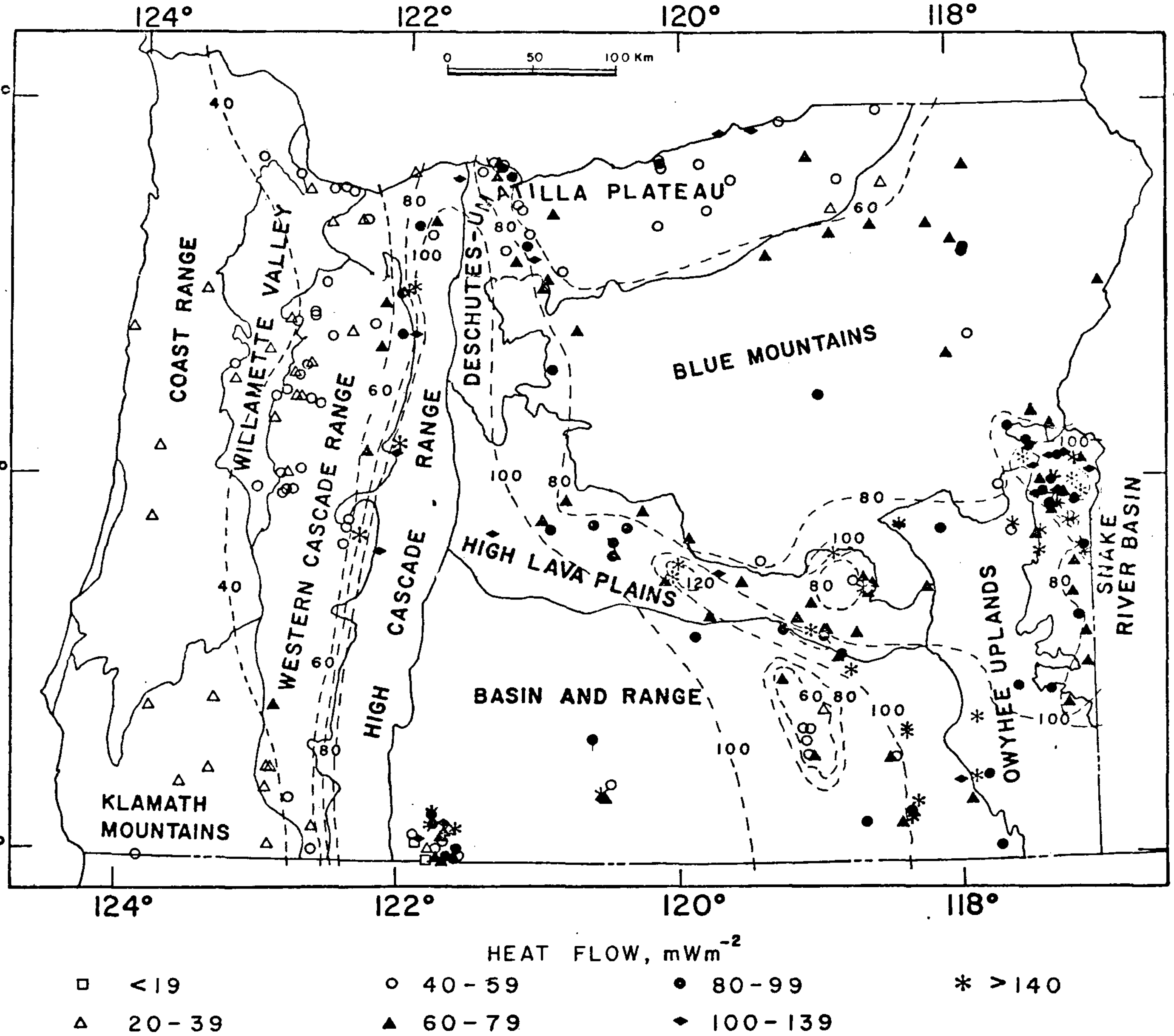

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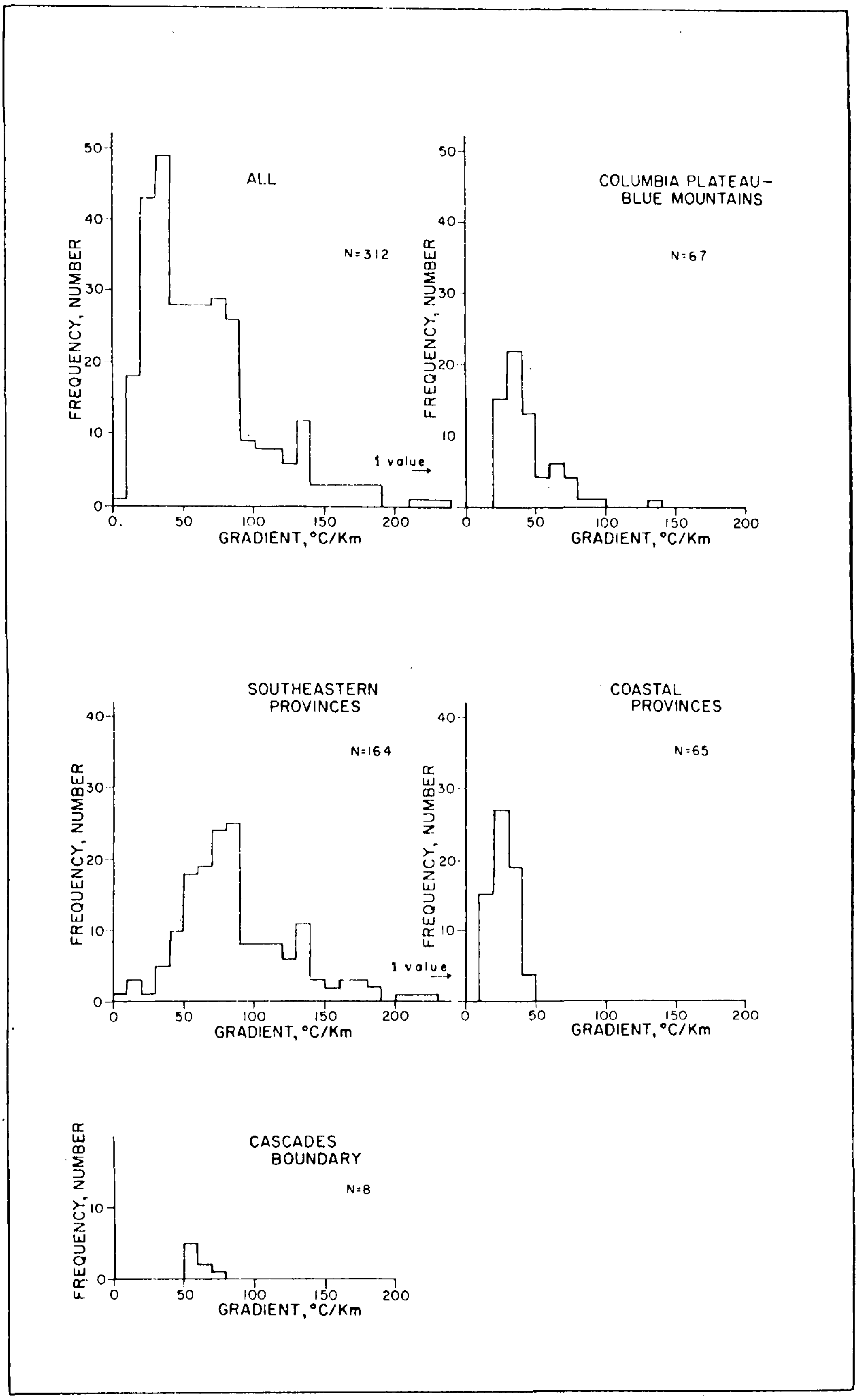


Figure 5. Histograms of geothermal gradient for various combinations of physiographic provinces in Oregon (Blackwell et al, 1978). N is the total number of samples.

RESISTIVITY METHODS IN EXPLORATION FOR HYDROTHERMAL RESOURCES

George R. Jiracek
Department of Geological Sciences
San Diego State University
San Diego, CA 92182

Introduction

This paper discusses the practical aspects of using d.c. resistivity in the exploration for hydrothermal resources. There are several reasons why low electrical resistivity is expected in hydrothermal aquifers but the association is not without pit falls. Besides outlining the reasons why resistivity has proven a successful geothermal exploration tool, a section on how resistivity is practiced is included. Here, the common electrode arrays are considered with their major advantages and disadvantages pointed out. The current status in resistivity interpretation schemes is touched upon with emphasis on computer modeling. Finally, a successful resistivity case history of a low-temperature resource at Las Alturas Estates, New Mexico is included to illustrate a specific resistivity exploration philosophy. The case history concludes with drilling results which are, of course, the ultimate test.

Hydrothermal Resistivity Model

From the outset we should understand that the actual complex hydrogeological situation undergoes a simplifying transformation in the face of resolution and modeling capabilities in resistivity. Figure 1 illustrates a hydrothermal target which is a thermal aquifer perhaps first detected by a warm well. The plumbing system may be some complicated migration of thermal waters up various fault zones. We visualize this situation as a resistivity model with a conductive slab approximating the aquifer which is surrounded by electrically resistive blocks. Rarely can the plumbing system be seen with resistivity and usually surface topography is not considered. Dipping fault contacts are usually modeled as stepped resistivity boundaries (Figure 1).

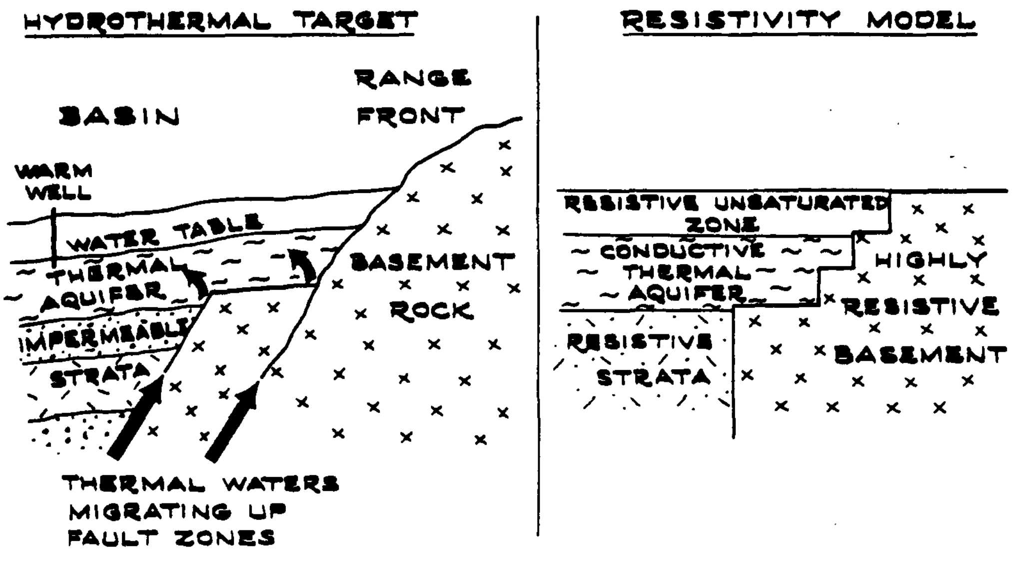


Figure 1. Comparison of complex hydrogeothermal target and simplified resistivity model.

Use of Resistivity in Hydrothermal Exploration

The two major uses of resistivity measurements in hydrothermal exploration are outlined in Figure 2 as: 1) detection and delineation of aquifers and 2) estimation of aquifer temperature and water content.

USE OF RESISTIVITY IN HYDROTHERMAL EXPLORATION

DETECTION AND DELINEATION OF AQUIFERS

HYDROTHERMAL AQUIFERS HAVE LOW RESISTIVITY BECAUSE:

1. Water **Decreases Resistivity**
2. Increased temperature **Decreases Resistivity** of Water (Increased Ion Mobility)
3. Increased Water Temperature Increases Salinity by Dissolving More Minerals Which **Decreases Resistivity** (Increased Number of Charge Carriers)
4. Hydrothermal Alternation Minerals **Decrease Resistivity** (e.g. Igneous Rocks into Clay and Zeolites)

WARNING: Nonthermal Clay Zones May Have Resistivities Indistinguishable from Hydrothermal Aquifers.

ESTIMATION OF AQUIFER TEMPERATURE AND WATER CONTENT

LOWER RESISTIVITY IMPLIES HIGHER TEMPERATURE

POROSITY ESTIMATION BY ARCHIE'S LAW

$$\frac{\rho_R}{\rho_W} \simeq \Phi^{-2}$$

ρ_R: Resistivity of Saturated Rock

ρ_W: Resistivity of Saturating Water

Φ: Fractional Porosity

WARNING: Archie's "law" Is Empirically True for Clay-Free Aquifers Only.

Figure 2

First, and foremost, is the detection and delineation of aquifers. Here, the key property is that hydrothermal aquifers have low resistivity. This is true because of the inherent effect of water itself coupled with the temperature-resistivity dependencies of ion mobility, salinity, and hydrothermal mineral alternation. Despite all of these factors favoring the association of low resistivity with hydrothermal resources we are cautioned (Figure 2) that nonthermal clay zones may have d.c. resistivities indistinguishable from hydrothermal aquifers. This warning cannot be overemphasized; it is the most serious limitation regarding the identification of a thermal aquifer with resistivity alone. Thermal confirmation such as a warm well, a hot spring, or a high thermal gradient is necessary to connect low resistivity with a thermal anomaly.

The second application of resistivity is in the estimation of aquifer temperature and water content. Simply stated, lower resistivity implies higher temperature; however, quantitative relationships between fluid resistivity and temperature often yield incorrect temperature estimates in practice. Again, the presence of clay in the hydrothermal aquifer will produce erroneous results. An estimate of the amount of water-filled porosity can be accomplished by using some form of Archie's law. The simplist form of the equation is given in Figure 2. Here, a surface measurement of the resistivity of the saturated aquifer (ρ_R) is coupled with the resistivity value of the saturating fluid (ρ_W), e.g. obtained from a spring or well, to compute the fractional porosity, ϕ. Clay is again the major problem in such a calculation since Archie's relationship is really not a "law" but is empirically true for clay-free aquifers only. The application of Archie's relationship in geothermal exploration using surface resistivity is discussed by Meidav (1970).

Resistivity Techniques and Common Arrays

Resistivity techniques can be described by three categories as listed in Figure 3. These are: 1) vertical sounding accomplished by expanding an array and effectively sensing to greater depths, 2) horizontal profiling using a constant array spacing moved laterally along the surface, and 3) a combination of sounding and profiling to produce either a section or a map. Various electrode arrays are used to accomplish these purposes; the Wenner, Schlumberger, equatorial, dipole-dipole, and bipole-dipole arrays are the most common. The resistivity case history included later in this paper presents examples of one-dimensional (vertical) sounding, two-dimensional sounding-profiling producing a resistivity pseudosection, and bipole-dipole mapping allowing a three-dimensional interpretation.

RESISTIVITY TECHNIQUES AND COMMON ARRAYS

1. **VERTICAL SOUNDING (1-D)**
   WENNER, SCHLUMBERGER, EQUATORIAL ARRAYS
2. **HORIZONTAL PROFILING (LIMITED 2-D)**
   WENNER, SCHLUMBERGER ARRAYS
3. **COMBINED SOUNDING — PROFILING**
   PSEUDOSECTION (2-D)
   WENNER, SCHLUMBERGER, DIPOLE-DIPOLE ARRAYS
   MAPPING (3-D)
   BIPOLE-DIPOLE (ROVING DIPOLE) ARRAY

Figure 3

The common arrays where the voltage and current electrodes are emplaced in a collinear fashion are the Wenner, Schlumberger, and dipole-dipole arrays (Figure 4). Probably the best-known array is the Wenner array which has four equally spaced contacts with the earth; current is usually passed through the

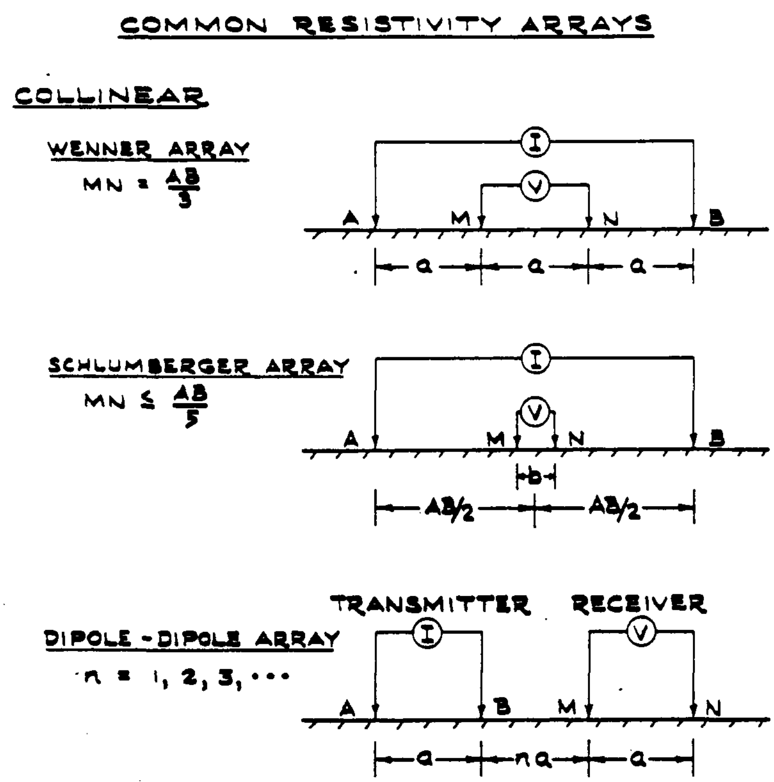


Figure 4

two outer electrodes (A and B) and the resulting voltage drop is measured by the inner electrode pair (M and N). The Schlumberger array uses the MN electrodes much closer together, less than one-fifth the outer (AB) spacing. The dipole-dipole array separates the current transmitting dipole (AB) from the voltage receiving dipole (MN) by integer multiples of the dipole spacing (a).

There are two common areal resistivity arrays shown in plan views in Figure 5. In the equatorial array the voltage measuring MN electrodes are maintained parallel to a fixed AB current source but are, additionally, moved along the perpendicular bisector (or equator) of AB. It is obvious that the equatorial array is identical to the Schlumberger array (Figure 4) in the limiting case when MN is between AB, providing MN $\leq$ AB/5. The bipole-dipole technique in Figure 5 employs a long, fixed AB current source called a bipole to distinguish it from the much shorter MN voltage measuring dipoles. Usually two voltage dipoles (Figure 5) are used to calculate the gradient of the potential at each location which is the total vector electric field. The bipole-dipole method is also called roving dipole since voltage measurements are made by roving about the fixed current source. Further discussion of the particulars of various resistivity arrays may be found in Keller and Frischknecht (1966) and Zohdy (1970).

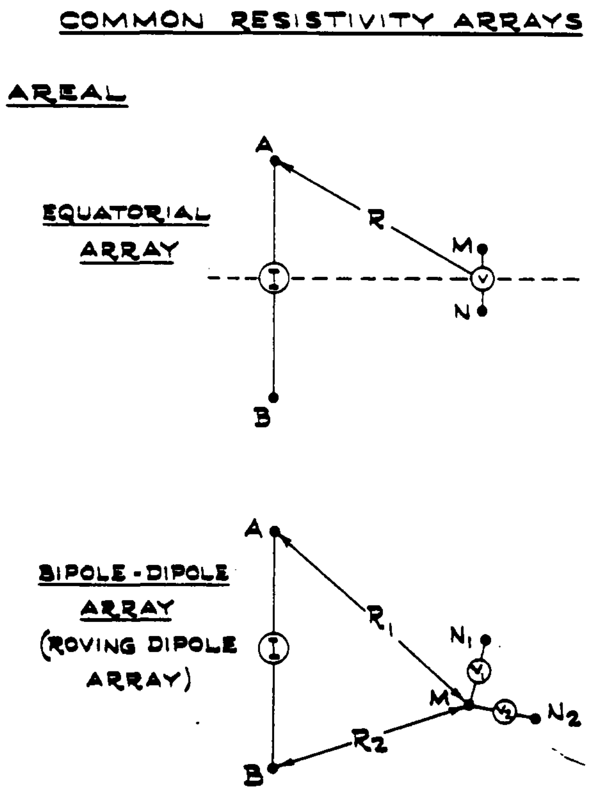


Figure 5

It is very important to consider the advantages and disadvantages of the various resistivity arrays before any field data are collected. The expected resistivity target, the logistics of the survey area, and the equipment capabilities are all crucial factors. These are usually not adequately appreciated prior to hiring a contractor or embarking on one's own field survey. A good practical comparison of the Wenner, Schlumberger, and dipole-dipole arrays appears in Zohdy et al. (1974). An excellent presentation of the use of various resistivity arrays in geothermal exploration is included in the Geothermal Resources Council Technical Training Course on "Geophysical Exploration Methods for Geothermal Resources". A considerably abridged version of the training material has been published by the instructor (Meidav, 1974).

Figure 6 summarizes the major advantages and disadvantages of the three common collinear resistivity arrays. The major advantage of the Wenner array is the large voltage drop that one obtains with a given current input. This is important if the field equipment has a limited capability. The only other advantage is the simple apparent resistivity formula; however, this is hardly a consideration, especially when using a pocket calculator in the field. The major disadvantage of the Wenner array is that a constant "a" spacing (Figure 4) must be maintained when expanding the array during soundings. Consequently, all four electrodes must be moved for each expanded spacing. Besides the additional time considerations, the movement of all four electrodes to new locations results in a higher susceptibility to surficial resistivity irregularities. Since the current and voltage wires are adjacent to each other, one must consider the possibility of errors due to inductive coupling. This may not be serious and is easily corrected for by physically separating the wires.

ADVANTAGES AND DISADVANTAGES OF COMMON RESISTIVITY ARRAYS

	ADVANTAGES	DISADVANTAGES
WENNER ARRAY	1. **Large Voltage Drop for Given Current** 2. Simple Apparent Resistivity Formula	1. **Sounding Requires Moving All Four Electrodes** 2. More Susceptible to Surficial Resistivity Irregularities 3. Coupling Between Wires
SCHLUMBERGER ARRAY	1. **Sounding Usually Requires Moving Only Outer Two Electrodes** 2. Less Susceptible to Surficial Resistivity Irregularities 3. Slightly Greater Probing Depth than Wenner for Given Outer Spacing 4. Inverse Schlumberger Allows Safer Operation 5. Inverse Schlumberger Reduces Requirement of Heavy, Expensive Current Wire	1. **Lower Voltage Drop for Given Current** 2. Coupling Between Wires 3. Inverse Schlumberger More Susceptible to Telluric Noise
DIPOLE-DIPOLE ARRAY	1. **Very Sensitive to Lateral Resistivity Variations** 2. Shorter Lines Required for Given Probing Depth 3. Reduced Coupling Between Wires	1. **Low Voltage Drop for Given Current** 2. Very Susceptible to Surficial Resistivity Irregularities 3. Complicated Pseudosections

Figure 6

The Schlumberger array has many advantages over the Wenner array (Figure 6) and should be used rather than the Wenner array provided the field equipment can cope with the disadvantages of a lower voltage drop. This means a more sensitive voltage measuring device and/or a higher current source. Since the Schlumberger array requires only that MN is maintained less than AB/5 (Figure 4), a sounding can be accomplished by moving only the outer electrodes. In practice, a point is reached where the MN spacing must be expanded for a measureable voltage but usually several AB spacings can be used with each MN spacing. This aspect of the Schlumberger array is a major advantage logistically with considerations to manpower and time; it also reduces the susceptibility to surficial irregularities since only two electrodes are changed, not four as with the Wenner array. The slightly greater probing depth for a given spacing with the Schlumberger array compared to the Wenner array is a minor advantage. However, the advantages of the inverse Schlumberger array (Figure 6) are very important and the array is used by most experienced resistivity practitioners.

The inverse Schlumberger array is obtained by the simple exchange of the current and voltage electrodes, i.e. current is passed through MN and voltage drop is measured between A and B. Identical values of apparent resistivity result from this interchange (even for an anisotropic earth) as a consequence of the theorem of reciprocity (Keller and Frischknecht, 1966). Reciprocity is always a good check in the field to help insure that valid results are being obtained. Additionally, there is a major safety and logistical advantage using the inverse array (Figure 6). For example, when spacings are large, say AB is several kilometers, the MN current spacing may only be a few hundred meters in the inverse case. Such MN electrodes (and current wires) would be in sight of the transmitter operator rather than out of view, over a kilometer away, in the case of the normal Schlumberger operation. Besides the safety considerations, since the current wire is often heavier and more expensive, the shorter MN current source can save money and setup time.

Along with the lower voltage drop using the Schlumberger array as discussed previously, the only significant disadvantage is the increased telluric noise at large AB spacings when employing the inverse array. The time varying voltages due to telluric or natural earth currents increase linearly with contact spacing, thus degrade the received signals.

The dipole-dipole method has been shown to be very sensitive to lateral resistivity variations (Beyer, 1977). This is an advantage when looking for such boundaries but it, unfortunately, also results in high susceptibility to surficial irregularities (Figure 6). The dipole-dipole method has the advantages of shorter lines for a given probing depth and the decrease of inductive coupling between wires. However, the physical separation of the current and voltage dipoles results in much smaller voltage drop for a given current. This requires a larger current generating source when doing dipole-dipole soundings. The pseudosection method of plotting dipole-dipole results produces very complicated plots which can only be effectively analyzed by computer modeling. Dipole-dipole pseudosection plotting and interpretation is discussed later in the Las Alturas Estates field study.

Resistivity Interpretation

Figure 7 summarizes the current status of the interpretation of resistivity vertical sounding, horizontal profiling, and combined sounding-profiling. None of the resistivity results obtained in practice are amenable to unambiguous interpretation although layered, sounding results are theoretically unique.

Resistivity sounding results have been classically analyzed by logarthmic curve matching using albums of theoretical master curves such as those presented by Orellana and Mooney (1966). These curves, in conjunction with auxiliary charts, require considerable time even in experienced hands. Consequently, the manual methods have been superseded in cases where many layers are evident by automatic computer routines. The layered inversion program published by Zohdy (1974) is readily available and widely used. Probably the most readily available two-dimensional resistivity modeling program is that of Dey and Morrison (1976) and Dey (1976). Dey and Morrison's finite difference routine is not automatic since it is a forward solution not an inversion. Thus, the desired two-dimensional model is obtained by trial-and-error matching. No three-dimension resistivity computer code is readily available; however, a catalog of theoretical bipole-dipole results has been published by Hohmann and Jiracek (1979).

RESISTIVITY INTERPRETATION

1. **VERTICAL SOUNDING**
   Theoretically Unique Solutions (Nonunique in Practice)

   Logarithmic Plotting of Sounding Data Facilitates Curve-Matching
   Using Master Curves

   Fast, Efficient 1-D Computer Programs
   Readily Available

2. **HORIZONTAL PROFILING**

   Nonunique Solutions

   2-D Computer Programs Available

3. **COMBINED SOUNDING-PROFILING**

   Nonunique Solutions

   PSEUDOSECTIONS
   2-D Computer Programs Available

   MAPPING
   3-D Computer Programs Not Readily Available

Figure 7

Las Alturas Estates Resistivity Case History

Introduction. The proximity of New Mexico State University to domestic wells producing warm water at Las Alturas Estates motivated the electrical resistivity evaluation of the prospect. Figure 8 is the base map of the area showing the campus located 2 km northwest of the two warmest domestic wells. Except for Tortugas Mountain (Figure 8), which is a 200 m high outcrop of Permian limestone, the area is covered by recent piedmont-slope alluvium. The Las Alturas prospect has no surface geothermal manifestations; a well drilled in 1960 at T.23 S, R.2 E, Sec. 34.214 (Figure 8) discovered water of 45°C at 100 m depth.

Resistivity evaluation of the area will be discussed in a sequence progressing from one-dimensional sounding, to two-dimensional combined sounding-profiling, and finally three-dimensional mapping. Crossed inverse Schlumberger soundings were centered on the warmest well (Figure 8). Two dipole-dipole lines were measured in east-northeast directions across the area and over 110 roving dipole points were used to map approximately 40 km^2 surrounding the bipole transmitters shown in Figure 8.

Schlumberger soundings. Crossed Schlumberger soundings centered at the warmest well (Figure 8) are presented in Figure 9 with five layer interpretations using the inversion method developed by Zohdy (1974). Both soundings clearly detect a significant conductive zone beyond approximately 30 m AB/2 spacing. Close scrutiny of the field results reveals that beyond this spacing the 78° sounding systematically yields apparent resistivity values less than the 348° sounding. This is considered to be a consequence of the paradox of anisotrophy (Keller and Frishknecht, 1966) whereby the minimum apparent resistivity is measured perpendicular to the strike of the conductive body. Hence, the conductive zone extends more north-south than east-west. Interpreted results in Figure 9 have not been corrected for anisotrophy; however, the interpreted conductive layers of 8 and 9 ohm-m are identified with the thermal aquifer. Considering all of the results (including well data and the dipole-dipole results to be described) the 78° sounding is considered more representative of the section. Thus, the shallow geothermal target is estimated to extend in depth from approximately 100 to 250 m and to be roughly 10 ohm-m resistivity.

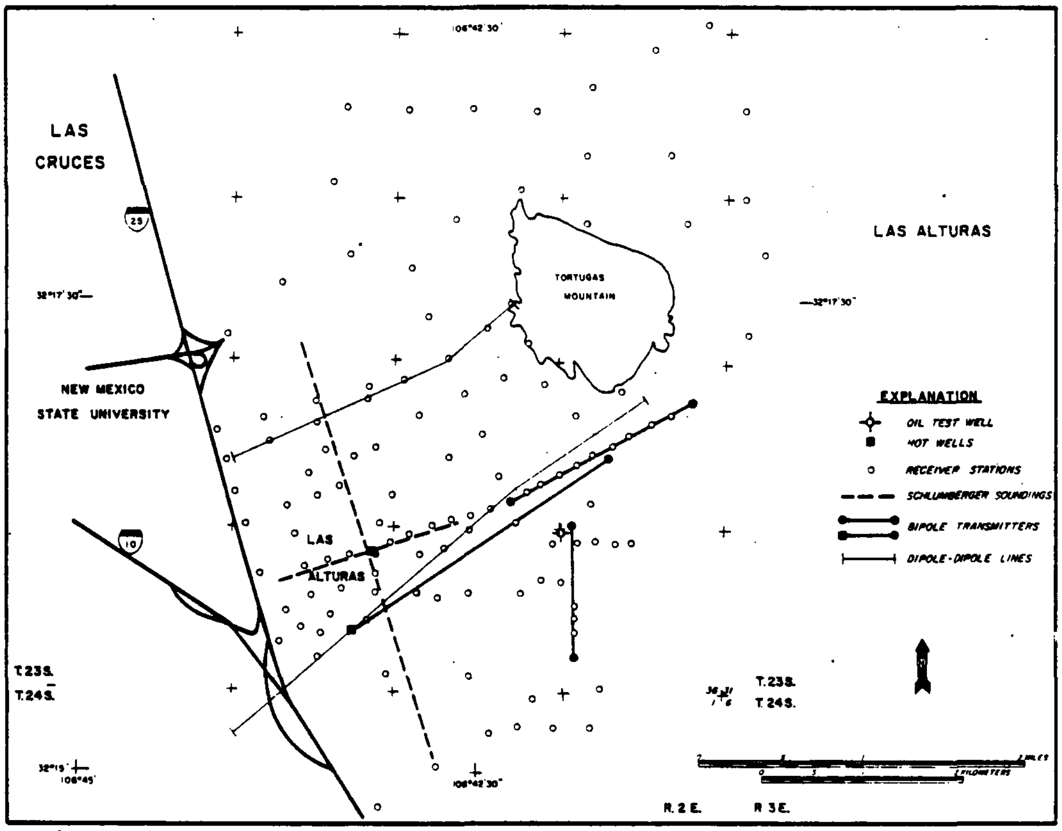
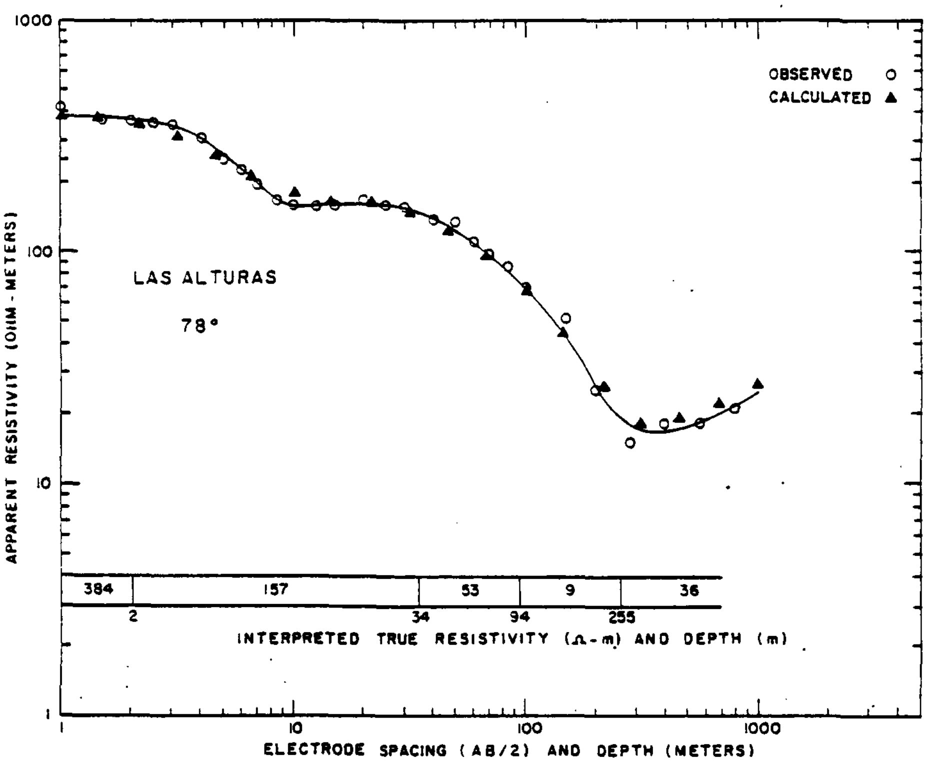


Figure 8. Las Alturas Estates base map (after Smith, 1977).

Dipole-dipole sounding-profiling. To further define the conductive zone associated with the geothermal aquifer, two dipole-dipole profiles were surveyed across the Las Alturas prospect. The lines were oriented (Figure 8) to better define the east-west extent of the conductive target. Figure 10 illustrates the method of plotting dipole-dipole apparent resistivity results in so-called pseudosection form. The values are first plotted at depth points determined by the intersection of lines drawn at 45° angles from the centers of the current and voltage dipoles. The combined values are then contoured to give the final pseudosection. Figure 11 presents the observed pseudosection for the southernmost dipole-dipole line (Figure 8) together with two-dimensional numerical model calculations (Dey, 1976). The modeling results are in good agreement



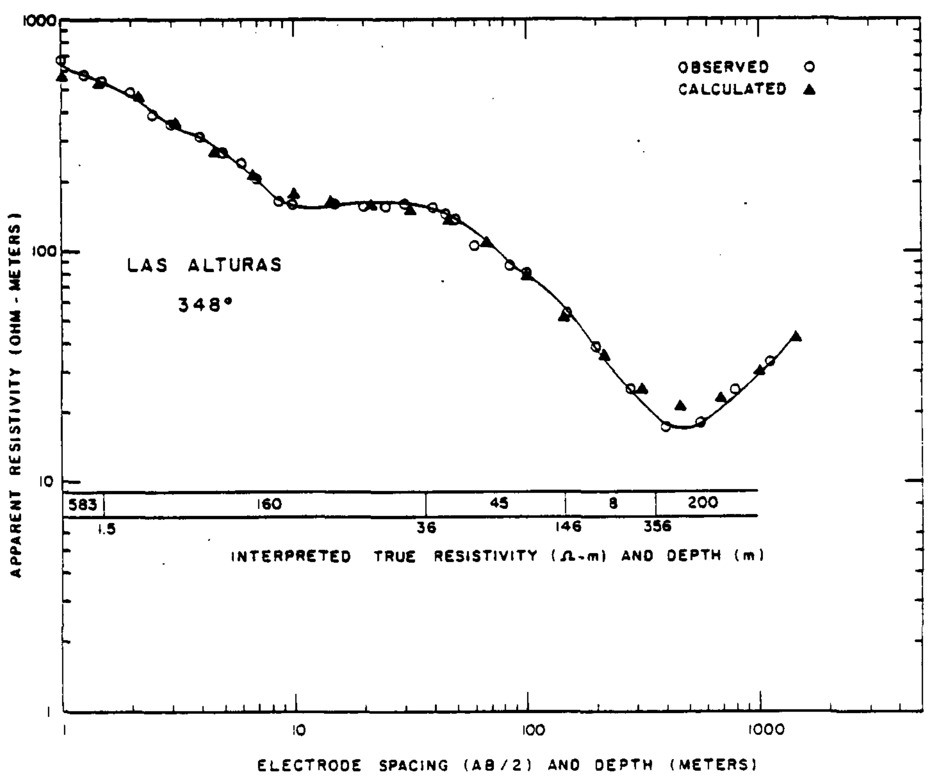


Figure 9. Crossed Schlumberger soundings at Las Alturas Estates and interpreted five-layer models (after Smith, 1977).

with the Schlumberger results (Figure 9). Furthermore, the independent two-dimensional modeling allows more quantitative conclusions to be drawn. For example, the geothermal zone (~10 ohm-m) clearly extends more to the east from the warm wells in Figure 8 than west; the southwestern extent of the layer is terminated laterally 600 m from the well at station 9 by a resistive (~50-100 ohm-m) barrier. The zone is bounded on the east (~ station 42) by a very prominent resistive block (~300 ohm-m), the depth to the top of the 10 ohm-m layer varies from about 40-100 m, and the bottom of the zone appears to be at about 300 m. The conductive geothermal reservoir is similarly evident on the northern dipole-dipole line. Using the results from both dipole-dipole lines we are now able to define the lateral boundaries of the conductive aquifer along the survey lines. These boundaries are shown in Figure 12 by heavy black marks. Such delineation is clearly not sufficient for complete reservoir assessment but it does provide valuable lateral bounds on the conductive aquifer. Such boundaries are important in the placement of bipole current sources for the most effective mapping (Hohmann and Jiracek, 1979).

DIPOLE-DIPOLE PLOTTING METHOD

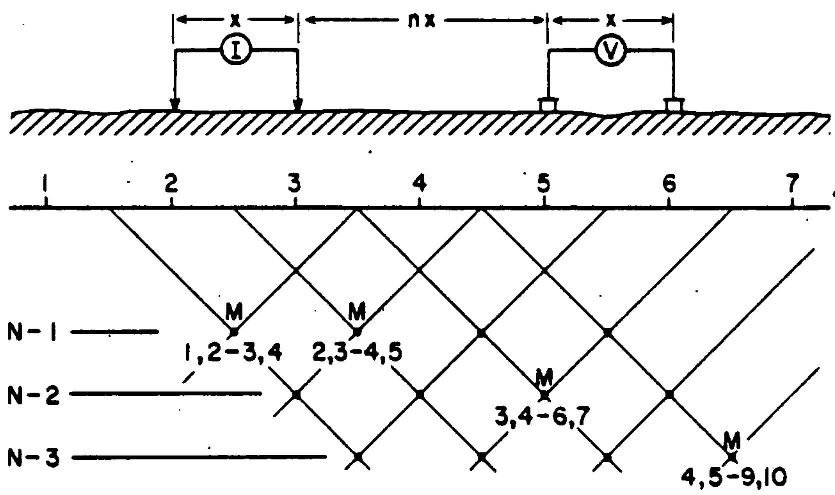


Figure 10

<u>Bipole-Dipole Mapping</u>. The 2 km bipole transmitter shown in Figure 13 has been the most effective in mapping the boundaries of the shallow conductive aquifer at Las Alturas Estates. Figure 13 was prepared by contouring the total field apparent resistivity values which were plotted at the reciever locations marked in Figure 8. Regions of apparent resistivity less than 60 ohm-m are shaded by the dot pattern in Figure 13.

Conductive regions (<60 ohm-m) mark the area of the hot wells in Figure 13. This zone is sharply circumscribed by a resistive pattern (>60 ohm-m) on the east, north, and west sides of the survey area. Three-dimensional modeling was applied interactively to these results to determine to what extent the observed patterns reflect the true subsurface resistivity distirbution, i.e., the areal extent of the shallow geothermal aquifer.

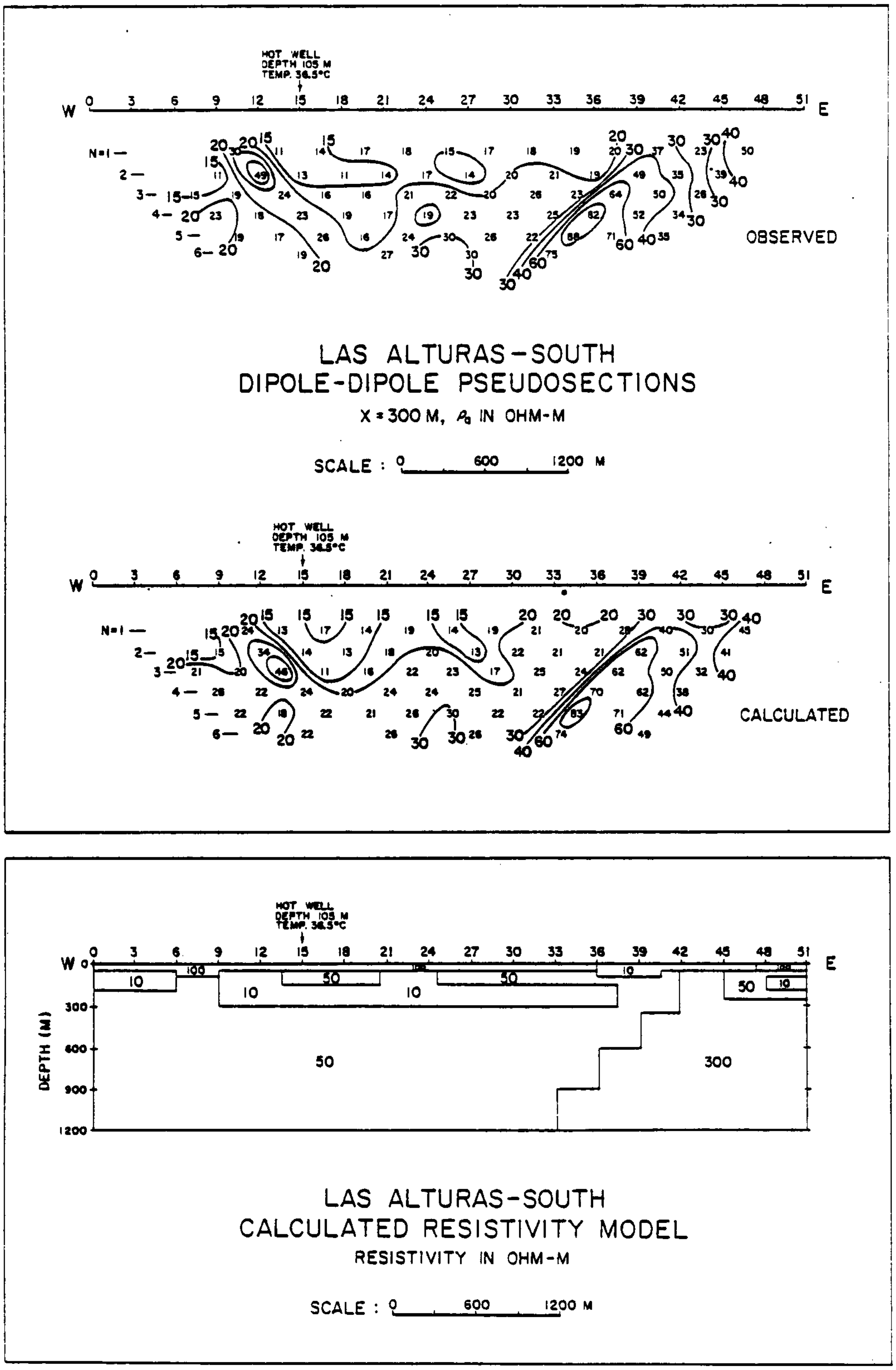


Figure 11. Dipole-dipole pseudosection at Las Alturas Estates-south and two-dimensional model calculations (from Jiracek and Gerety, 1978).

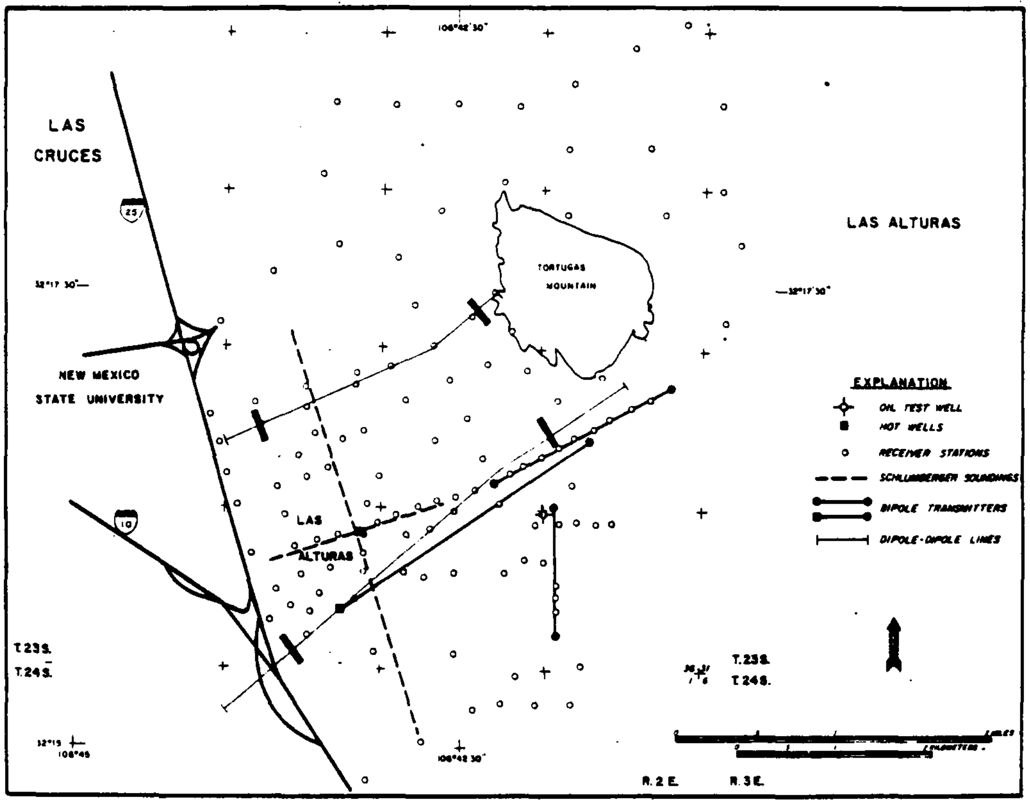


Figure 12. Lateral boundaries (heavy marks) of conductive hydrothermal aquifer along dipole-dipole survey lines at Las Alturas Estates.

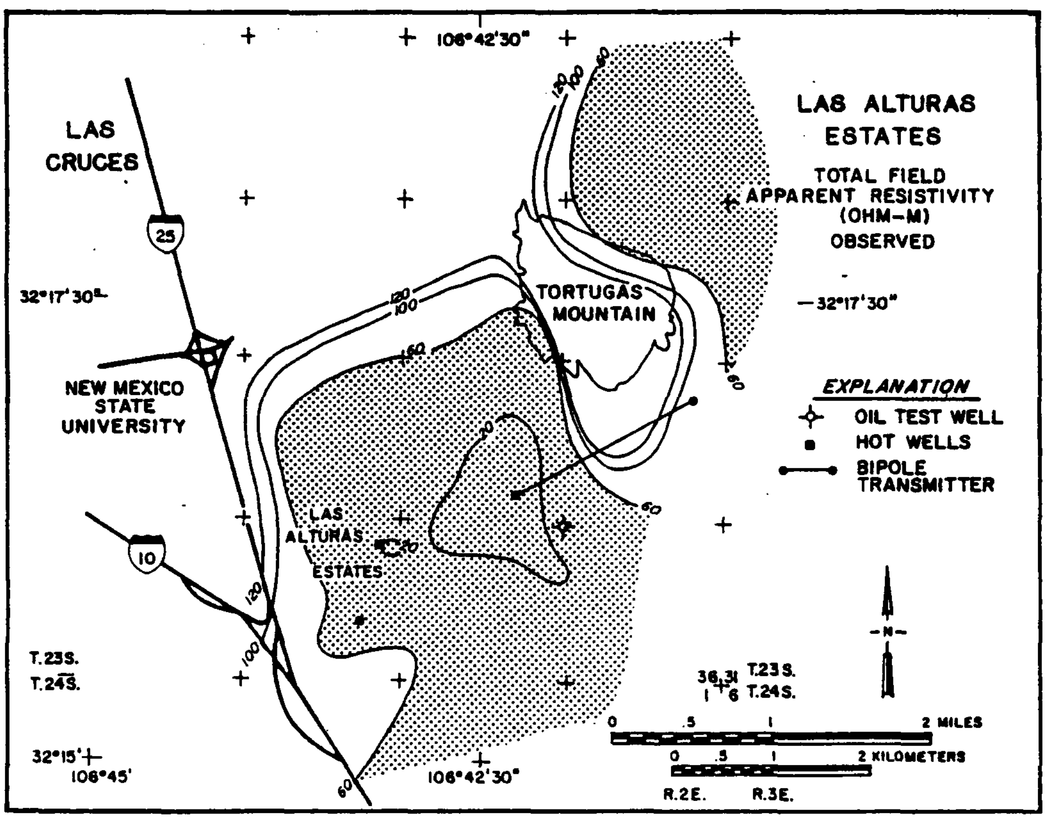


Figure 13. Observed total field apparent resistivity map of Las Alturas Estates (from Hohmann and Jiracek, 1979).

Figure 14 shows a plan view of the three-dimensional model used to approximate the conductive geothermal reservoir. This model was defined from the combined Schlumberger and dipole-dipole modeling (Figures 9 and 11) and by a comparison of the observed bipole-dipole patterns with theoretical results presented by Hohmann and Jiracek (1979). The reservoir is approximated by a slab of 10 ohm-m material which is 100 m deep and 500 m thick. The body has an approximately north-south length of 6 km and east-west width of 2.5 km. The more complex resistive body of 300 ohm-m to the east of the conductive slab in Figure 14 models the resistive limestone of Tortugas Mountain and its buried extension to the south. The conductive and resistive slabs are immersed in a half-space of 75 ohm-m.

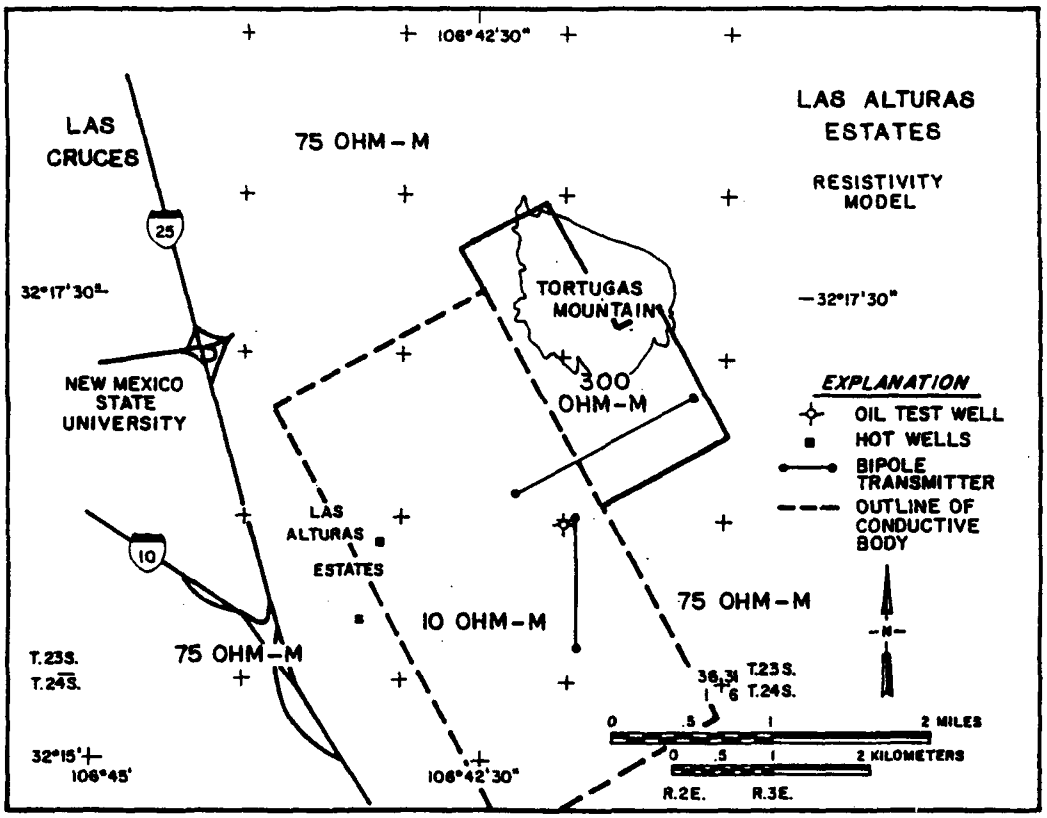


Figure 14. Simple three-dimensional resistivity model of Las Alturas Estates (from Hohmann and Jiracek, 1979).

A comparison of the theoretical results in Figure 15 with the field data of Figure 13 reveals a remarkable similarity. It is apparent that the actual body is wider than the 2.5 km body modeled and may extend farther to the south. However, the 60 ohm-m contour may be considered as very nearly outlining the body in Figure 13. It is emphasized that no attempt was made to model the regions near the bipole electrodes which reflect shallow resistivity variations. The main interest was in duplicating the major patterns of the field results using all available constraints, e.g., Schlumberger, dipole-dipole, and drilling results. Results of the bipole-dipole mapping allow us to outline the areal extent of the conductive aquifer as shown in Figure 16. This outline should be taken as approximate only; dipole-

dipole boundaries (included in Figure 16) are better constrained. The outline is dashed to the south and southeast since our field data do not permit complete definition of this sector.

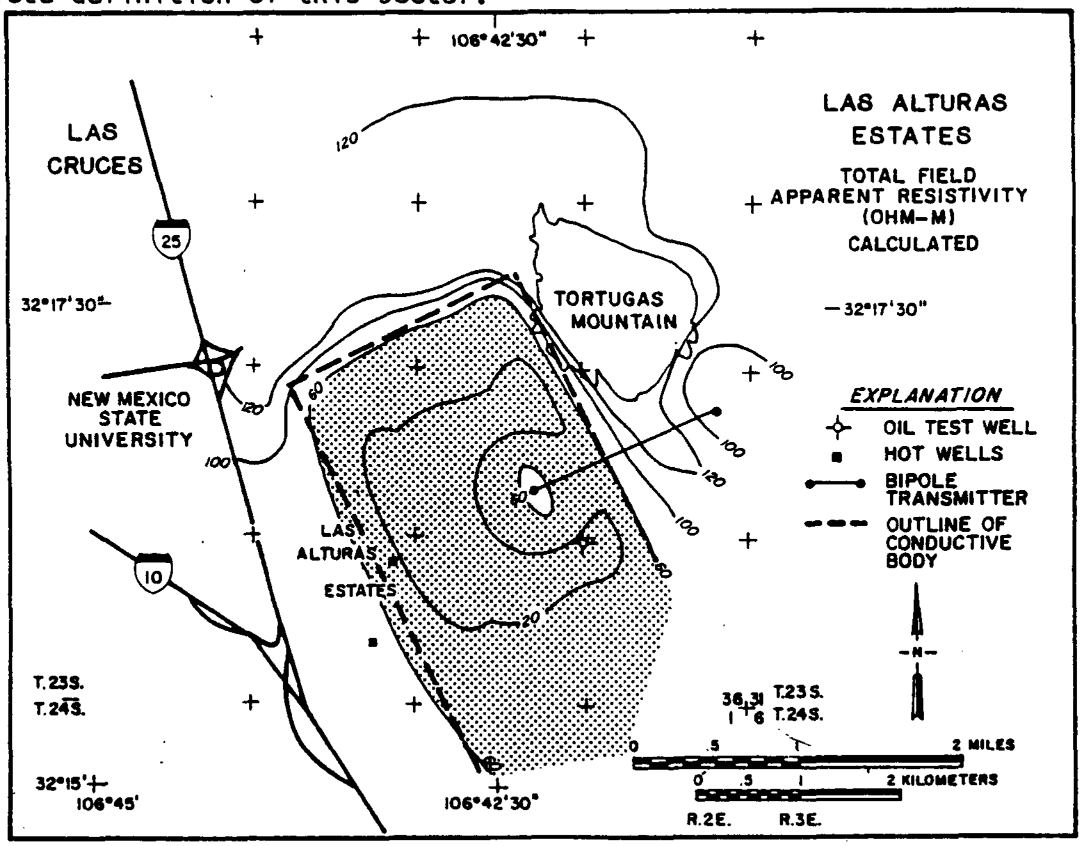


Figure 15. Calculated total field apparent resistivity map of Las Alturas Estates (from Hohmann and Jiracek, 1979).

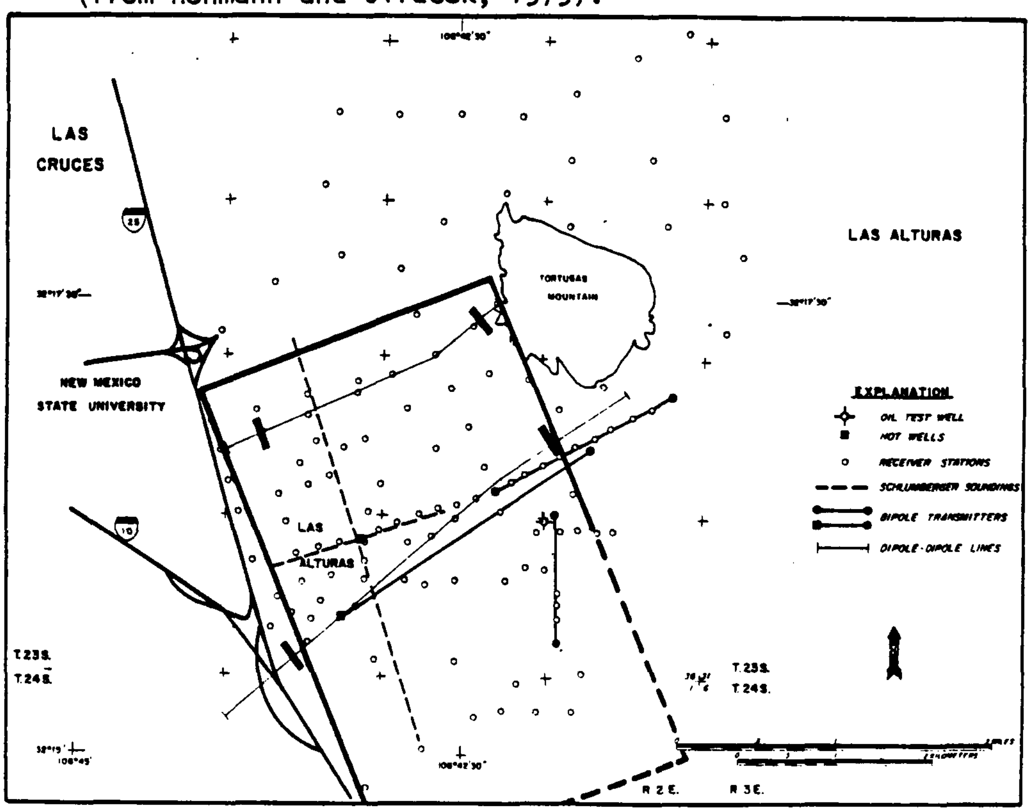


Figure 16. Outline (heavy solid and dashed lines) of conductive hydrothermal aquifer at Las Alturas Estates.

Interpretation and Drilling Results. The major resistivity contrast detected by the dipole-dipole results in Figure 11 occurs in the vicinity of station 42 where a highly resistive (300 ohm-m) block terminates the conductive aquifer. This is the same location where the western boundary of the high apparent resistivity lobe surrounding Tortugas Mountain forms a distinct, sharp, and linear gradient (Figure 13). These results and the combined modeling suggest a fault zone west of Tortugas Mountain with an extension to the south. This geologically unmapped fault zone may govern the occurrence of hot water at Las Alturas Estates as hypothesized in Figure 17. This section would be appropriate to a profile running approximately along the northernmost dipole-dipole line (Figure 8).

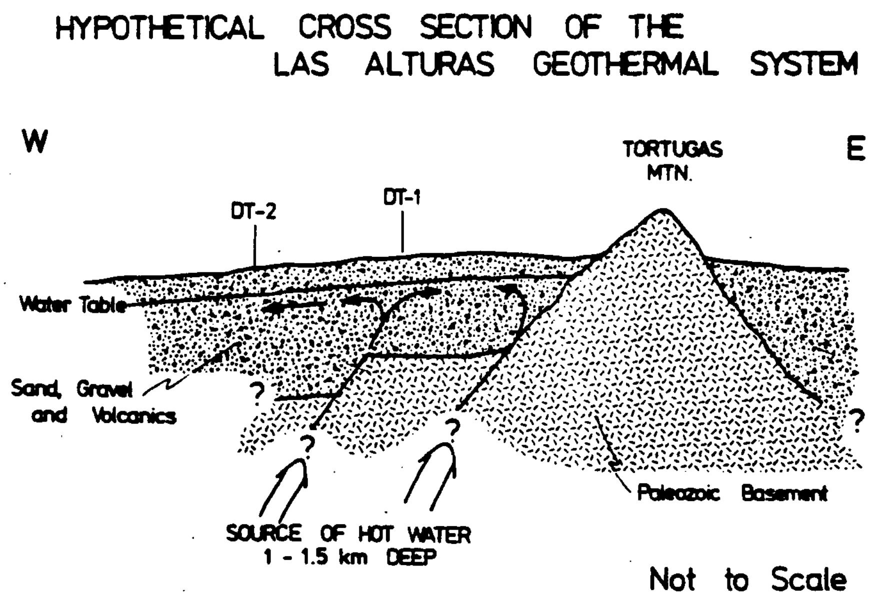


Figure 17. (after Morgan et al., 1980).

A consequence of the hypothetical cross section in Figure 17 is the expectation of higher shallow temperatures on the eastern side of the survey area. This original suggestion (Jiracek and Gerety, 1978) has been proven correct as is evident by the two temperature gradient curves in Figure 18. Locations of the two gradient wells are shown in Figure 17. It is significant to add that test well DT1 is located over the region of lowest modeled dipole-dipole resistivity (5 ohm-m).

A production well near DT1 is currently being used to heat a new home for the New Mexico State University president and plans are being formalized for more extensive use on campus.

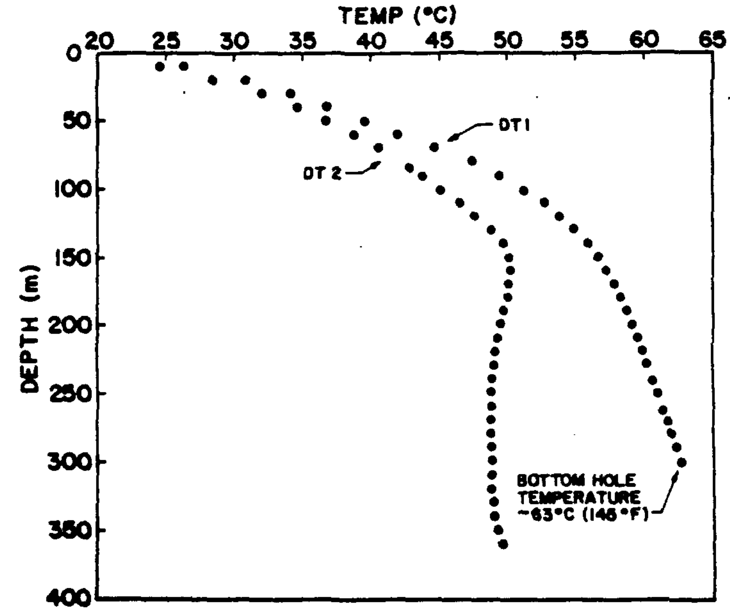


Figure 18. Temperature data from test wells DT1 and DT2 at Las Alturas Estates (from Morgan et al., 1979).

Acknowledgments

The Las Alturas Estates field activities were completed with the able assistance and leadership of many students from the University of New Mexico and New Mexico State University. The evaluation program was supported by New Mexico Energy Resources Development Program Grants No. 22 and No. 117-69-B and by U.S. Geological Survey Grant No. 14-08-0001-E-255.

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REGIONAL ASSESSMENT FOR HOT DRY ROCK RESOURCES

A. W. Laughlin and M. J. Aldrich

The Hot Dry Rock (HDR) Geothermal Energy Development Program began about eight years ago at the Los Alamos National Laboratory. It was conceived as a means of extracting geothermal energy from high temperature reservoir rocks with very low natural permeability. The basic concept involves drilling to a depth where temperatures are economically attractive, generating a large heat exchange surface by hydraulic fracturing, and finally drilling into the fractured zone with a second hole to form a closed-loop circulating system. Cold water injected down one borehole is heated as it flows along the fracture or through the fracture zone and then is returned to the surface via the second borehole.

Three types of geologic studies form parts of the HDR program. The first of these is directly concerned with the development of the prototype system at Fenton Hill, New Mexico. It is primarily concerned with providing direct geologic support to the drilling, fracturing, and reservoir formation operation at Fenton Hill. Subjects such as rock permeability, composition, fracture spacing, and orientation are addressed. In addition, as funding is available, exploration and assessment techniques are being tested around the Fenton Hill site. It is hoped that eventually a detailed case history will be developed that will allow comparison of the various geothermal exploration techniques.

The second group of geologic investigations is directed towards the selection of a second site for the testing of the HDR method of geothermal energy extraction. The Fenton Hill site is representative of one type of high grade HDR site, i.e., a site associated with a young silicic volcanic center. If the HDR concept is ever to be widely used it must be thoroughly evaluated in other geologic environments. A number of possible sites are currently being examined. These range from sites in the eastern U.S. best suited for direct utilization to sites in the western U.S. associated with both high regional heat flow and young volcanism.

The third group of geologic investigations are the regional evaluations of the HDR resource base. The ultimate objective of these studies is the assessment of the HDR resource base of the entire United States. It is the results of these studies that are the most relevant to the Low Temperature Program.

The first step in exploration for any type of geothermal resource is the identification of a heat source. This holds true for high and low temperature hydrothermal, geopressured, and HDR resources. Clearly, without the heat nothing else matters. Evaluation of permeability is relegated to the second phase of the assessment activity.

To assess the HDR resource base of the United States, the country has been divided into broad regions with similar geologic settings. One or two persons at Los Alamos are responsible for the evaluation of each of these regions. Exploration and assessment work in these areas is being done by Los Alamos staff and academic and industrial subcontractors.

Arizona, New Mexico, and the Trans-Pecos area of west Texas comprise one of these regions. To help perform an assessment of this region, we are preparing a series of eleven maps at a scale of 1:1,000,000. The first two of the maps in this series evaluate direct evidence for regional and local (igneous point sources) heat sources. In press is an update of the volcanic rock map of Luedke and Smith (1978). Our version incorporates data published since 1978. Because rocks older than 3 m.y. have probably lost their original magmatic heat, we have restricted our map to a delineation of rocks younger than 3 m.y. A separate report, also in press, tabulates the location and age of volcanic rocks younger than 5 m.y. for the region. A heat flow map of this region is also being constructed using all available data. This map will rely heavily on work by the USGS, Decker and his colleagues, and Reiter and his students.

Indirect evidence for possible elevated temperatures in the crust will be presented on two planned maps; one, a map showing the location of deep electrical conductors in the crust and the second, a depth to Curie point map. The electrical conductor map will be mainly a product of our own extensive, contracted MT work in Arizona and New Mexico. Because of the correlation between depth to electrical conductor and heat flow, it is hoped that this technique will eventually prove effective as a reconnaissance tool.

We anticipate eventually synthesizing the aeromagnetic data for the Arizona-New Mexico-Trans-Pecos area of Texas and compile a map for the region. This aeromagnetic map will then serve as a data source for generating a depth to Curie point map. Preliminary correlations in northwest Arizona between depth to Curie point, geothermal gradient, P-wave delay, seismic wave attenuation, and gravity suggest that the Curie point depth may also be useful as a reconnaissance technique.

Using all of the direct and indirect evidence we plan to develop a map showing depths to the 100°C and 150°C isotherms. Obviously this map will be of use to any geothermal explorationist.

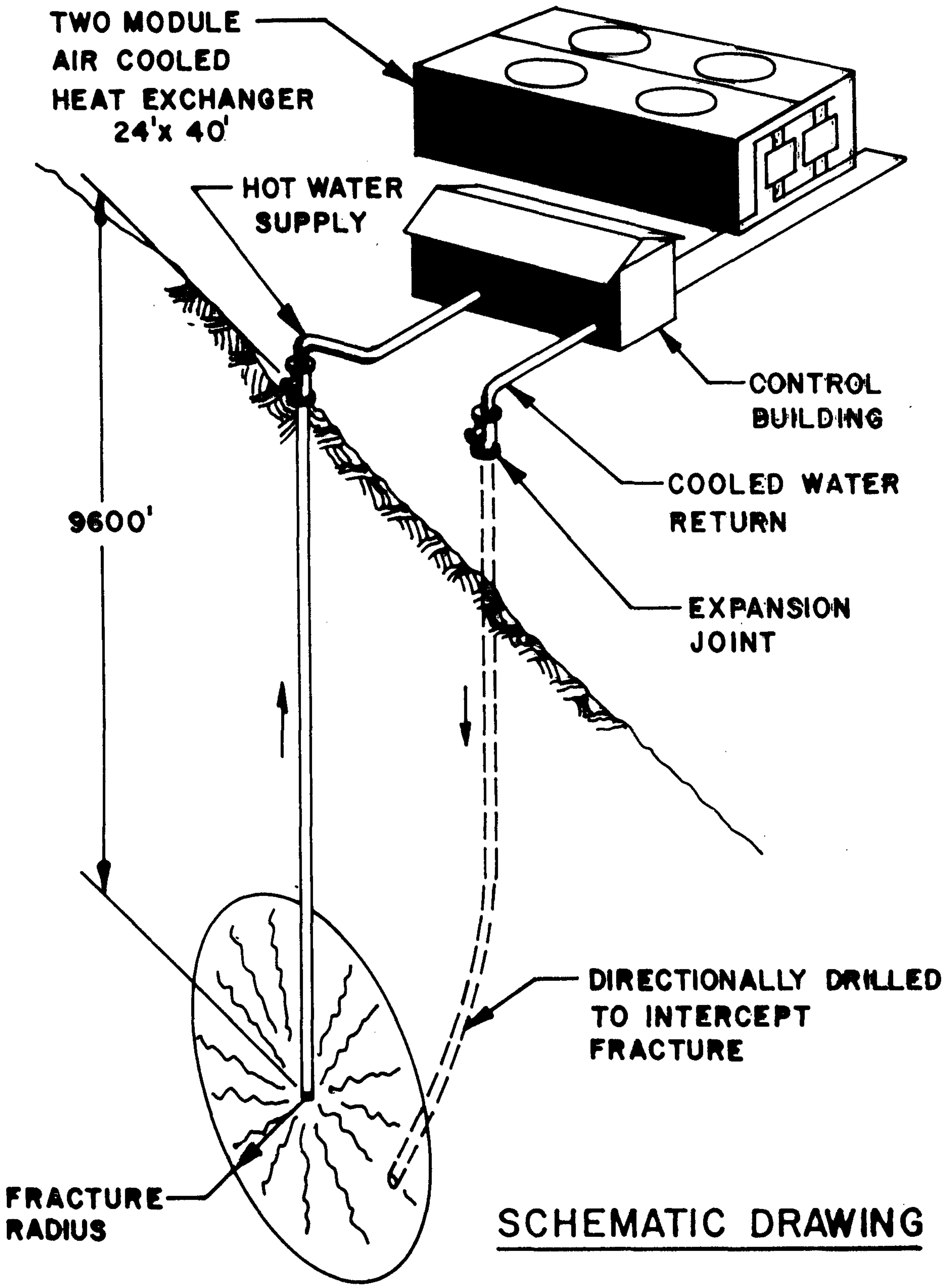
The remaining maps in the series illustrate the stress-orientation, depth to potential reservoir rocks, seismicity, gravity, potential users, and the utility grid for this region.

The stress-orientation map builds on the recent map by Zoback and Zoback (1980) of the stress orientation throughout the United States. We have acquired considerably more data for the southwestern U.S. This map is being drafted at the present time and should be published this summer.

Work has begun on the depth to reservoir rock map and it is perhaps 30% finished. Hopefully it should be published in the fall.

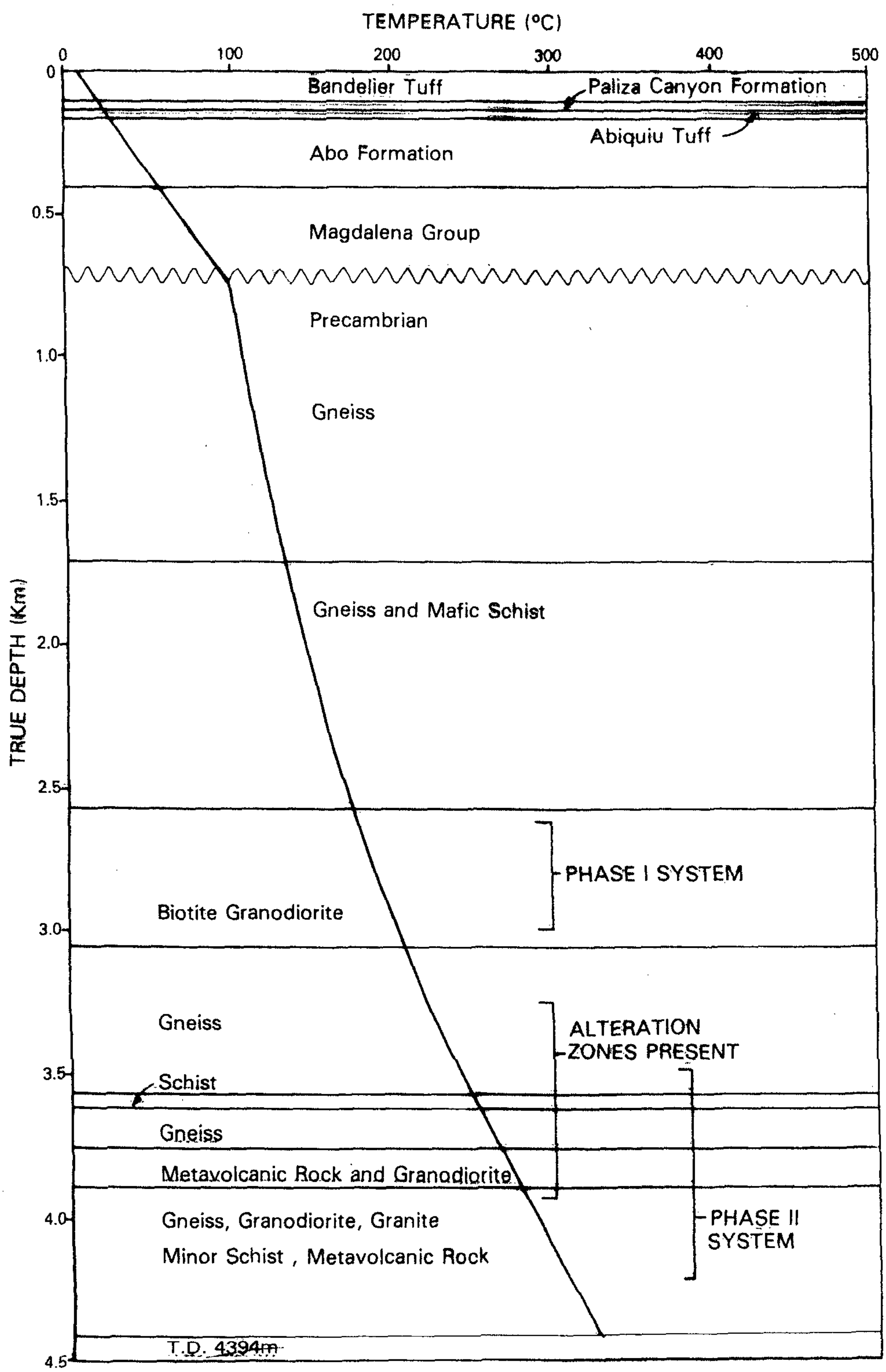
Work on the remaining maps should begin during FY-82.

20 MW (THERMAL) DRY HOT ROCK ENERGY SOURCE DEMONSTRATION

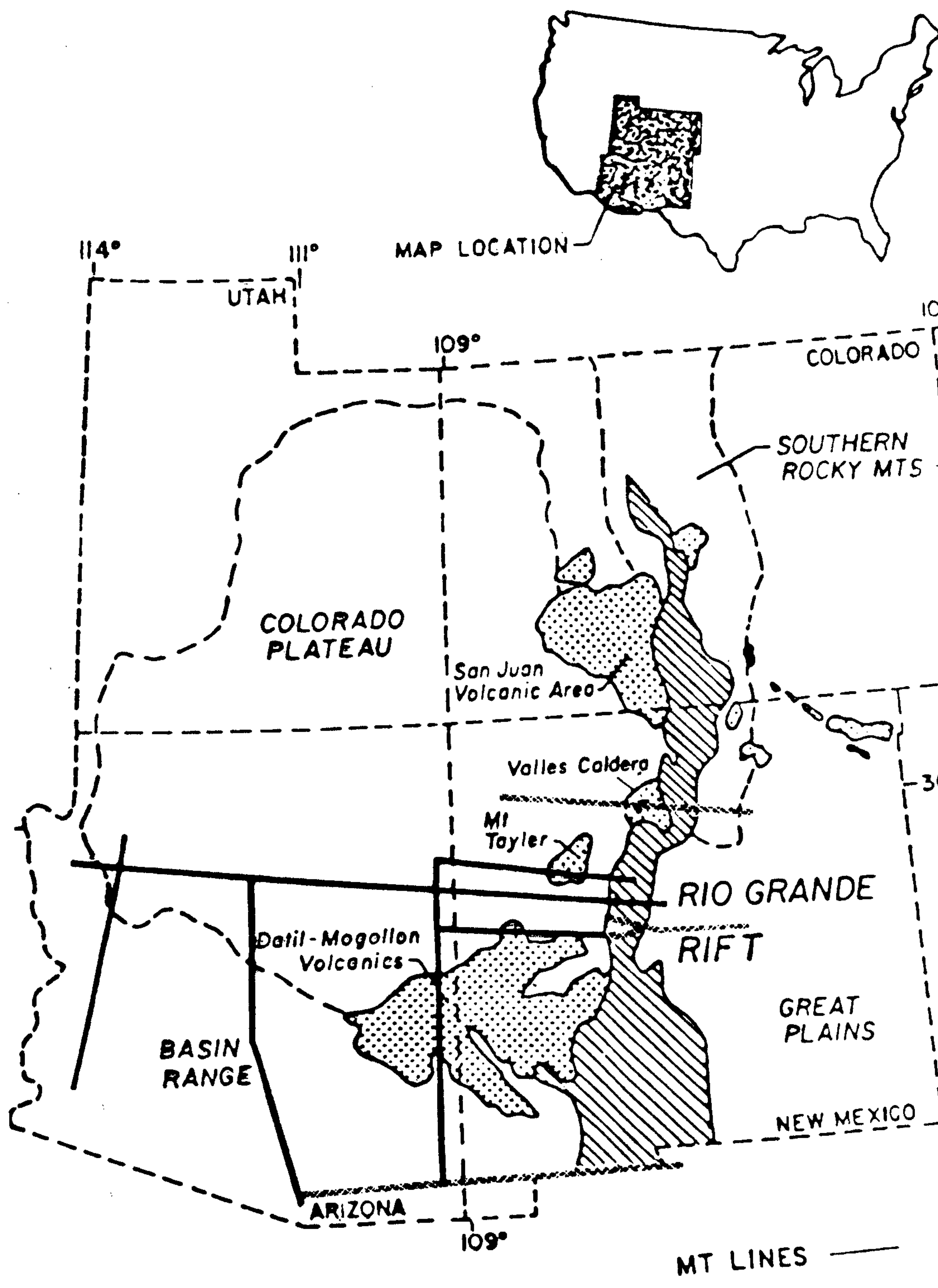


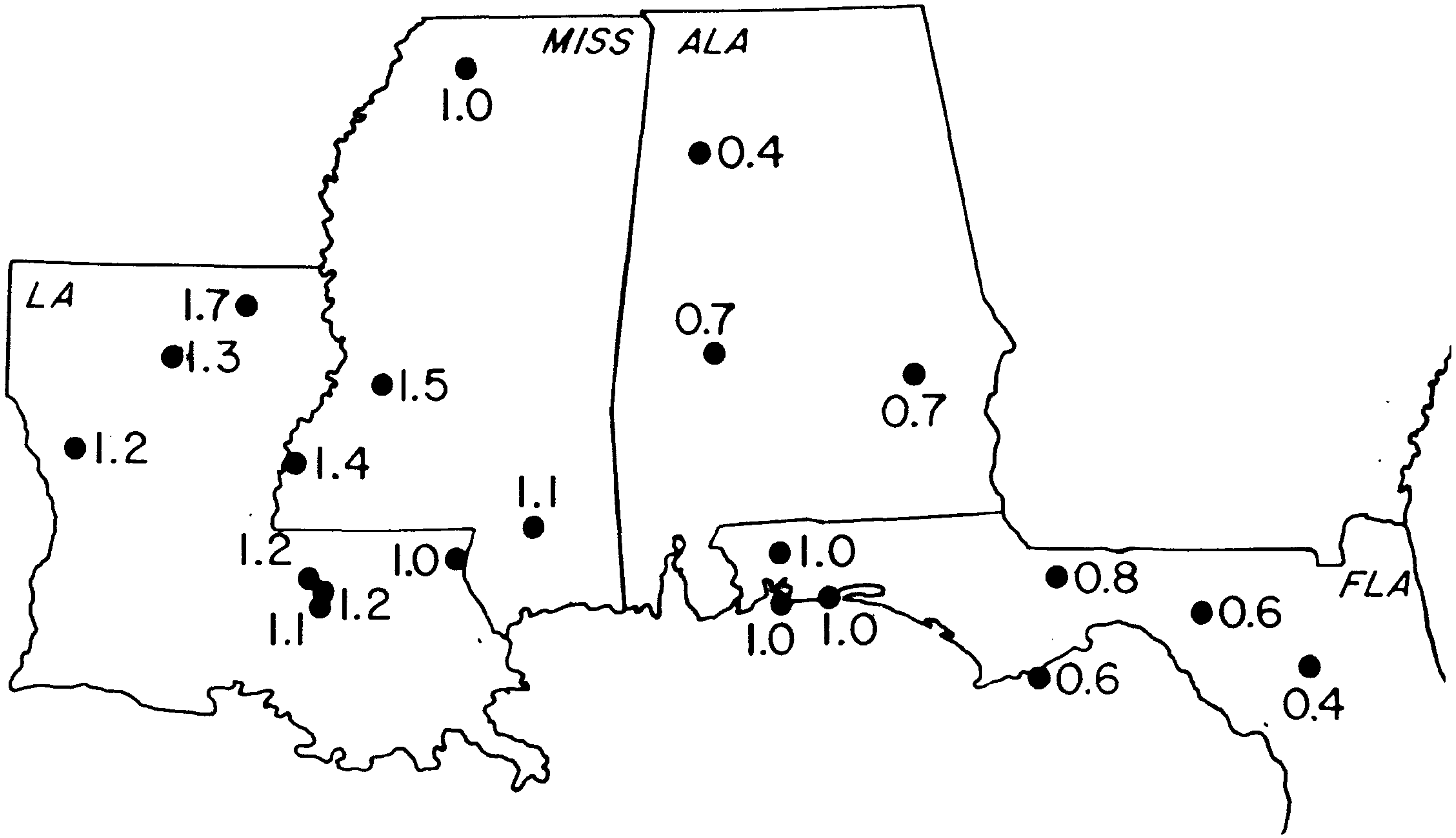
SCHEMATIC DRAWING

GEOLOGIC CROSS SECTION, FENTON HILL HDR SITE NEW MEXICO



GRADIENT SHOWN IS DERIVED FROM MEASUREMENTS IN GT-2, EE-1 AND EE-2

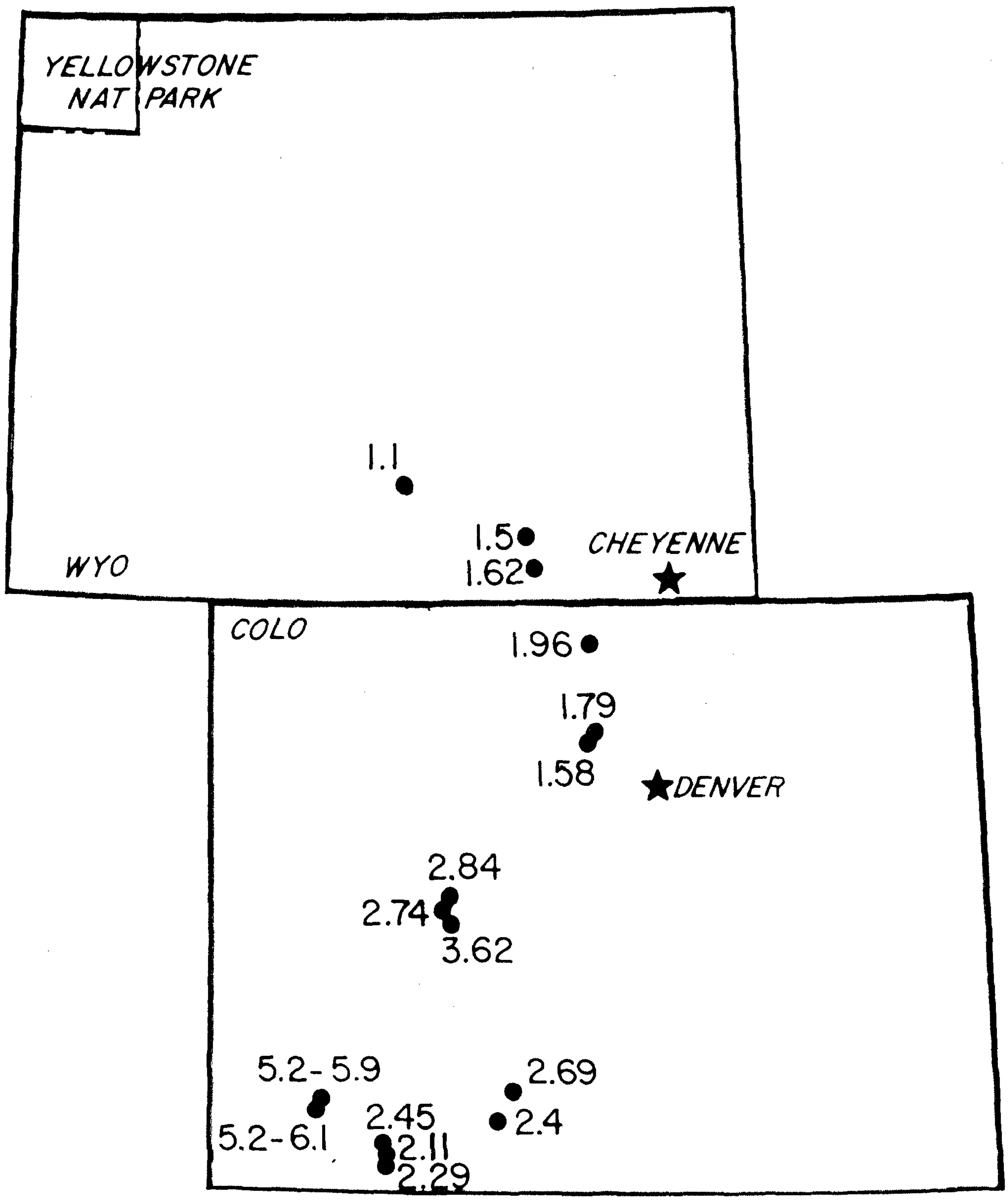

MAP LOCATION
114°
111°
UTAH
109°
102°
41°
COLORADO
SOUTHERN
ROCKY MTS
COLORADO
PLATEAU
San Juan
Volcanic Area
37°
Valles Caldera
36°
Mt
Tayler
RIO GRANDE
RIFT
Datil-Mogollon
Volcanics
GREAT
PLAINS
BASIN
RANGE
NEW MEXICO
32°
103°
ARIZONA
109°
MT LINES



NEW HEAT FLOW VALUES IN S.E. U.S.

MEASURED by D.L. SMITH, UNIV. of FLORIDA

FOR LASL



NEW HEAT FLOW VALUES IN W. U.S.

MEASURED by E.R. DECKER FOR LASL

Problems of Trace Element Ratios and Geothermometry in a Gravel Geothermal-Aquifer System

Sonderegger, J. L., and Donovan, J. J.

Introduction

The utilization of trace element concentrations and ratios of trace-elements to each other or to chloride in an attempt to understand the chemical evolution of a particular water was first applied to brines, seawater, and mineral springs (Rankama and Sahama, 1950, Chap. 6; Rubey, 1951; White, D. E., 1957). Tonani's (1970) overview paper on geochemical methods applied to geothermal exploration emphasized vapor phase separation of the volatiles hydrogen sulfide, ammonia, and boric acid; figure 1, from a more recent manuscript (Tonani, 1980), shows the fractionation of boron between liquid and vapor phases as a function of temperature. Attempting to utilize B/Cl, Li/Cl, and B/Li ratios, without consideration of total concentration levels can result in misleading interpretations of relatively shallow thermal systems, which undergo a significant amount of dilution.

The system studied is a Tertiary-age, block-faulted basin in which a Pleistocene gravel bed acts as a confined aquifer and permits the lateral dispersion of the geothermal fluids. Vertical movement of the hot water is currently believed to be controlled by faults on the east side of the valley. An aerial magnetic anomaly (Geodata International, 1981, lines 300 and 320) and a Bouguer gravity anomaly (Donovan, et al., 1980) appear to correspond with these eastern faults. Basic data on the geology and trace element halos has been presented previously (Donovan, et al., 1980); figures 3, 4, and 5 from that article are reproduced below, along with the fluoride data (figure 2).

Evaluation of the mixing phenomena in this system was attempted using a dissolved silica-enthalpy graph. Figure 6 differs from most figures of this type in that a chalcedony curve is also plotted. Figure 7, an enthalpy versus chloride plot, suggests that either conductive cooling occurs before mixing or that higher chloride content background waters are available for mixing.

Ion Ratios from Various Sources

First let us consider the results of short-term leaching of volcanic rocks. The data used to construct Table 1 are from Ellis and Mahon (1977); the temperature was maintained at 250°C for two weeks, a rock/water ratio of 2 was used for all but the rhyolite pumice. For the pumice, a rock/water ratio of 1 was used by the experimenter, and in the absence of a 250°C experimental run, the values from the 200 and 300°C have been averaged. Values for F/Cl, B/Cl, and Li/B for the existing Camp Aqua well were 0.120, 0.020, and 0.136, respectively; these ratios suggested that the thermal potential might be greater than calculated, particularly if the chalcedony curve in figure 6 is used to estimate the reservoir temperature.

Table 2 compares ranges for ion ratios; the first three rows are from White (1957) and the second three rows are from Montana systems. At Ennis, a Na-HCO_3-SO_4 water with 120 mg/kg Cl and a surface temperature of 83°C, yields a 153°C Na-K-Ca-Mg reservoir temperature and a 135°C quartz reservoir temperature, assuming no mixing. A 540 foot well encountered 93°C water. At Warm Springs the surface temperature (77-78°C) agrees with the Na-K-Ca calculated 79.5°C, and chalcedony calculated value of 77.7°C. A 1,500 foot drill hole at Warm Springs encountered 78°C water. The Ennis system is in gneiss, while the Warm Springs system is in Tertiary(?) age valley fill materials and is probably fed by a Paleozoic limestone (high Ca and SO_4 content). The Camp Aqua data generally fit in the same order of magnitude for trace element ratios; however, the total dissolved solids content is less than half as great.

The test well in the Camp Aqua vicinity was drilled with state funds to evaluate a request for funding for a deep, large-diameter production well by a private concern. Table 3 contains chemical data and geothermometer calculations for well 211, (believed to represent the background water in the gravel aquifer), the Camp Aqua well, the test well in the gravel zone, and the test well in the bedrock zone. These data suggest to us that the 124°C source water projected from figure 6 may exist somewhere in the system and that dilution has had a greater effect on the quartz geothermometer than on the Na-K-Ca-Mg geothermometer.

A note of warning should be stressed at this point. Secondary calcite, and possibly dolomite, have been noted in the cuttings, both from the gravel and bedrock zones. Thus the correlation between the Na-K-Ca-Mg

Table 1. Ratios by weight of leach solutions

	F/Cl	B/Cl	Li/B
Basalt	0.024	0.006	0.214
Andesite	0.192	0.048	0.200
Dacite	0.375	0.048	0.368
Rhyolite Pumice	0.438	0.006	---

--- insufficient data, value @ 200°C is 3.33

Table 2. Approximate ranges of ion ratios from chloride dominated waters (From White, 1957) and data from the Camp Aqua well and two springs

	F/Cl	B/Cl	Li/Na	As(ppm)	H_2S(ppm)	Li/B
Ocean	7×10^{-5}	2.4×10^{-4}	10^{-5}	2×10^{-3}–1.7×10^{-2}	0–60	N.L.
Oil-field Brines	10^{-5}–10^{-3}	10^{-5}–2×10^{-2}	10^{-4}–3×10^{-3}	?	0–3,000	N.L.
Volcanic Hot Springs	5×10^{-4}–10^{-1}	10^{-2}–10^{-1}	3×10^{-3}–3×10^{-2}	10^{-1}–10^{-1}	0–10	N.L.
Camp Aqua	1.2×10^{-1}	2×10^{-2}	5.7×10^{-4}	---*	---	0.136
Ennis	9.2×10^{-2}	5×10^{-3}	7.6×10^{-4}	2.5×10^{-2}	---	0.426
Warm Springs	7.8×10^{-1}	2×10^{-2}	3×10^{-3}	---	0.7	3.60

*halo from lower temperature wells up to 0.1 ppm

N.L. - Not Listed

--- - Data not available

Table 3. Chemical data from Camas area wells

	Ca	Mg	Na	K	SiO_2	Field Alka.	Cl	SO_4
Well LB-211	29.4	6.6	24.0	0.8	20.8	125	1.4	23.8
Well LB-32	3.2	0.3	152.	4.0	42.2	293	32.5	4.1
Test Well 264'	3.4	0.3	159.	3.2	45.9	297*	35.8	0.4
Test Well 324'	10.7	2.1	139.	2.9	38.8	328	35.9	<.1

	F	Li	B	As	H_2S	Temp ^oC	TDS	Field pH
Well LB-211	0.4	<.008	<.09	0.009	<.1	10	179	8.26
Well LB-32	3.9	0.087	0.64	<.0001	--	52	437	8.71*
Test Well 264'	5.2	0.083	0.64	0.0005	--	49	432	8.72*
Test Well 324'	4.6	0.050	0.63	<.0001	--	47	406	7.82

	Chalcedony	Quartz	Na-K-Ca-Mg
Well LB-211	33^o	65^o	-58^o
Well LB-32	66^o	94^o	117^o
Test Well 264'	68^o	98^o	119^o
Test Well 324'	60^o	90^o	79^o

*Laboratory value

temperatures and the 124°C projection from figure 6 may be wholly fortuitous. It is our belief, at this time, that only the chalcedony curve in figure 6 can be used with any confidence, and that the 77°C source water is the more probable of the two. The trace element data ratios are probably, in part, controlled by fines (clays) within the gravel due mainly to ion exchange. At least one zeolite (either heulandite or clinoptilolite) has been detected in the bedrock cuttings (along with pyrite and chalcopyrite(?)). We are definitely looking at an old hydrothermal system. What is uncertain at the present time is whether the hydrothermal alteration and mineralization are related synoptically to the current hot water system, or whether the modern goethermal circulation systems simply following the same flow pathway followed by older (possibly Tertiary) hydrothermal fluids and intrusives.

Conclusions

The valley fill aquifer system in the vicinity of Camp Aqua is still poorly understood with respect to the temperature of source waters. For a low TDS (420 mg/kg) Na-HCO_3 water in valley fill materials it is advised that trace element ratios be used with great caution, and the absolute values of the trace elements and chloride must be considered. The role of cation (Li) and base (B) exchange by clays in a system affected by waning or prior hydrothermal alteration is not presently clear, and needs further study.

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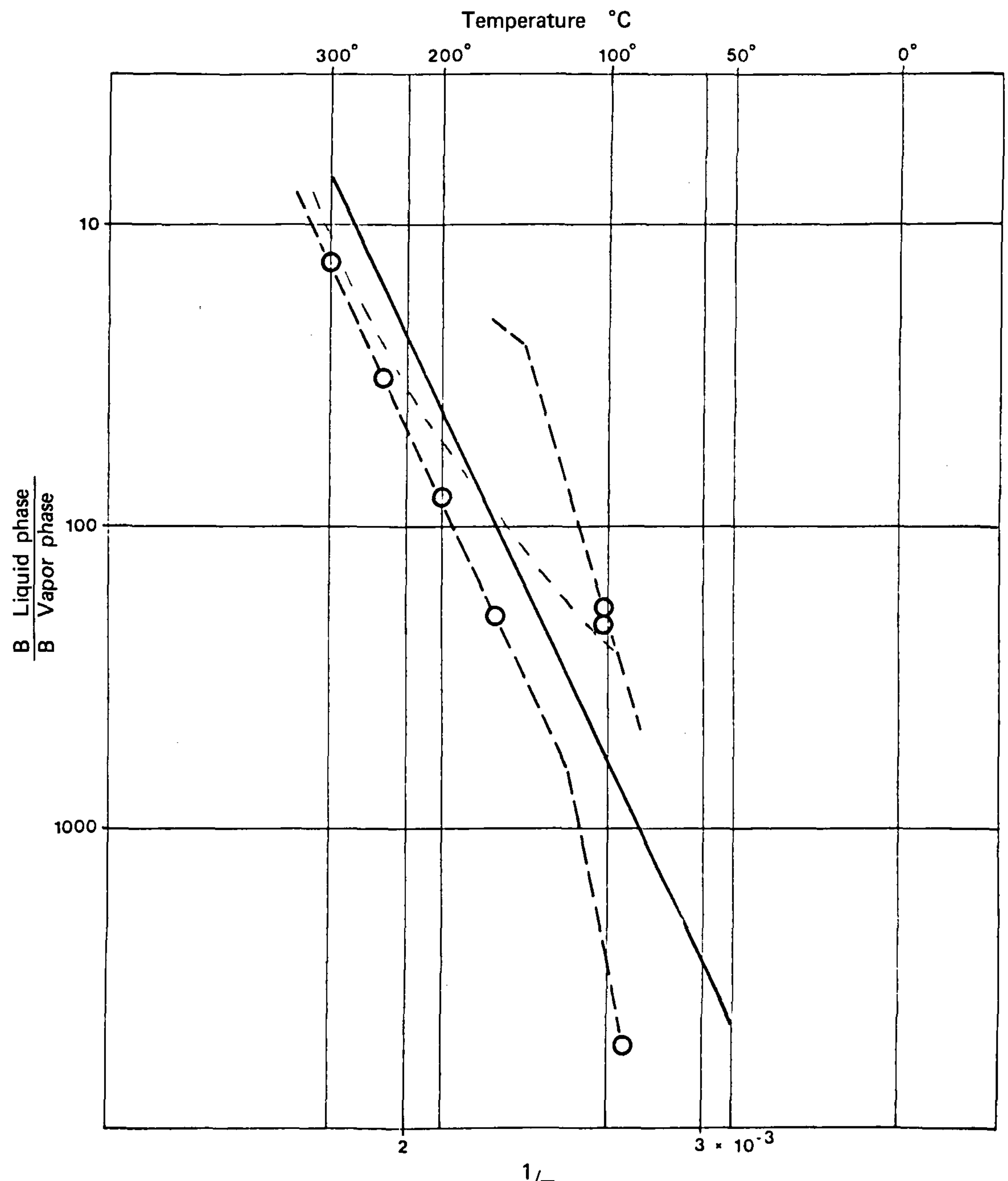

Figure 1. Plot of boron fractionation between the liquid and vapor phases as a function of temperature (from Tonani, 1980).

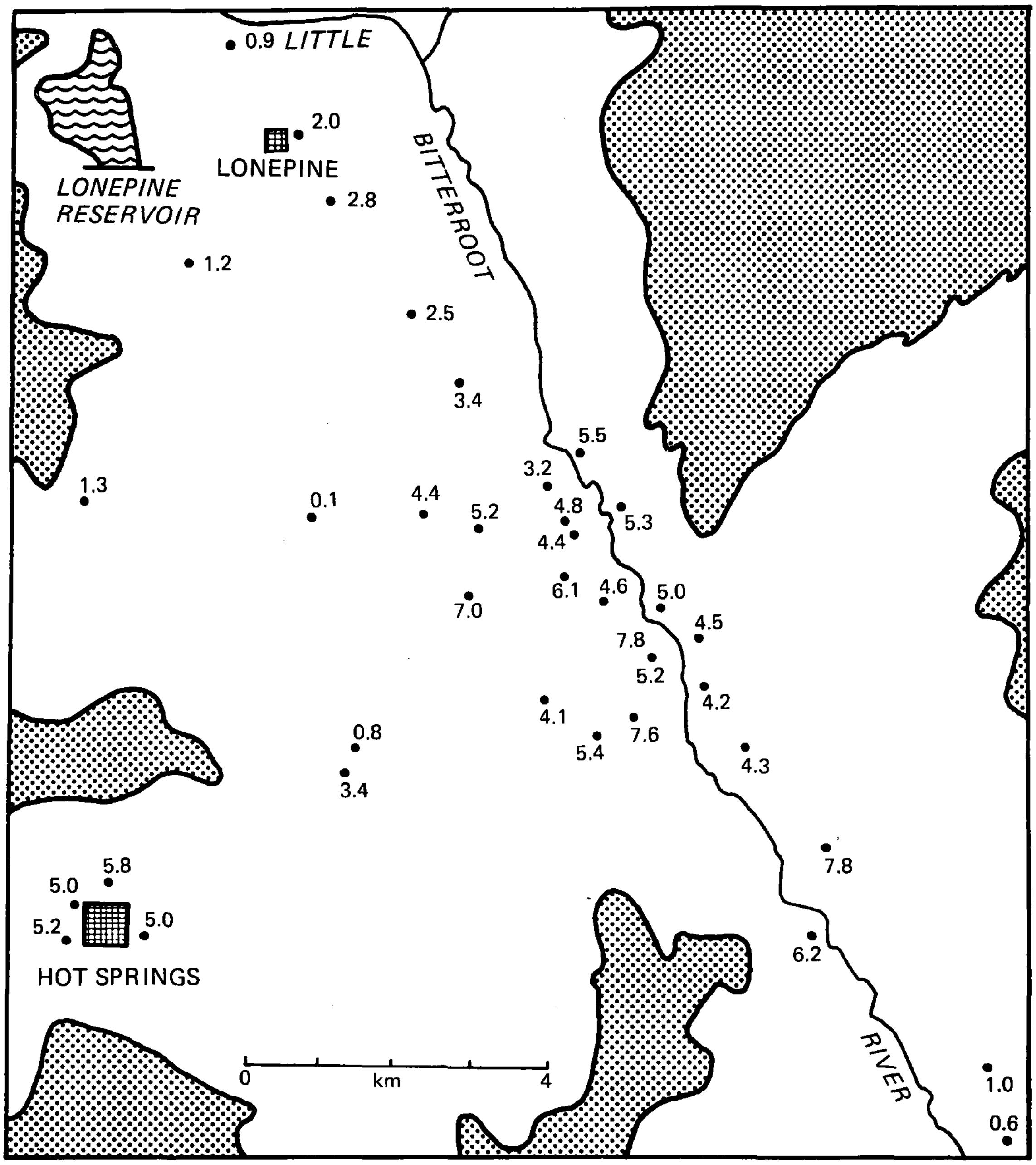

Figure 2. F^- concentrations (mg/kg) in well waters.

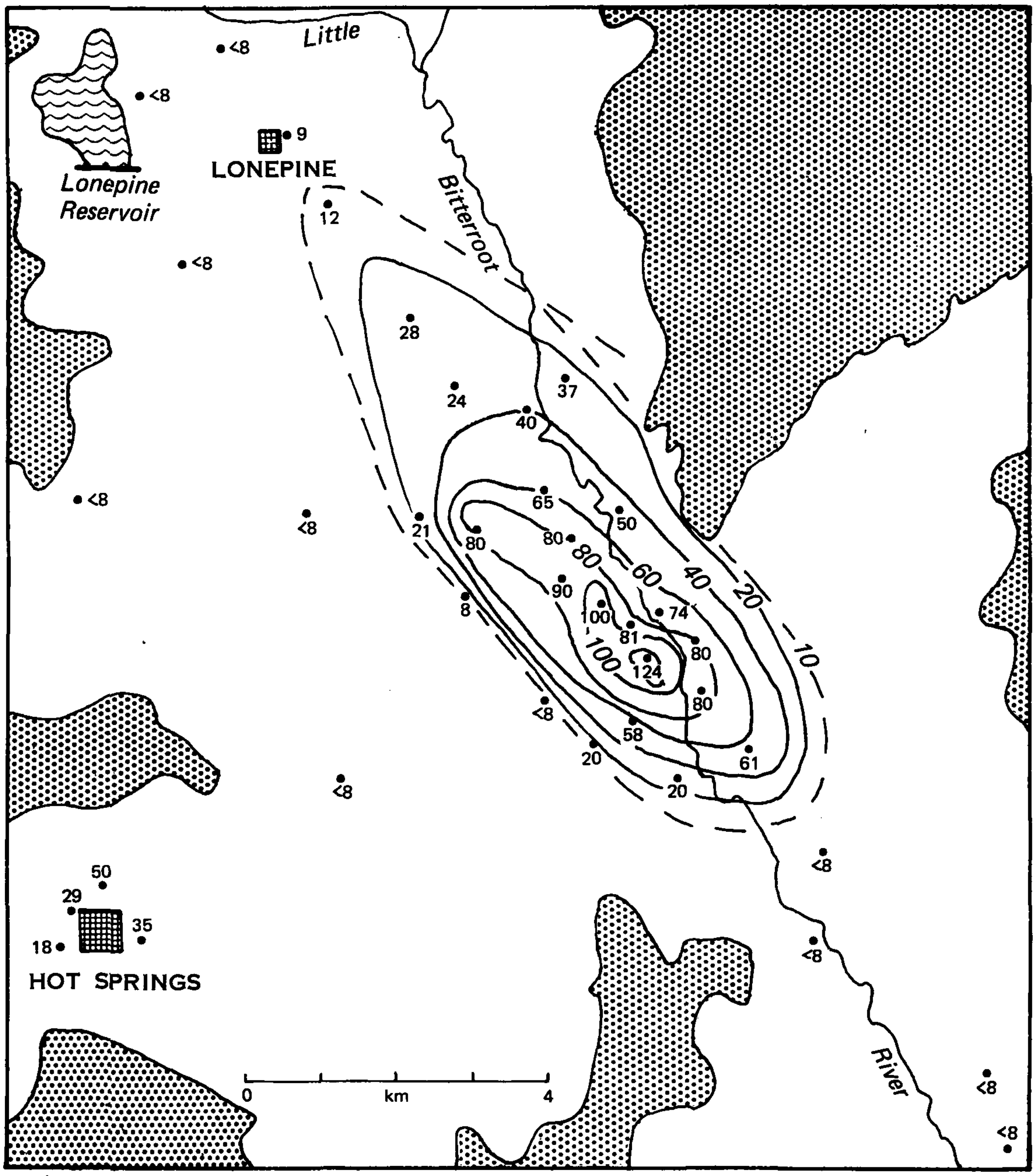

Figure 3. Li^{+} concentrations (μg/kg) in well waters.

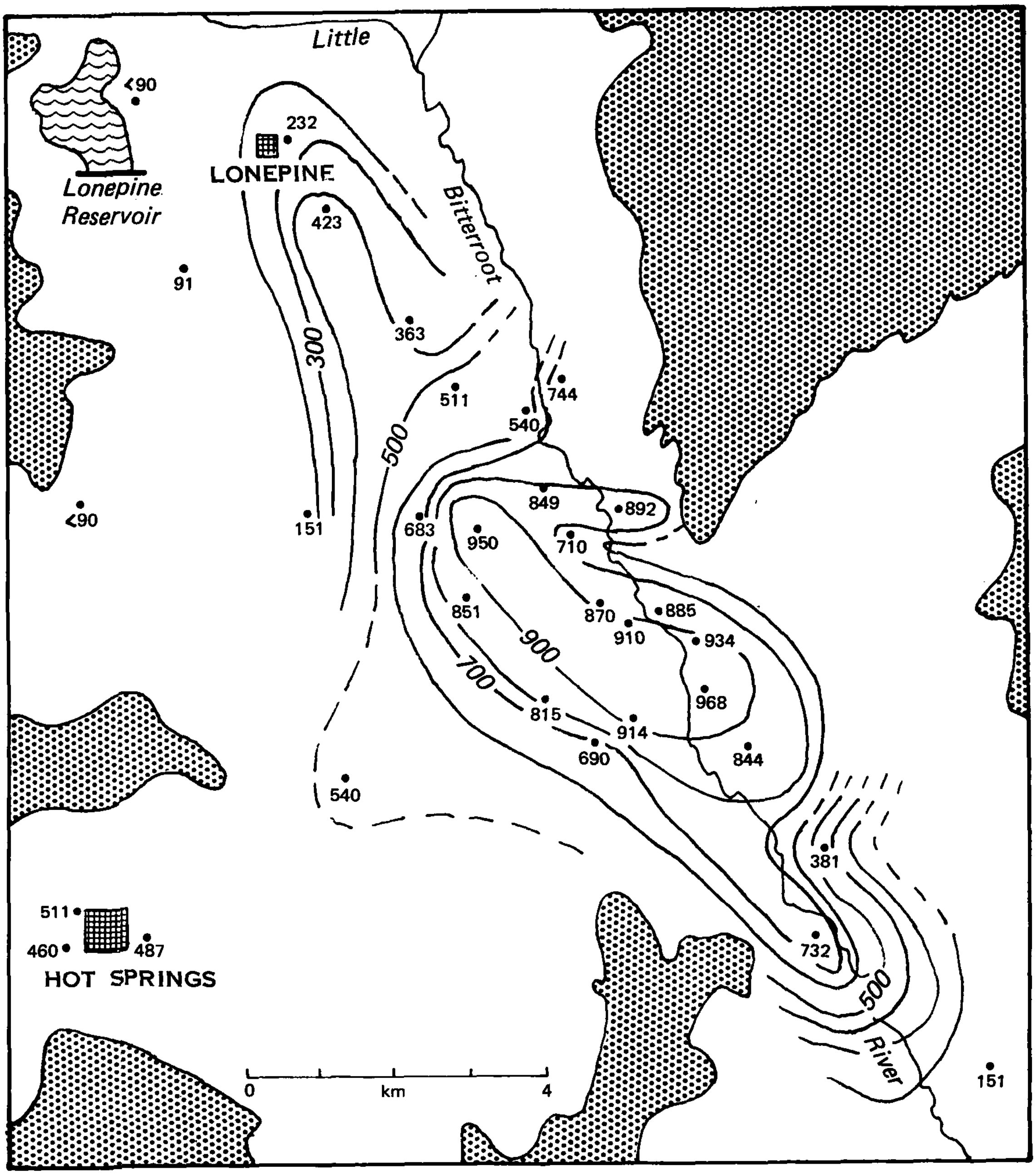

Figure 4. B concentrations (μg/kg) in well waters.

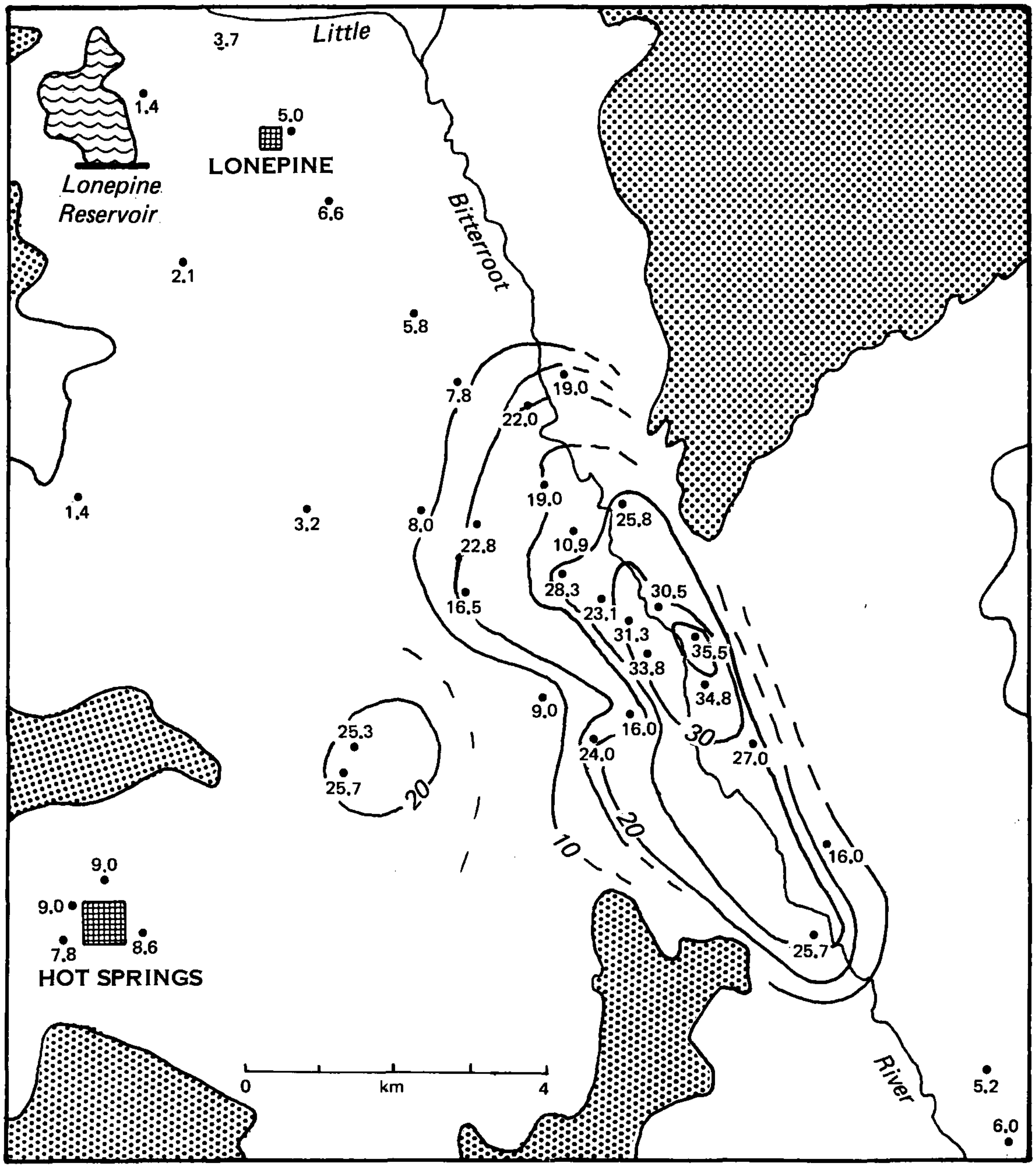

Figure 5. Cl^- concentrations (mg/kg) in well waters.

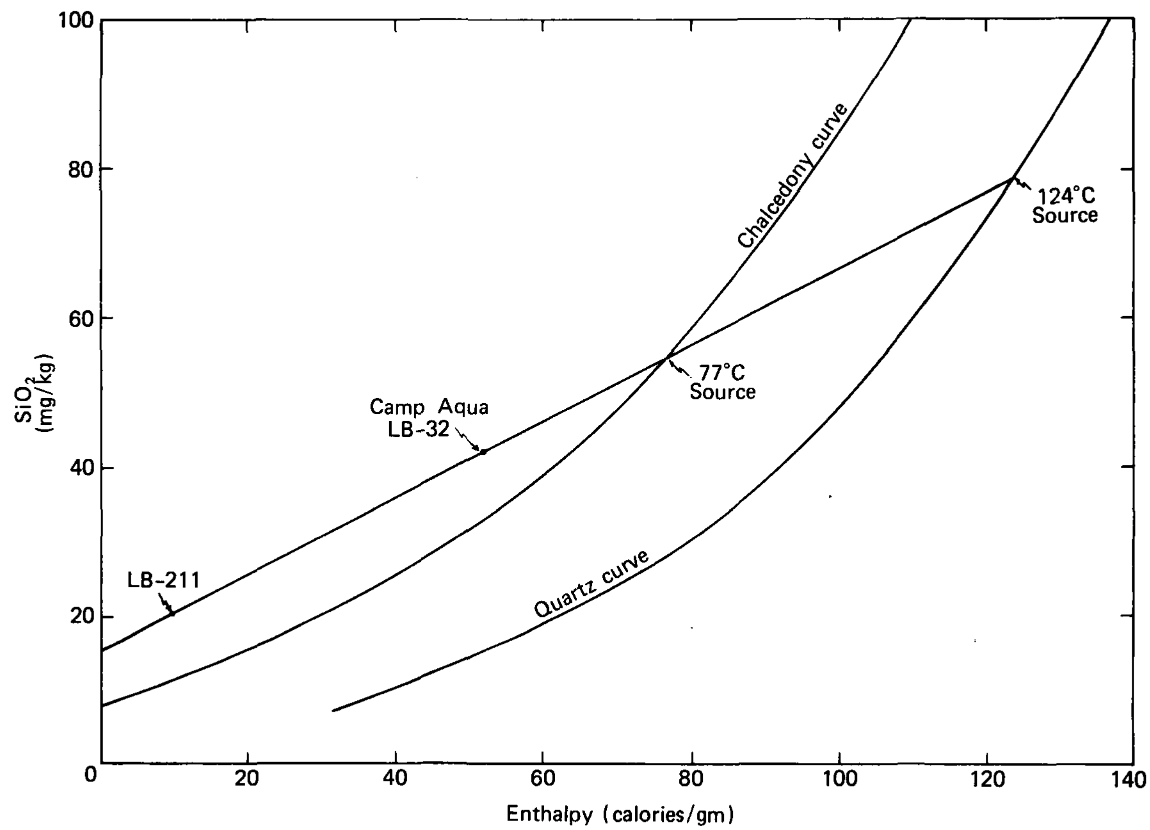

Figure 6. Enthalpy - SiO_2 plot of data from a typical background well (211) and the Camp Aqua well. The chalcedony curve has been added using data from Fournier (1980).

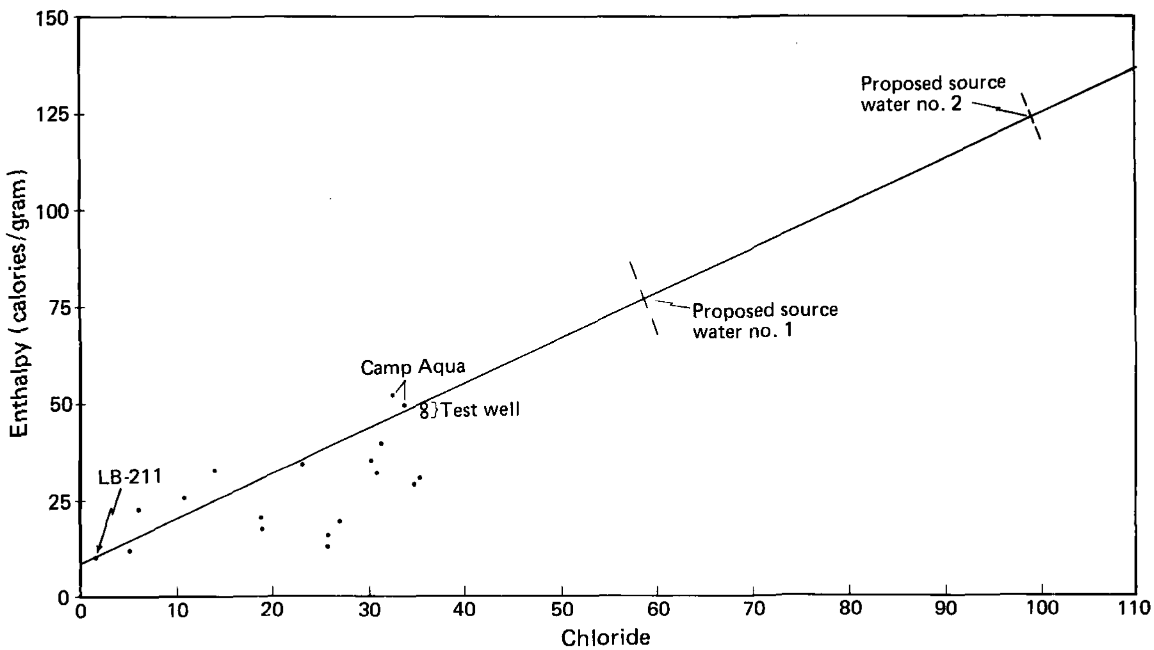

Figure 7. An enthalpy-chloride plot for waters from the gravel aquifer in the Camp Aqua vicinity.

HELIUM AND MERCURY IN THE CENTRAL SEWARD PENINSULA RIFT SYSTEM, ALASKA

by

Eugene Wescott
Geophysical Institute
University of Alaska
Fairbanks, Alaska 99701

Abstract

The central Seward Peninsula, Alaska, has one Known Geothermal Resource Area (KGRA) at Pilgrim Springs, and has recent volcanic flows, fault systems, topographic and tectonic features which can be explained by a rift model. As part of a geothermal reconnaissance of the area we used helium and mercury concentrations in soil as indicators of geothermal resources. The largest helium concentrations were found in the vicinity of the Pilgrims Springs KGRA, and indicate prime drilling sites. Five profile lines were run across the suspected rift system. Significant helium anomalies were found on several of the traverses, where future exploration might be concentrated. Mercury values showed a great range of variability on the traverses, and seem unreliable as geothermal indicators except in the vicinity of the Pilgrim Springs. Permafrost at the surface resulting in variations in sampling depth may contribute to the mercury variations.

Introduction

Turner et al., 1981, have proposed "that an interconnected system of late Tertiary to Quaternary rifts and transform faults extends 250 km across the central Seward Peninsula from Port Clarence to the eastern Koyuk River Valley". The rift model appears to explain many late Tertiary to Quaternary topographic, structural, tectonic and volcanic features, and should be useful as an exploration model for geothermal energy resources. Figure 1 shows a diagram of the proposed rift system on a generalized geological map.

As part of a study of the Seward Peninsula geothermal energy potential during summer 1980, geological mapping, remote sensing, geophysical exploration and geochemical sampling methods were used to develop the rift model and search for geothermal resources. Previous work in 1979 at the Pilgrim Springs KGRA had delineated the near surface reservoir, which shallow drilling confirmed.

The Seward Peninsula covers approximately the same area as the state of West Virginia, but with very few roads. In order to search for evidence of other hidden geothermal resources in this large area we tried using helium and mercury in soil samples as indicators. Both have been reported as useful in exploration for geothermal areas and they can be sampled fairly rapidly and are inexpensive to analyse.

Helium anomalies near geothermal sites have been detected near geothermal sites throughout the world (Bergquist, 1979). Two factors may contribute to helium anomalies in conjunction with geothermal areas: The deep, nonatmospheric mixing source of most geothermal waters, and the radioactive decay of uranium and thorium in the vicinity of the source waters. Helium is unusual in that its solubility in water increases with temperature above 30°C [Figure 2 after Mazor (1972)]. Pressurized hot water will be a very efficient scavenger of helium produced by radioactive decay of uranium and thorium contained in the rocks at depth, and will release it as it rises towards the surface, cools and de-pressurizes. Since helium is highly mobile it will find faults, minute fractures and paths to rise to the surface. Some helium will be trapped in the rocks and sediments, but because its atomic structure is nearly spherical the entrapment is difficult (Bergquist, 1979).

We sampled the soil for helium in two ways: The first was to drive a probe 30 in. into the ground and draw off a ground gas sample which was then inserted into a small evacuated steel ampule and sealed for later analysis. This method does not work well in wet soil or where the soil is rocky or frozen. In such conditions we used a soil sampling auger to drill a hole 30 in deep. The soil core at the bottom was then quickly placed in a tin can and sealed. Western Systems, Inc., performed the helium analysis to a precision of 10 parts per billion. Normal atmospheric He concentration is 5.24 ppm, and any significant soil concentration above this is an anomaly.

Mercury content in soils has also been reported as a possible indicator of geothermal resources (Matlick and Buseck, 1975). They confirmed a strong association of Hg with geothermal activity in three of four areas tested (Long Valley, California; Summer Lake and Klamath Falls, Oregon). Mercury deposits typically occur in regions containing evidence of geothermal activity, such as hot springs (White, 1967).

Mercury is very volatile. The high vapor pressure makes it extremely mobile, and the elevated temperatures near a geothermal reservoir tend to increase this mobility. The Hg migrates upwards and outwards away from the geothermal reservoir, creating an aureole of enriched Hg in the soil above a geothermal reservoir larger in area than a corresponding helium anomaly.

We collected soil samples about 10 cm below the organic layer. The samples were air dried in the shade and sized to -80 fine using a stainless steel sieve. The -80 portions were stored in airtight glass vials for analysis. The Hg content of the sample was determined by use of a Jerome Instrument Corp., model 301 Gold Film Mercury detector with sensitivity to better than 0.1 ng of Hg. A standard volume of -80 mesh soil (0.25 cc) is placed in a quartz bulb and heated red hot for one minute to volatize all of the Hg, which is collected on a gold foil. Heating of the gold foil in the analysis procedure releases the Hg for analyses as a gas in the standard manner. Calibration is accomplished by inserting a known concentration of Hg vapor with a hypodermic syringe.

The background concentration of Hg in soils varies widely from area to area, and must be determined from a large number of samples. It is generally the order of 10 parts per billion.

A question remained as to the application of helium and mercury sampling to Alaska geothermal exploration: How does the presence of permafrost affect the diffusion of He and Hg from source to the soil surface? Further basic resarch on this problem is needed, but we do know that we found both He and Hg anomalies in both thawed and in thick permafrost areas.

Prior to the work on the Seward Peninsula rift system we tried both He and Hg sampling in the vicinity of Chena Hot Springs, Alaska. Figure 3 shows a map of Chena Hot Springs with isothermal contours at a depth of 0.5 m, and the soil concentrations of He and Hg in the area. A high value of 795 ppm He was found near the center of the 40°C isotherm at the west end of the area. In general the mercury values tended to outline the same linear anomaly presumed to be a fault in the underlying quartz monzonite crystaline rocks.

He and Hg Results in Central Seward Peninsula

In order to further assess the usefulness of He and Hg surveys some limited profiles were made in the approximately 1 km^2 thaw ellipse at Pilgrim Hot Springs. Figure 4 shows a map of Pilgrim Springs and the location of anomalous He samples. The highest soil concentrations of about 100 ppm He were found near, but not at the highest temperatures at 4.5 m depth (80°C). Figure 5 shows the T, He and Hg values along a profile west to east across the hottest temperature anomaly. In general the helium and mercury values are in agreement. Both are anomalously high at station 20°W which is suggested as a prime drilling site. Along a north-south profile on the 0.0 line the helium values were all close to atmospheric levels, yet a mercury anomaly of 55 ppb was found at a location where the ground temperature was only 20°C. Curiously samples next to the main hot springs pool were low in both He and Hg. The cause of this is likely due to the soil which is a porous sand, and the elevated temperature. The porosity of the soil could allow helium to readily pass through to the atmosphere. Mercury would be easily vaporized by the high temperature and also escape through the porous soil.

Some anomalous helium values were found outside the thaw ellipse across the Pilgrim river as shown in Figure 4. Galvanic and EM-16R resistivity measurements show low resistivity layers beneath the surface and suggest the presence of geothermal water in a band along the river.

There is a second smaller thaw window in the Pilgrim river valley 4 km ENE of Pilgrim Springs. The temperature at 4.5 m was 20°C. The He concentration in the soil was anomalous, 5.52 ppm, and Hg samples were indeterminate, some low some higher than normal. No drilling or deeper temperature measurements have been made, but resistivity measurements indicate a low resistivity layer of 2.5 Ω-m at depth.

As geological mapping progressed in 1980 the general outlines of a proposed rift system emerged. Except for the Nome Taylor Road, all access to the area was by helicopter or boat. Five traverse lines were planned to cross the elements of the rift system to measure gravity, geology, VLF, mercury and helium wherever possible. The limited helicopter and field time did not permit stations as closely spaced as might be desirable.

Figure 6 shows a map of the Seward Peninsula with the location of stations on the five traverse lines. Also shown are the locations of anomalous He soil concentrations found. With the exception of two small anomalies at the north end of the Imuruk Traverse, all the helium anomalies lie within the proposed rift sections A, B, C, or D (Figure 1). Helicopter flying range did not allow us to work farther to the east in segment E.

Figure 7 shows a geologic cross section, and the He and Hg concentrations along the Imuruk Traverse. There are two significant helium anomalies two km apart in the lava fields not far from the recent Lost Jim Flow. The Hg values are also high at these two stations.

Figure 8 shows a geologic cross section at the western end of the rift system where the highest helium anomaly on the traverses was located. The corresponding Hg soil concentration is about normal. On the traverse several large Hg anomalies were found, particularly in a small stream valley at station 128. The cause of this anomaly is not evident. At one time during the gold rush mercury was used to amalgamate the fine placer gold in streams. We tested a soil sample in Quartz Creek which was heavily mined and found the Hg content about average. There was no evidence of mining actively in the valley of station 128, or at 131 which was also anomalous. The Hg aureole is expected to be much larger than that around an He source, so the fact that no He anomaly was found at either station does not rule out a geothermal source of the Hg.

Space does not permit the inclusion of other traverses. In general however the Hg values showed great variability from one station to another, while the He values were almost all near background except for a few anomalies shown on Figure 6.

Conclusions

Our use of He and Hg in the study of the Central Seward peninsula was in part research into the usefulness of these geothermal associated elements. We found that both are probably useful in a hot springs or known thermal anomaly. However as reconnaissance tools we found that the Hg soil concentrations showed great variability. If we had made closer spaced measurements we might be able to explain the variability, but as it is we can only speculate. Perhaps the presence of permafrost affected the ability to collect samples at a uniform soil horizon. There are probably more varied sources of Hg in the rocks than it is the case for He. We found the He sampling produced anomalies in the rift zones, and several significant concentrations which indicate areas of interest for future geothermal exploration.

APPENDIX A

Geological Map Units

Q QTu	Tertiary to Quaternary alluvium, valley fill, includes Kougarok gravels and equivalents, till and alluvium.
QTb Qb	Tertiary to Quaternary basalts of the Kuzitrin Flats and Eva Mtn. Alkalic to Tholeiitic in composition with ultramafic inclusions in the alkali basalts.
Ki	Cretaceous intrusives, mostly quartz monzonite.
Pz Pzc	Thrust sheets of Paleozoic carbonates and meta-carbonates.
P€ P€ms	Precambrian to lower paleozoic metasediments. Schists and gneisses of the Nome Group and York Slate. Locally migmatized in the Northern Bendeleben Mts.

Geologic units generalized from Hudson (1977) and Sainsbury (1972, 1974) with faults on the Imuruk Traverse by Hopkins (1963).

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FIGURE CAPTIONS

Figure 1. Diagram of proposed rift model for the central Seward Peninsula. The graben structure offshore (PCR) is the Port Clarence Rift (Hopkins et al., 1974). The geology is generalized from Hudson (1977). QTb unit are late Tertiary to Quaternary basaltic lava flows. See Appendix A for geologic units.

Figure 2. Solubility of noble gases in fresh water (after Mazor, 1972).

Figure 3. Map of Chena Hot Springs, Alaska with 0.5 m depth isothermal contours, helium and mercury soil concentrations.

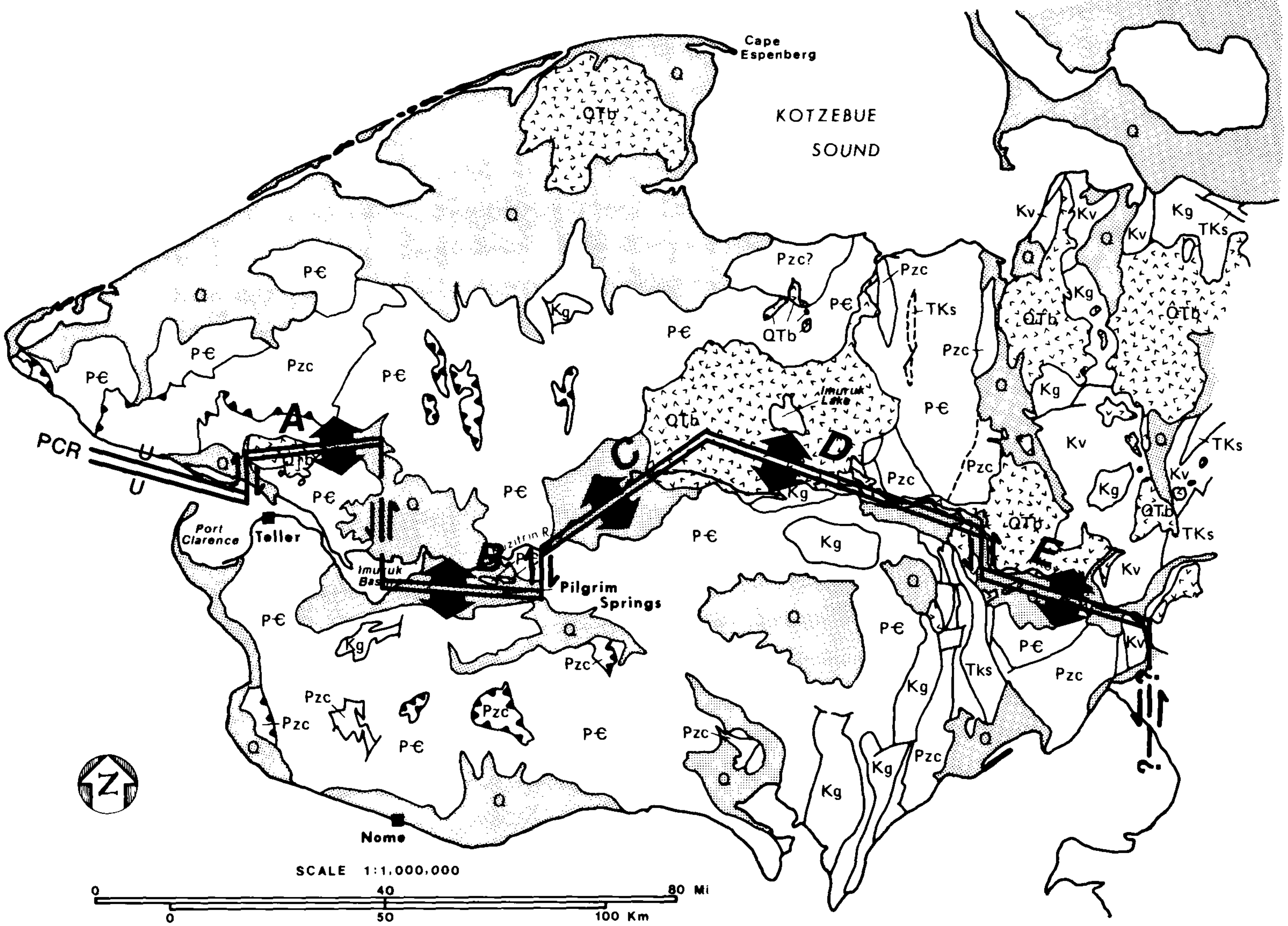
Figure 4. Map of Pilgrim Hot Springs, Alaska showing the elliptical area of thermally disturbed ground and anomalous He values found.

Figure 5. Temperature at 4.5 m depth, He and Hg soil concentrations in a west to east profile across the Pilgrim Hot Springs thaw ellipse.

Figure 6. Map of Seward Peninsula, Alaska showing 5 traverse lines across sections of the proposed rift system and locations of anomalous helium soil concentrations.

Figure 7. Imuruk traverse, helium and mercury soil concentrations. Two significant helium anomalies are found at stations 43 and 44. The mercury values are also higher than the mean at those sites. The nearby Lost Jim Flow is of very recent age. See Appendix A for geologic units.

Figure 8. Agiapuk traverse, helium and mercury soil concentrations. A significant helium anomaly was found at station 123 near lava flows of probable Tertiary age. Two large Hg anomalies were also found without He concentrations. See Appendix A for geologic map units.


Cape
Espenberg
KOTZEBUE
SOUND
Q
QTb
PЄ
Pzc
Pzc?
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TKs
Tks
Imuruk
Lake
PCR
U
A
B
C
D
E
Port
Clarence
Teller
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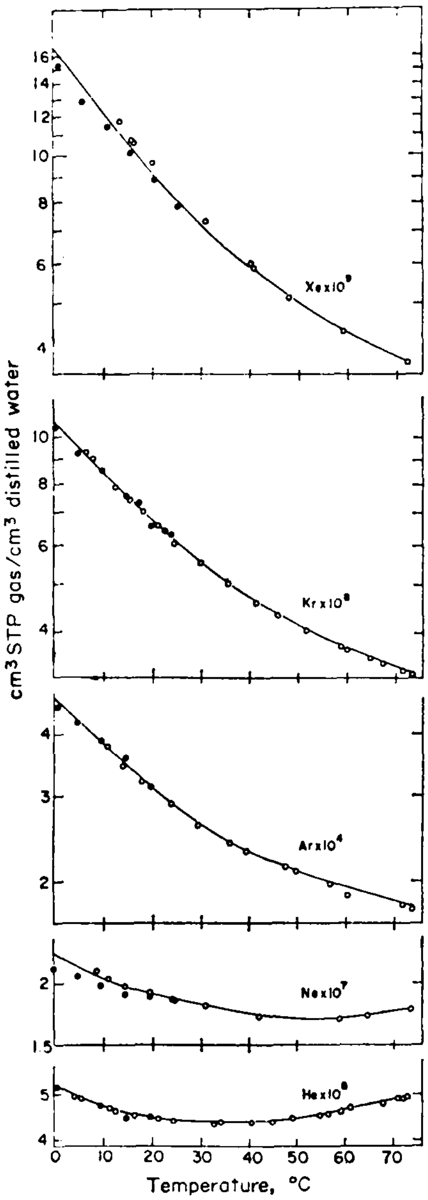
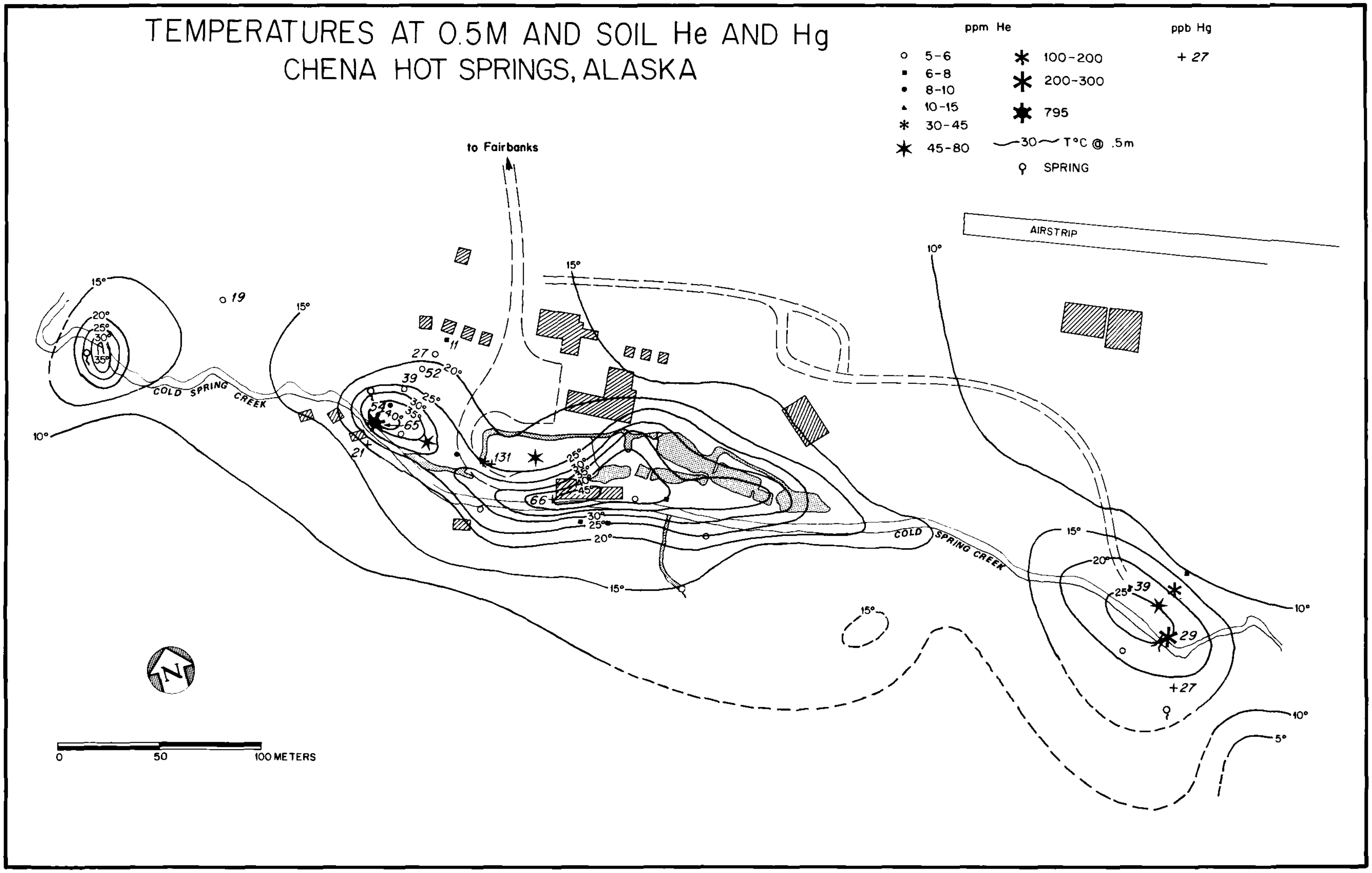
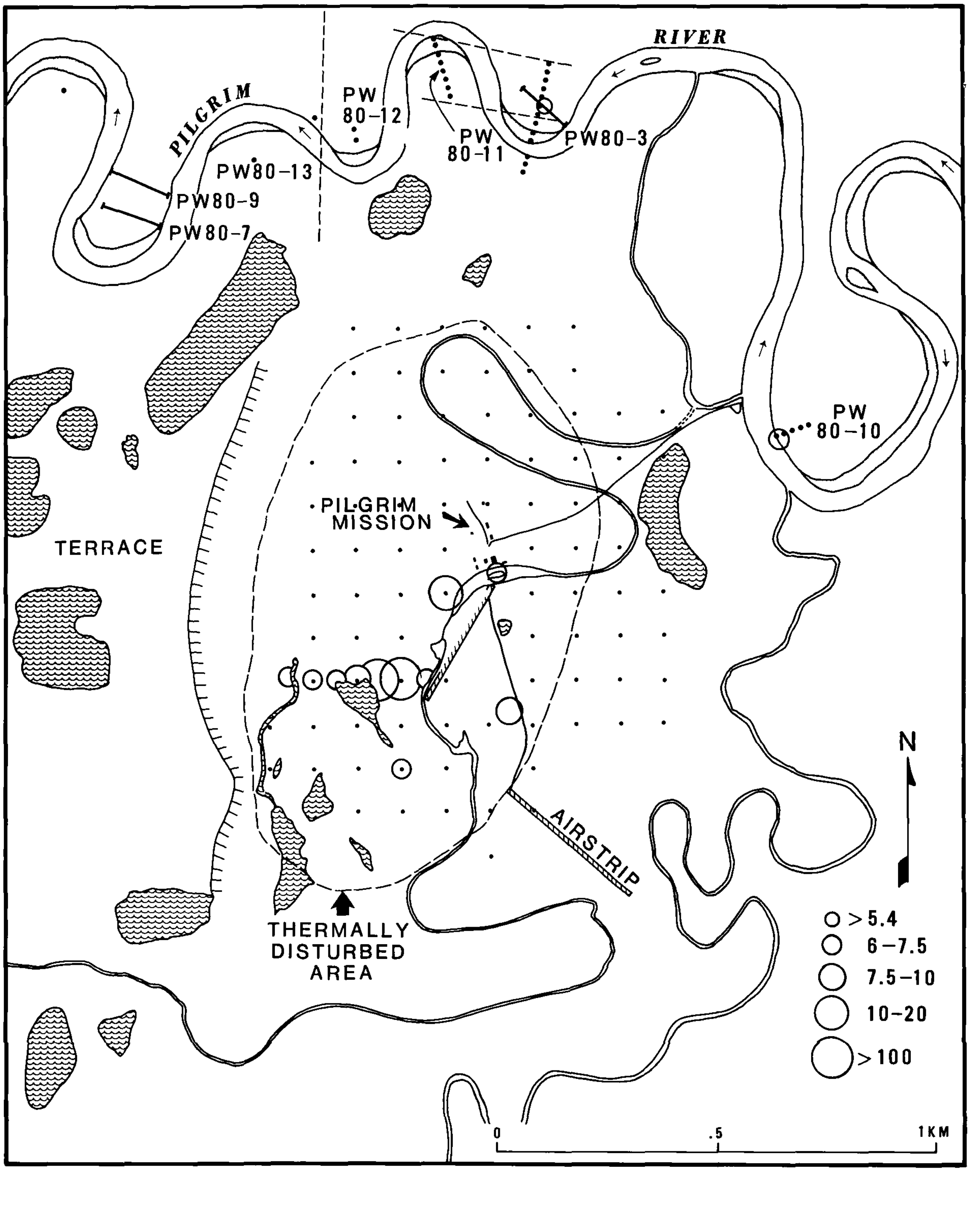


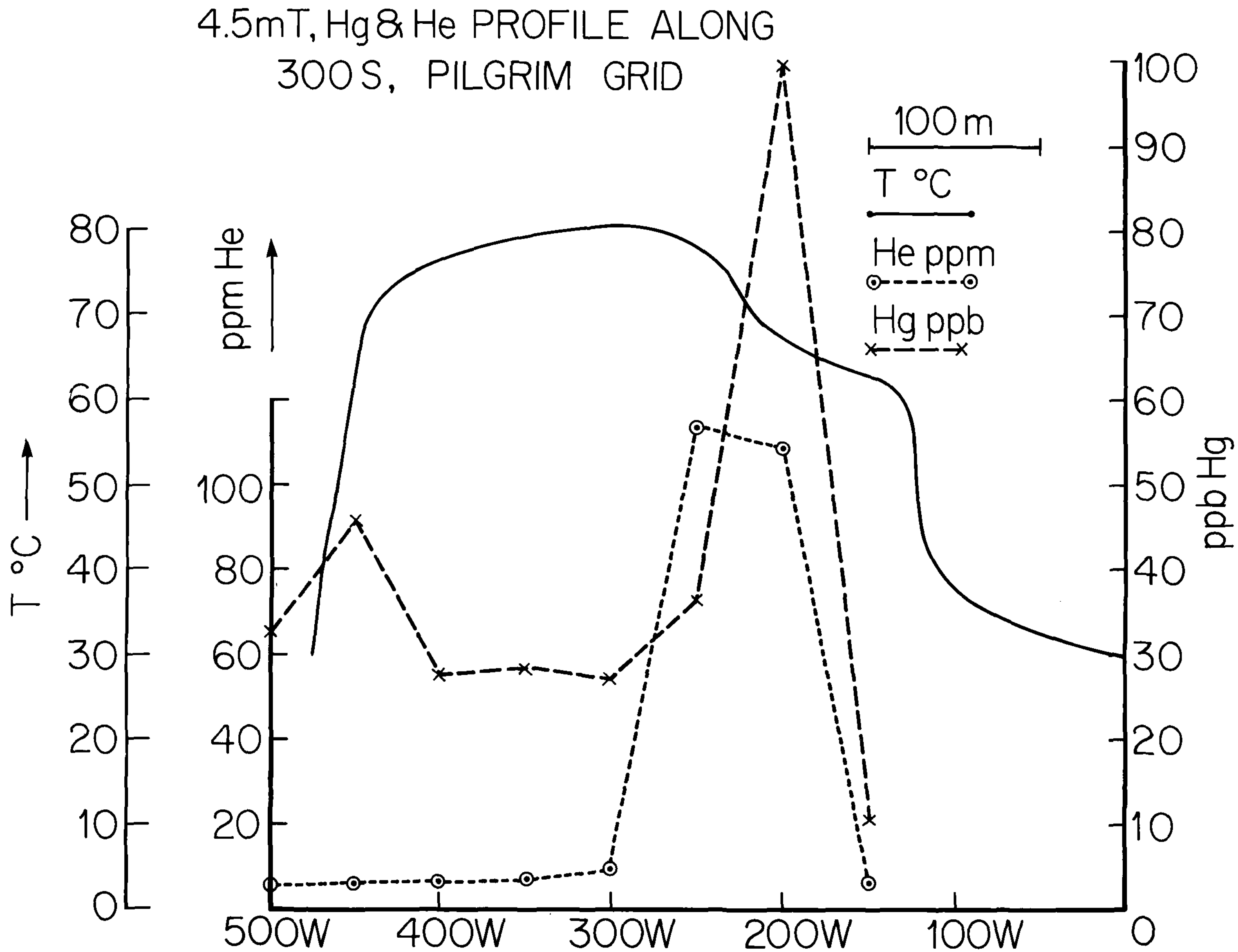
cm3STP gas/cm3 distilled water
Xe x 10^9
Kr x 10^8
Ar x 10^4
Ne x 10^7
He x 10^8
Temperature, °C



TEMPERATURES AT 0.5M AND SOIL He AND Hg
CHENA HOT SPRINGS, ALASKA
ppm He
5-6
6-8
8-10
10-15
30-45
45-80
100-200
200-300
795
ppb Hg
+ 27
30 T°C @ .5m
SPRING
to Fairbanks
AIRSTRIP
COLD SPRING CREEK
0 50 100 METERS

HELIUM SOIL CONCENTRATION




4.5mT, Hg & He PROFILE ALONG
300 S, PILGRIM GRID
100 m
T °C
He ppm
Hg ppb
T °C
ppm He
ppb Hg
0
10
20
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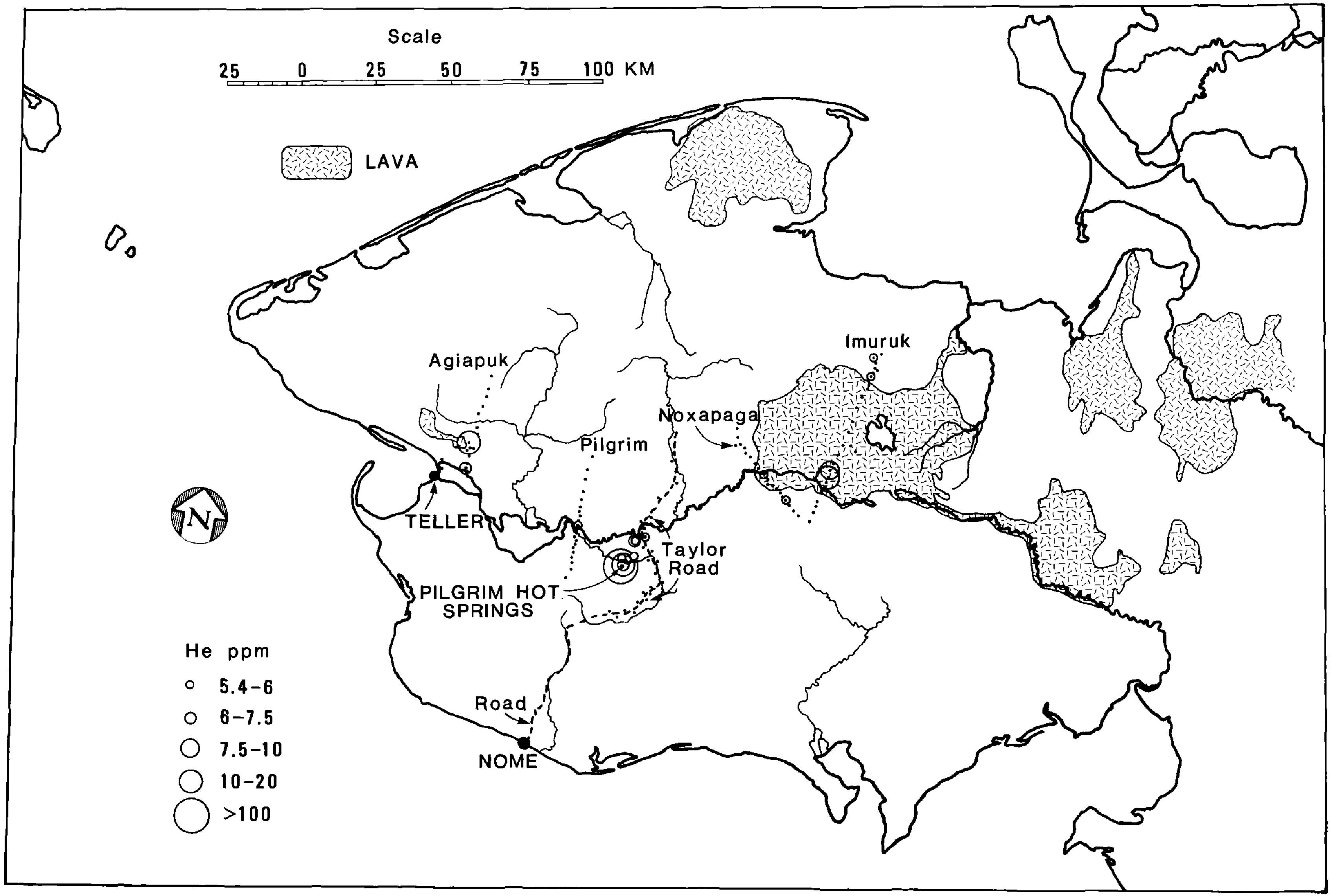
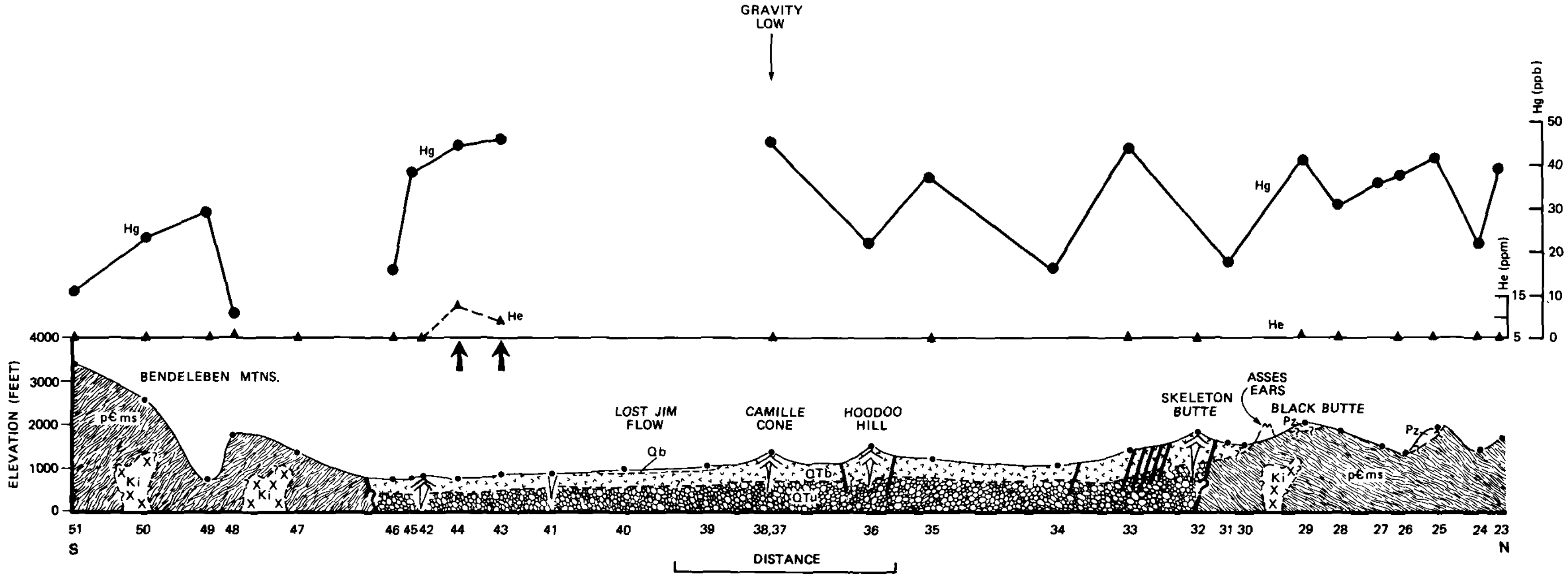
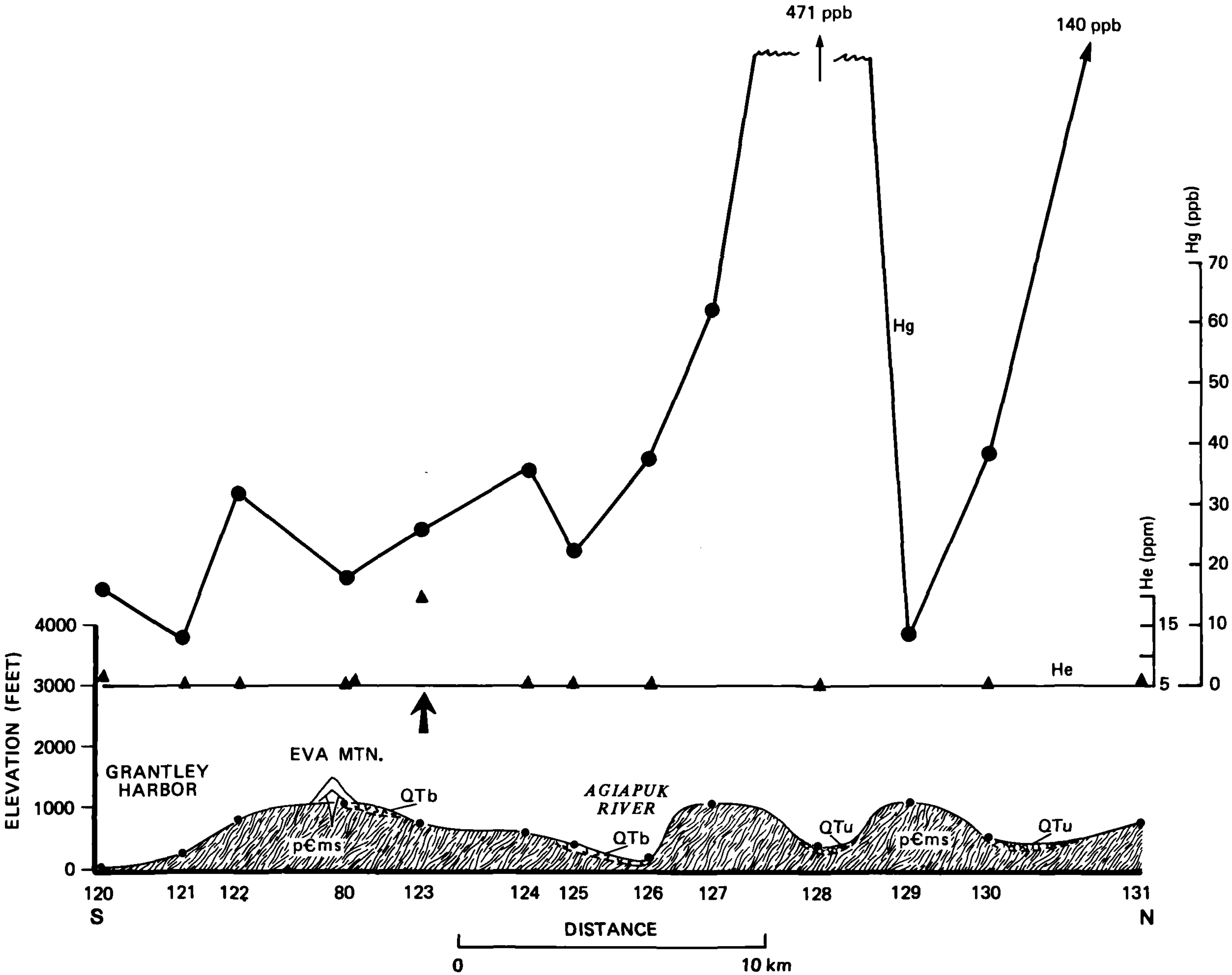
200W
100W


Scale
25 0 25 50 75 100 KM
LAVA
Agiapuk
Pilgrim
Noxapaga
Imuruk
TELLER
Taylor
Road
PILGRIM HOT
SPRINGS
He ppm
5.4-6
6-7.5
7.5-10
10-20
>100
Road
NOME



IMURUK TRAVERSE



AGIAPUK TRAVERSE

KANSAS GEOLOGICAL SURVEY
Environmental Geology Section

1930 Avenue "A", Campus West
The University of Kansas
Lawrence, Kansas 66044
913-864-4991

June 12, 1981

Mr. Carl Ruscetta
Earth Science Laboratory
University of Utah Research Institute
420 Chipeta Way
Salt Lake City, Utah 84108

Dear Carl:

Enclosed is a manuscript that more or less summarizes our geothermal program, along with preliminary results.

Thank you for your patience.

Sincerely,

Don W. Steeples, Chief
Environmental Geology
and Geophysics Section

DWS:ep
Enc.

Copy.

GEOTHERMAL EVALUATION OF KANSAS - PRELIMINARY RESULTS

by

Don W. Steeples

Kansas Geological Survey
The University of Kansas
Lawrence, Kansas 66044

June, 1981

INTRODUCTION

A low-temperature geothermal resource investigation was begun in Kansas in 1979. This paper should be considered a progress report of that investigation, so the results and speculations should be considered preliminary in nature. The individual facets of the study are discussed individually after the outline of the geologic and tectonic framework.

GEOLOGIC AND TECTONIC FRAMEWORK

The Midcontinent portion of North America is the most stable part of the continent, tectonically. The region is relatively aseismic, and there has been no significant deformation of the crust since at least the Late Paleozoic. The major tectonic elements of this region are (1) the southern extension of the Central North American Rift System (Midcontinent Rift; Midcontinent Geophysical Anomaly; Figure 1); (2) the Nemaha Ridge; and (3) the Central Kansas Uplift.

The Central North American Rift System [Ocola and Meyer, 1973; Chase and Gilmer, 1973] can be traced from central Kansas across southeastern Nebraska, Iowa, and Minnesota to its outcrop area in the Great Lakes region. The rift is marked by pronounced gravity and magnetic anomalies [King and Zietz, 1971; Lyons, 1950; Thiel, 1956] and is underlain by mafic igneous rocks, mostly basalt and gabbro, and arkosic sedimentary rocks. The feature is generally regarded as an abortive continental rift which occurred about 1100 m.y. ago [Goldich et al., 1961; Silver and Green, 1963, 1972; Goldich et al., 1966; Chaudhuri and Faure, 1967; Van Schmus, 1971].

The Nemaha Ridge is a striking tectonic feature which was intermittently active during Paleozoic time. It is certainly a major crustal fracture zone, for mylonitized basement rocks have been brought up from within it, and cataclasis is a common feature along its extent from northeastern Kansas into Oklahoma [Bickford et al., 1981]. The fault zone is upthrown on the western side, forming the feature known as the Nemaha Ridge. The eastern flank of the Nemaha Ridge is bounded by the Humboldt Fault Zone [Steeples et al., 1979]. Earthquakes as large as Modified Mercalli Intensity VII have occurred along the Humboldt Fault Zone in historic time [DuBois and Wilson, 1978].

The Central Kansas Uplift (Figure 1) is a broad region in which basement rocks have been moved upward and which is characterized by fault zones and cataclasis. The feature is evidently coextensive with the Cambridge Arch in Nebraska. Although the Central Kansas Uplift was active during the Paleozoic, little is known about its Precambrian history. A relatively high level of microearthquake activity (more than 20 events per year larger than magnitude 1) occurs along this structural trend [Steeples, 1980].

The crystalline crust in the Midcontinent is buried under about 1000 m of sedimentary rocks and is thus mostly known from studies of numerous drill holes [Muehlberger et al., 1966; Goldrich et al., 1966; Lidiak et al., 1981; Kisvarsanyi, 1980]. The crust in this area is notable for its predominantly granitic composition. Mafic rocks are rare, and metamorphic rocks, though present in many places, are not abundant. A major feature of the crystalline crust in the Midcontinent is its division into a northern terrane, consisting of somewhat deformed and sheared granitic rocks and lesser amounts of metamorphic rocks that occur in northern Missouri, northern Kansas, and Nebraska, and a southern terrane totally dominated by silicic volcanic rocks and associated epizonal granitic plutons. The southern terrane can be traced from northern Ohio across Indiana, Illinois, southern Missouri, southern Kansas, and Oklahoma into the Texas Panhandle. Geochronological studies [Bickford et al., 1981; Denison et al., 1981] indicate that the northern terrane is generally older, with many rocks yielding ages of 1640 m.y. (U/Pb, zircon) to 1740 m.y. (Rb-Sr), whereas the southern terrane varies in age from about 1475 m.y. in the St. Francois Mountains of southeastern Missouri [Bickford and Mose, 1975] to about 1380 m.y. in southwestern Missouri, southeastern Kansas, and Oklahoma [Bickford and Lewis, 1979; Bickford et al., 1981].

Lying upon the crystalline crust in the Midcontinent region is a section of sedimentary rocks ranging from about 150 m in thickness over parts of the buried Nemaha Ridge to as great as 2 to 3 km thick in basins such as the Hugoton Basin of southwestern Kansas and northwestern Oklahoma. The average thickness of the sedimentary rock section in eastern Kansas where our drilling projects were done is about 1 km. The rocks range in age from Late Cambrian to Pennsylvanian or Permian in eastern Kansas, but there is a thick Cretaceous section in central Kansas, and rocks of Tertiary age occur on the western plains. Paleozoic rocks in the Midcontinent region are mostly marine in origin and are dominated by carbonate units and shale.

PIGGYBACK DRILLING

Normal exploration procedure for most resources involves drilling as a culmination of geologic and geophysical investigations. It would seem that drilling four holes in an area the size of Kansas at the outset of a regional geothermal evaluation is a reverse approach.

However, the drilling was done as a "piggyback" operation at a relatively low cost compared to the total project. The data obtained from the boreholes has enabled us to provide high quality heat-flow data for Kansas and to better evaluate thermal information available from the petroleum drill holes.

THE DRILLING OPPORTUNITY

In 1976 a cooperative program between the U.S. Geological Survey and the Kansas Geological Survey was begun. The purpose of the study was to determine the regional geohydrologic characteristics of the Arbuckle Group to include definition of flow patterns in the Arbuckle and in relation to other overlying units, determination of hydraulic parameters in the various units, and determination of regional chemical quality. Test data procured from oil exploration companies are being analyzed to accomplish this objective.

Additional funding to the Kansas Geological Survey became available in FY 1979 and FY 1980 for the purpose of test drilling and installation of deep monitor wells. This funding was matched and increased by the U.S. Geological Survey, the U.S. Army Corps of Engineers, and the Kansas Department of Health and Environment; the project was expanded to include determination of hydrologic properties of the Arbuckle and other units by drilling at specific locations.

We were not involved professionally in any aspect of the Arbuckle project when it was originally funded. However, we realized that valuable petrologic and geophysical data could be obtained if these holes could be deepened by about 100 m to penetrate the crystalline basement. Four drilling sites were selected. Three of the four sites were selected so that, in addition to the hydrologic study, basement rock samples and geophysical data could be obtained from the same holes. Two of the sites were located above intense, circular magnetic anomalies; the third was located in a region where sparse well control indicated a terrane of silicic volcanic rocks in the basement.

We received funding from LASL and the U.S. Department of Energy (DOE) to deepen the first two holes, recover core from the basement, and perform high-quality heat flow measurements in the holes. Funding was also received from a separate LASL contract to deepen the third hole; however, severe circulation problems developed within the Arbuckle Formation before the "piggyback" experiment could begin. All of the money from that contract was returned to LASL.

Cost of Drilling Operations

A contract was awarded in September 1979 in the amount of $444,701 for work to be performed at the first three sites. A second contract, in the amount of $129,341 for work to be performed at the fourth site, was awarded in April 1980. Seventeen drill-stem tests were completed for the purpose of determining the hydraulic relationships between the Arbuckle and other major overlying aquifers. Three cores of the Arbuckle were collected in addition to the basement cores, and these are currently being analyzed under a third contract. A complete suite of geophysical logs, including vertical flow determinations, were completed at each site. Also, acoustical televiewer photographs were taken over selected intervals at three sites. Water samples have been collected at three sites and are undergoing complete chemical analyses, including determinations for age dating with lithium, bromine, strontium, deuterium, carbon 14, and tritium.

The increased incremental cost of approximately $9,000 per hole for coring in the Precambrian section of two holes is included in the above figures. Most of the preliminary scientific results reported here are a direct result of this small additional expenditure, thus demonstrating the value of "piggybacking." Less than $30,000 of the total geothermal funding was used for the heat flow studies and the "piggyback" drilling.

SCIENTIFIC RESULTS FROM DRILLING

The author had significant input as to the location of the holes, and their sites were chosen to maximize potential information from the basement, subject only to the general suitability of the location to the primary mission of the drilling project, i.e., the hydrologic study of the Arbuckle. The legal descriptions and locations of holes drilled are given in Table 1.

Drilling at the first hole (Miami County) was completed on December 10, 1979. Approximately 8 m of 6.7-cm-diameter core of fresh granite were recovered from a depth of 658 to 666 m. This hole was located on a sharp 1000-γ circular aeromagnetic high, shown as locality 1 on Figures 1 and 2.

The second hole (Douglas County; locality 2 on Figures 1 and 2) was also located on a circular magnetic high with an amplitude of about 1100-γ; drilling was completed on March 19, 1980. Three meters of 10-cm-diameter core of fresh granite were recovered from a depth of 905 to 908 m. The 3 meters represent only 58 percent recovery of the 5.2 meters cored. We were very fortunate not to lose all of the core, as it started slipping out of the core barrel during the trip up the hole. The core catcher barely hooked the core again and prevented disaster. We were not charged for the core that was lost.

Two additional holes (localities 3 and 4 on Figures 1 and 2) were drilled to depths of 1117 m and 554 m, respectively. Severe lost-circulation problems developed on both of these holes within parts of the Arbuckle Formation, and drilling was halted at that depth because the primary objective of the drilling had been met. Penetration of Precambrian basement at sites 3 and 4 would have cost an additional (possibly very large) undetermined amount of money.

The scientific data we ultimately expect to obtain from the drill core and from the geophysical measurements include the following: age, petrography, major and trace element chemical composition, density, and remanent magnetism of the rocks encountered; heat flow; and heat production of the rock material. The holes into basement can be made suitable for hydrofracturing experiments to measure in situ stress, provided future funding becomes available. The holes will be available to other scientists for other experiments within 2 years. Interested individuals should contact the author of this report.

Some preliminary data are available on age, thermal gradient, and heat flow. The age of the Precambrian cores is about 1350 m.y. (U/Pb of zircons) [Steeples and Bickford, 1981] indicating that the circular magnetic anomalies represent a suite of intrusions younger than the "normal" 1650 m.y. age for the crust in the area.

TABLE 1. Legal Description of Drill-Hole Locations in Kansas

	Location	Total Depth
Douglas County	SE 1/4 NW 1/4 NW 1/4 Sec..13, T12S, R17E	908 m
Labette County	Center of SE 1/4 Sec.22, T31S, R20E	553 m
Miami County	SE 1/4 SW 1/4 SE 1/4 Sec.18, T18S, R23E	666 m
Saline County	SW 1/4 SW 1/4 SW 1/4 Sec.32, T13S, R2W	1117 m

Geothermal Gradients

Preliminary thermal logging has been performed on all four holes by personnel from David Blackwell's laboratory at Southern Methodist University. The thermal logging equipment was not capable of reaching the bottom of the holes, so these data should be considered preliminary, pending results from deeper logging. Samples of core or well-cuttings have been sent to Blackwell's laboratory for thermal conductivity measurements. The following geothermal gradients have been measured to date in the four holes drilled on this project:

Location	Gradient	Depth Logged
Douglas County	30.3°C/km	565 m
Labette County	28.5°C/km	520 m
Miami County	36.0°C/km	395 m
Saline County	30.7°C/km	565 m

Preliminary data from Blackwell indicate an unusually high rate of radioactive heat generation, about 11 heat generation units, in the core from the Miami County hole, compared with the 5 to 6 heat generation units for typical granites. The heat generation from the core obtained from the Douglas County hole was not anomalously high.

High quality heat flow measurements have been made in the four drill holes [Blackwell et al., 1981]. The average heat flow in the holes was 54 ± 5 milliwatts/square meter (1.3 ± 0.1 heat flow units). These heat flow values are typical for areas of stable Precambrian-aged crust. Additional results will be available after the holes are logged to total depth.

OTHER STUDIES IN PROGRESS

The aeromagnetic map of Kansas (Figure 2) was funded in part with geothermal money. A preliminary interpretation of that map has been published by Yarger (1981). Gravity data are being gathered to supplement the interpretation of the aeromagnetic map. A gravity map contoured at one milligal intervals will be available for the eastern half of Kansas by mid-1982.

We are continuing to measure geothermal gradients statewide and will complete that phase of the field work in time to publish a geothermal gradient map in 1982. Preliminary indications are that geothermal gradients state wide are consistently in the range of 30 to 35°C/km.

Geothermal Speculations for Kansas

Figure 3 shows structural contours on the Precambrian surface relative to sea level. There are areas on this map that bear discussion with respect to low grade geothermal prospects. They will be discussed in counter-clockwise order around Kansas starting with the southeast corner of the state.

1. In the southeastern portion of the state, dip is toward the northwest away from the Ozark Uplift. Sedimentary cover is roughly 500 meters thick with relatively good quality water in much of the sedimentary section. The Arbuckle group of early Ordovician age is used as a fresh water source by several towns. The water temperature is in the range of 25° to 35°C, warm enough to be used for heat pump applications for space heating. This area is the most likely portion of Kansas to have any geothermal applications in the next 5 to 10 years. The geothermal gradient below depths of 500 meters is quite low - possibly in the range of 15 to 20°C/km.

2. The Nemaha Ridge is a buried granitic mountain range that reaches to within 200 meters of the surface in northeastern Kansas. Geothermal gradients in the sedimentary section exceed 50°C/km, but geothermal prospects are poor because the sedimentary section is so thin. The high geothermal gradient is not thought to persist below the bottom of the sediments.

3. To the west of the Nemaha Ridge is the Midcontinent Geophysical Anomaly (Figure 1). In Figure 3 the MGA is more or less bounded on the southeastern flank by six kimberlite intrusions of Cretaceous age. The northwest flank of the MGA is bounded by a zone of microearthquakes that have occurred since 1978. These boundary features along opposite sides of the MGA indicate structural zones that may allow circulation and convection of water deep into the crust.

   The MGA itself is caused by mafic intrusive and extrusive rocks of late Precambrian age as discussed earlier. Surrounding the MGA is an arkosic Precambrian age sandstone (Rice Formation) that developed during the later stages of rifting and subsidence. The thickness of the Rice Formation is unknown, but probably substantial. Modeling of the gravity and magnetics by Yarger (1981) indicates that the edges of the Rice Formation are probably fault-bounded. It is possible that the Rice Formation is several kilometers thick, based on differential arrival times at two seismograph stations in the vicinity of the MGA.

   A deep seismic reflection profile experiment is being performed across the MGA by the Consortium for Continental Reflection Profiling (COCORP) during 1981. The results of this experiment will probably allow calculation of the thickness of the Rice Formation.

   If there are, indeed, several kilometers of Rice Formation present, geothermal prospects for production of large volumes of warm water would be excellent. The quality of any such water is unknown.

4. The Central Kansas Uplift (Figure 1) is much the same as the Nemaha Ridge from a geothermal standpoint. However, thousands of oil wells produce water at temperatures of about 40 to 45°C in conjunction with oil production. The heat from this water is not purposefully extracted prior to reinjection in salt water disposal wells. Other than this by-product warm water, the geothermal prospects are not good.

5. In northwestern Kansas, the Dakota Formation may be locally useful for heat-pump applications. Will Gosnold (1981 personal communication) has discovered that the Dakota waters in western Nebraska are unusually warm, apparently as a result of convection updip to the east from the Denver-Julesburg Basin. It is not yet clear whether this same effect is present in Kansas, but it will be investigated in the near future.

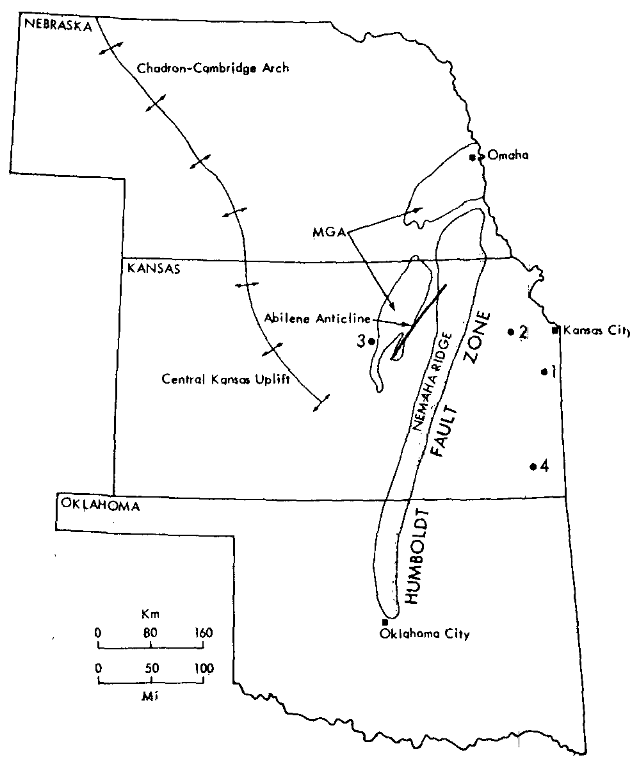
6. In southwestern and south-central Kansas, the sedimentary section is 2 to 3 km thick. Waters produced in conjunction with petroleum are at temperatures of 60 to 65°C. Water quality is generally poor and the prospects for geothermal development are not bright because of sparse population density and lack of industry in the area.

FIGURE CAPTIONS

FIGURE 1. Principal positive structural features in Kansas. Major basins include the Forest City Basin in northeast Kansas, the Cherokee Basin in southeast Kansas, the Salina Basin in north-central Kansas, and the Hugoton embayment of the Anadarko Basin in southwestern Kansas. Drill sites for heat flow measurements are shown chronologically by numbers 1, 2, 3, and 4.

FIGURE 2. Aeromagnetic map of Kansas. Contour interval is 50 gammas. Drill sites from Figure 1 are shown as 1, 2, 3, and 4.

FIGURE 3. Precambrian structural contours of Kansas with 1000 foot contour interval relative to sea level. Stars in northeast Kansas denote kimberlite locations. Faults and microearthquakes show locations where crustal fractures are present.



NEBRASKA
Chadron-Cambridge Arch
Omaha
MGA
KANSAS
Abilene Anticline
3
2
Kansas City
1
4
Central Kansas Uplift
NEMAHA RIDGE
HUMBOLDT FAULT ZONE
OKLAHOMA
Km
0 80 160
0 50 100
Mi
Oklahoma City

FIGURE 1.

AEROMAGNETIC MAP OF KANSAS

H. Yarger, R. Robertson, J. Martin, K. Ng, R. Sooby and R. Wentland

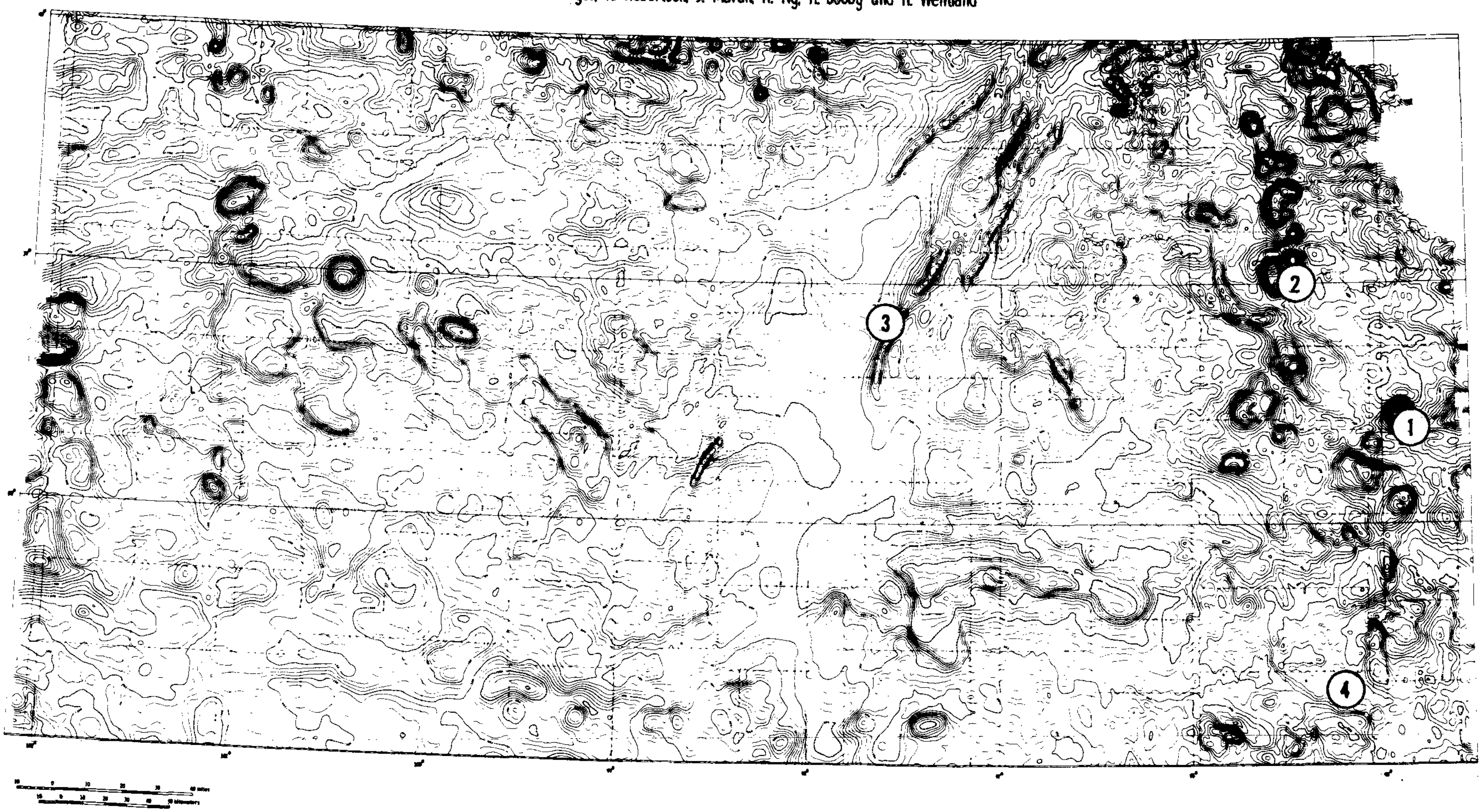
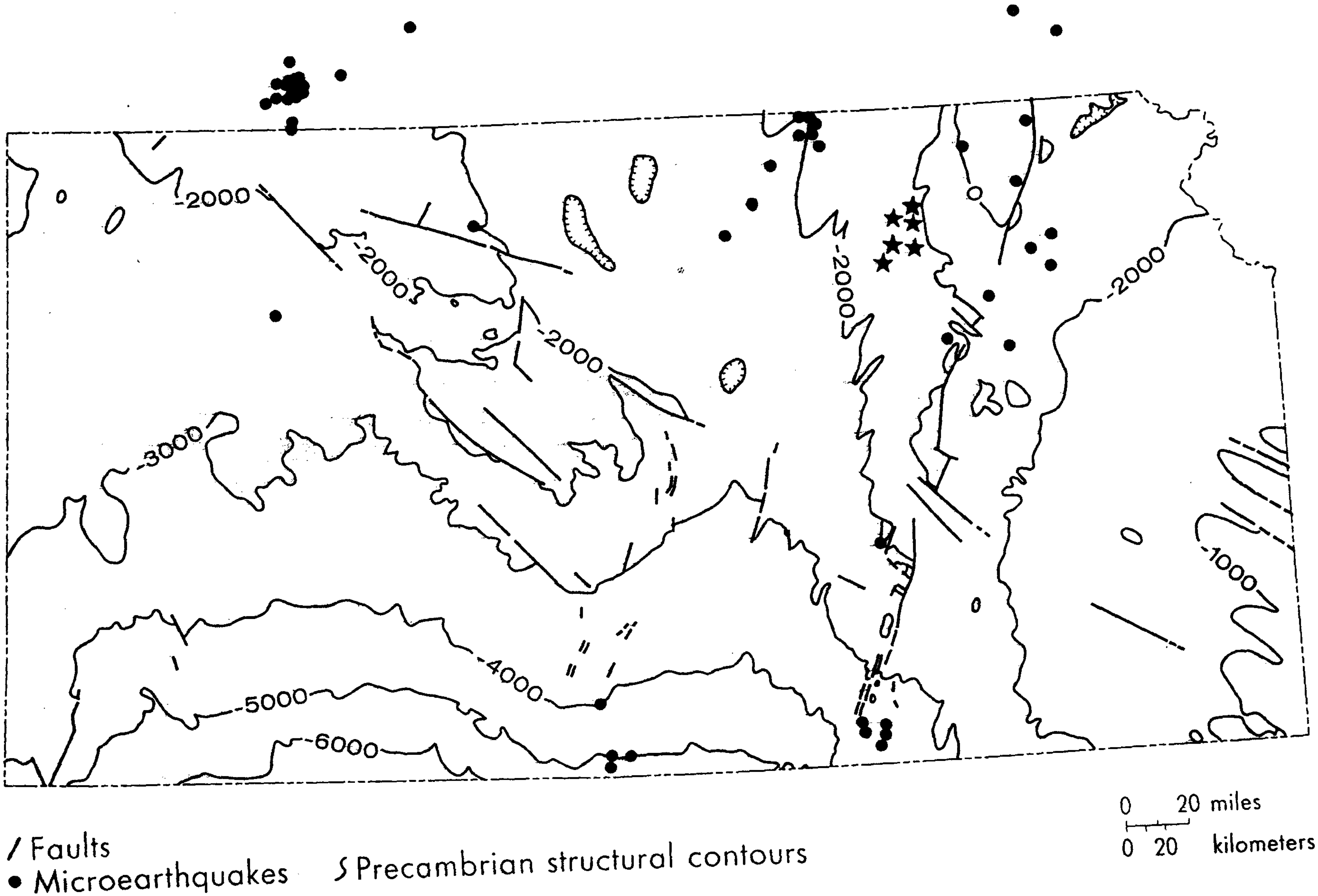


FIGURE 2.



-2000
-2000
-2000
-2000
-2000
-3000
-4000
-5000
-6000
-1000
0
0
0
0
20 miles
0 20 kilometers
Faults
Microearthquakes
Precambrian structural contours

FIGURE 3.

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THERMAL SPRINGS OF ARIZONA

James C. Witcher
Arizona Bureau of Geology and Mineral Technology
University of Arizona, Tucson

Identification of truly thermal springs is an indispensable aid in the assessment of a region's geothermal characteristics. Over the years numerous lists of thermal springs in Arizona have been compiled and we present yet another. Although the word thermal implies heat, there is considerable subjectivism or arbitrariness in its application. In geothermal work what is important is anomalous or unusual heat--something above a norm. I have devised a functional scheme useful in identifying those Arizona springs judged to be carrying anomalous heat. The method is readily applied to any new springs that may be encountered. The results of this updated version are shown in Table 1. Also, possible heat sources are briefly outlined in the text.

Defining Thermal Springs

Over the years, springs given the label "thermal" may or may not carry anomalous heat. Likewise, it is possible for springs not so labeled to be anomalously warm. The explanation for this is not difficult; it is to be found in Arizona's regional topographic-climatic variances.

Depending upon the season, the temperature of the earth down to 10 or 20 meters is slightly above or below the mean annual air temperature (MAT). Because springs are surface discharges of water contained in the pores and fractures of rock at very shallow depth, springs tend to have a temperature close to the MAT. Spring temperatures that are much higher than the MAT are thermal springs and their waters are heated by anomalously hot rock near the surface or by circulation through hot rock at much greater depths.

The MAT in Arizona ranges from less than 6°C to over 22°C, primarily because the surface elevation is quite varied; therefore, a similar range in spring temperatures is to be expected. Generally, a thermal spring at a high elevation will have a lower temperature than an equally significant thermal spring at a lower elevation where the MAT is higher. Thus, the MAT provides a baseline from which a thermal spring can be defined from place to place.

However, in order to actually classify a spring as being thermal, some comparisons, or temperature standard above the baseline temperature, is needed. This comparison temperature should fall somewhere between normal spring temperatures and those that are anomalously high and obviously thermal. The temperature distribution of Arizona's springs relative to the mean annual air temperature (MAT) is utilized to find this comparison temperature.

Spring temperatures measured during field work and reported in geologic literature covering Arizona were compiled. All available MAT data for Arizona were plotted and contoured on a map of Arizona in order to determine the MAT at the spring locations. The MAT for individual spring locations is subtracted from the individual measured spring temperatures and plotted on a frequency diagram in Figure (1). A mostly normal distribution of spring temperatures relative to the MAT is evident. The mean spring temperature is slightly above the MAT. This mean spring temperature relates to the average circulation depth of these waters below the surface.

However, the distribution is not perfectly normal when all springs in Arizona are considered. Actually, the distribution appears to have two means with similar standard deviations. When the mean spring temperature of the Basin and Range province is compared to the mean spring temperature of the Colorado Plateau province (Figure 2), a bimodal mean spring temperature is evident, the former being the higher. If the same average circulation depth and average rock thermal conductivities are assumed for both provinces, the difference may relate to the higher conductive heat flow observed in the Basin and Range province. If this is true, the higher mean spring temperature of the Basin and Range springs is caused by a higher average subsurface temperature gradient. It should be pointed out that other explanations are plausible such as differences in surface vegetative cover, average spring flow rates, and seasonal recharge.

The apparent deviation of spring temperatures below the means, assuming a normal distribution, is believed to be caused by discharge from perched water tables close to recharge sources and not discharge from the static water table.

Thermal waters may be subdivided arbitrarily into "hot" and "warm." Hot springs for all of Arizona are here defined as those having temperatures that exceed the MAT by the sum of the mean spring temperature for all springs and the standard deviation (Figure 1). Thus, the comparison temperature used to define a hot spring is $15^{o}C$ above a spring's MAT. In the Basin and Range province the comparison temperature used to define a "warm spring" is $10^{o}C$ above the appropriate MAT. For the Colorado Plateau province $6^{o}C$ above the MAT defines a "warm spring" (Figure 2). These definitions apply only to Arizona and may vary in other states having different geological terrains and subsurface geophysical properties.

Origin of Thermal Springs

Thermal springs, as herein defined, originate from a combination of special conditions. These conditions are basic elements in any geothermal system and they have to work in concert before a system can exist naturally. These elements are: (1) a heat source; (2) a recharge source; (3) a circulation framework or storage reservoir; and (4) a discharge mechanism.

The most basic element is the heat source because it alone separates geothermal spring systems from all others. In Arizona, igneous heat sources are tentatively ruled out because no Recent or Pleistocene silicic volcanism is known. Silicic magma is very viscous and tends to collect in large shallow storage sites. These bodies of magma contain enormous quantities of heat and may require several hundred thousand years to cool to ambient temperature, thereby providing significant heat to overlying rocks and contained fluids.

Recent and Pleistocene basaltic volcanism is known in Arizona; but intrusions related to this volcanism are small plugs, dikes and sills, because basaltic magma is very fluid. Small plugs, dikes and sills cool to ambient temperature in a few months or years and contribute only minor quantities of heat to the surrounding rocks.

The normal flow of heat from the earth's interior is probably the major source of heat for Arizona's thermal springs. Because the earth's internal heat flows or conducts through rock toward the surface, subsurface temperatures in Arizona generally increase at least 30^{o}C for every kilometer of depth; therefore, water circulating deeper than 300 meters for a period of time will be heated by subsurface rocks a minimum of 10^{o}C above the MAT on the surface. Provided little loss of heat occurs on the way back to the surface, these circulating waters will discharge as thermal springs.

The detailed mechanics and geologic conditions required for deep circulation of water are beyond the scope of this article. However, it is believed that forced convection accounts for Arizona's thermal springs because the vertical permeabilities in fault zones and Arizona's subsurface temperature gradients are too low for free convection. Free convection is buoyant flow of water caused by a temperature-induced vertical differential in water density. Forced convection is pressure-induced water flow caused by elevation differences between the recharge water table and the springs discharge elevation. Deep forced convection requires special structures, stratigraphic geometries and geohydrologic conditions.

Studies of Arizona's thermal springs are but a part of the Arizona Bureau of Geology and Mineral Technology's assessment and characterization of Arizona's geothermal resources. The entire study is being funded by the U.S. Department of Energy.

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THERMAL SPRINGS OF ARIZONA

#	NAME	LOCATION	T°C	T-MAT°C
1	Warm Spring	A-1-20-12AC (unsurv.)	29.4	14.4
2	Hanna Creek Hot Springs	A-1-31-29AD	55.5	42.5
3	Warm Spring	A-4½-20-36CB (unsurv.)	24.4	10.4
4	White River Salt Spring	A-4½-20-35AD (unsurv.)	28.3	13.3
5	Roosevelt Dam Hot Spring	A-4-12-19DDB	48.0	28.0
6	Hot Spring	A-9-6-26AB (unsurv.)	36.6	17.6
7	Verde Hot Springs	A-11-6-3B	41.0	23.0
8	Salado Spring	A-12-28-17DCA	21.7	11.7
9	Henderson Ranch Spring	B-8-1-33BAC	30.3	11.3
10	Alkalai Spring	B-8-1-33DB	31.2	12.2
11	Castle Hot Springs	B-8-1-34CC	54.7	35.7
12	Kaiser Hot Spring	B-14-12-10AD	37.0	19.0
13	Cofer Hot Spring	B-16-13-25CAD	37.0	18.0
14	Warm Spring	B-18-13-25DB	28.3	10.3
15	Warm Spring	B-18-19-33DC	29.2	10.2
16	Spring	B-20-9-30CC	27.0	14.0
17	Hot Spring	B-30-23-15CBD	32.0	12.0
18	Hot Spring	B-30-23-26BBC	30.0	10.0
19	Pakoon Spring	B-35-16-24BD	28.0	10.0
20	Agua Caliente Spring	C-5-10-19AA	40.0	18.0
21	Radium Hot Spring	C-8-18-12CC	60.0	38.0
22	Spring	D-2-31-35ABB (unsurv.)	25.6	10.6
23	Mescal Warm Spring	D-3-17-20CBC	29.1	14.0
24	Coolidge Dam Hot Spring	D-3-18-17DC	36.6	18.6
25	Miguel Raton Spring	D-3-31-3ADC	26.7	11.7
26	Spring	D-4-23-21AA	27.2	10.2
27	Spring	D-4-23-21AD	31.5	14.5
28	Tom Niece Spring	D-4-23-22BD	28.3	11.3
29	Eagle Creek Hot Spring	D-4-28-35ABB	42.5	25.5
30	Clifton Hot Spring	D-4-30-18CCD	70.0	53.0
31	Clifton Hot Spring	D-4-30-18CDC	50.0	33.0
32	Clifton Hot Spring	D-4-30-19CAA	33.0	16.0
33	Clifton Hot Spring	D-4-30-30DBC	38.0	21.0
34	Warm Spring	D-5-19-23BDD	26.0	11.0
35	Indian Hot Springs	D-5-24-17AD	48.8	30.8
36	Spring	D-5-24-16CB	33.0	16.0
37	Gillard Hot Spring	D-5-29-27AAD	84.0	67.0
38	Spring	D-7-24-13DC	29.4	12.4
39	Spring	D-10-29-23DD	26.1	10.1
40	Spring	D-12-21-31CA	32.5	17.5
41	Agua Caliente Spring	D-13-16-20CDD	32.0	12.0
42	Hookers Hot Spring	D-13-21-6AAC	52.0	37.0
43	Agua Caliente Spring	D-20-13-13BA	27.0	11.0
44	Antelope Spring	D-20-24-21DC	25.5	10.5
45	Monkey Spring	D-21-16-3C	28.3	13.3

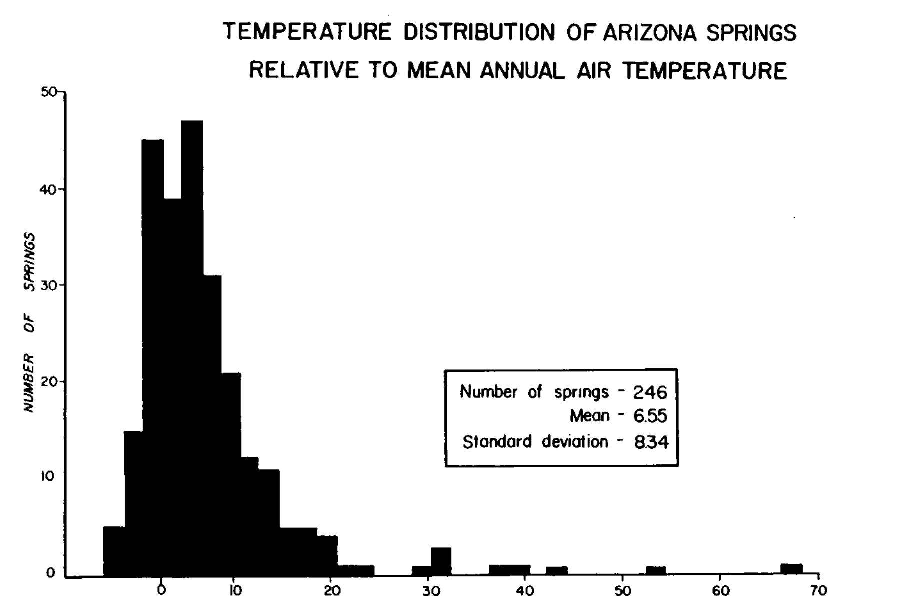

Figure 1. SPRING TEMPERATURE MINUS MEAN ANNUAL AIR TEMPERATURE (°C)

THERMAL SPRINGS OF ARIZONA

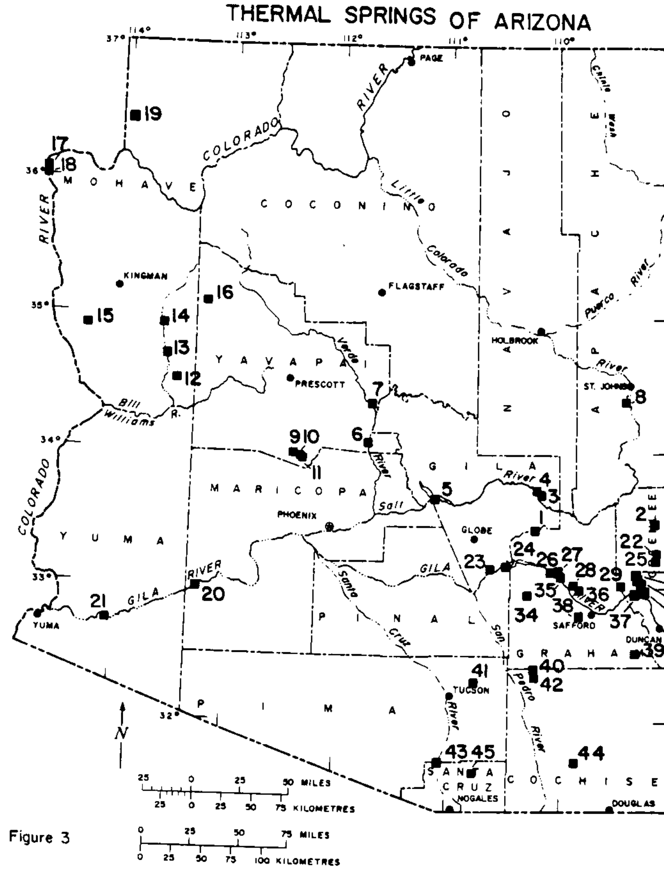

Figure 3

A COMPARISON OF TEMPERATURE DISTRIBUTION OF SPRINGS IN THE COLORADO PLATEAU AND BASIN AND RANGE PROVINCES

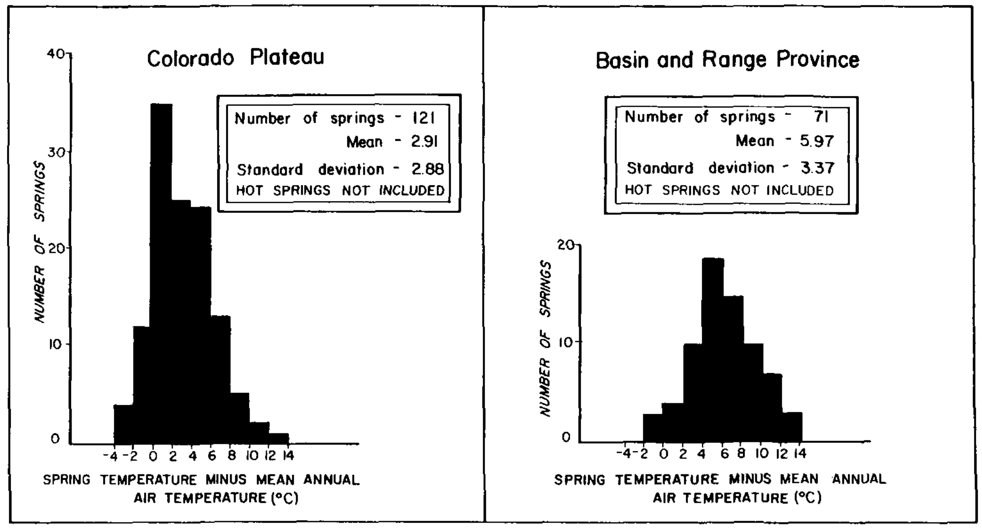

Figure 2.

MERCURY SOIL SURVEYS: A GOOD RECONNAISSANCE TOOL

Mercury geochemical soil sampling has proven to be a useful tool in determining subsurface geologic structure in the Basin and Range province of Arizona. Limited studies indicate that mercury vapor leakage along buried geologic structures is reflected in anomalously higher mercury concentrations in soil samples above the structures. The mercury anomalies reflect subsurface structures that are usually identified by gravity and seismic surveys. Therefore, when feasible, mercury soil sampling can easily and economically be substituted for geophysical surveys during a geothermal reconnaissance assessment in undisturbed basins in the Basin and Range province.

The mercury soil samples are generally collected at one-mile spacings at section corners, but have also been collected at half-mile and quarter-mile intervals. Sample spacing depends upon the size of the project area and what is to be defined by the survey. If one-mile spacings are inadequate, additional sampling at closer intervals is easily accomplished to improve resolution and coverage.

The sampling technique is simple and fast. Soil horizons, as developed in moist or wet climates, do not exist in the Basin and Range province of Arizona. Therefore, when taking the sample the first four or five inches of soil and organic material are removed, and two or three handfuls of bare, mineral soil are scooped into a plastic, ziplock sandwich bag. The sample bag is carefully sealed and labelled.

The samples are sent to a commercial laboratory where they are air dried inside the laboratory, sieved to the minus-80-mesh fraction, and analyzed. Air drying of the samples in the laboratory is important as sunlight tends to drive off the mercury vapor.

It is important to sample the same depth zone at each sample site since the results of a mercury survey are only relative. It is also important that the samples are kept away from direct sunlight, when stored for any length of time. We statistically estimate the mean mercury concentration for a given area and use all values greater than the mean plus one standard deviation to indicate an "anomalously" high value.

Three examples of mercury soil surveys are discussed, along with the gravity data. In the Avra Valley study, the gravity indicates a rather large alluvial-covered structure trending northwest from the southeast corner of the map (Fig. 1). The results of the mercury soil survey mirror this structure. The stippled area on the map indicates the geochemically anomalous area. These two linear patterns (gravity and mercury) define the Silver Bell-Bisbee discontinuity (Titley, 1976), a major mineralized lineament active in the Laramide and mid-Tertiary (Rehrig and Heidrick, 1976). In addition Lepley (1978) was able to define portions of this lineament using satellite imagery.

Figure 2 is from the San Pedro Valley study. In this area the gravity reflects the thick sequence of volcanic rocks in the basin that apparently mask most of the geologic structures. The results of the mercury soil survey also do not reflect geologic structures in the basement. Again, there is good correlation between the gravity data and the reconnaissance mercury soil survey.

The third survey site was in the northern Hassayampa Plain. The mercury soil survey (Fig. 3) reflects the northwest-trending range-bounding fault, as well as north-south and northeast-trending structures. At one-mile spacings the mercury soil survey clearly indicates gross subsurface structures suitable for a reconnaissance assessment. The survey provided realistic structural targets for the gravity survey to refine. Gravity data in this area were so sparse that it was necessary to run a gravity survey (Fig. 4). This more refined survey reveals a sinuous pediment edge that suggests en echelon(?) faulting. The north-south and northwest-trending structures seen in the mercury survey are also shown by the gravity survey.

We do not use mercury soil concentrations to identify geothermal anomalies for two reasons: (1) high mercury concentrations are also indicative of fossil geothermal systems; and (2) the high incidence of placer gold mining operations that used mercury for amalgamation have possibly contaminated large areas in the Arizona Basin and Range province. In both instances high mercury concentrations would erroneously suggest geothermal anomalies.

In summary, we have found an excellent correlation in southern Arizona between buried structures revealed by gravity and mercury soil surveys. The latter type of survey has several advantages over the former as a reconnaissance tool: (1) location and elevation control are not critical; (2) nearly anyone can be trained to collect samples; (3) inexpensive; (4) fast; and (5) easy.

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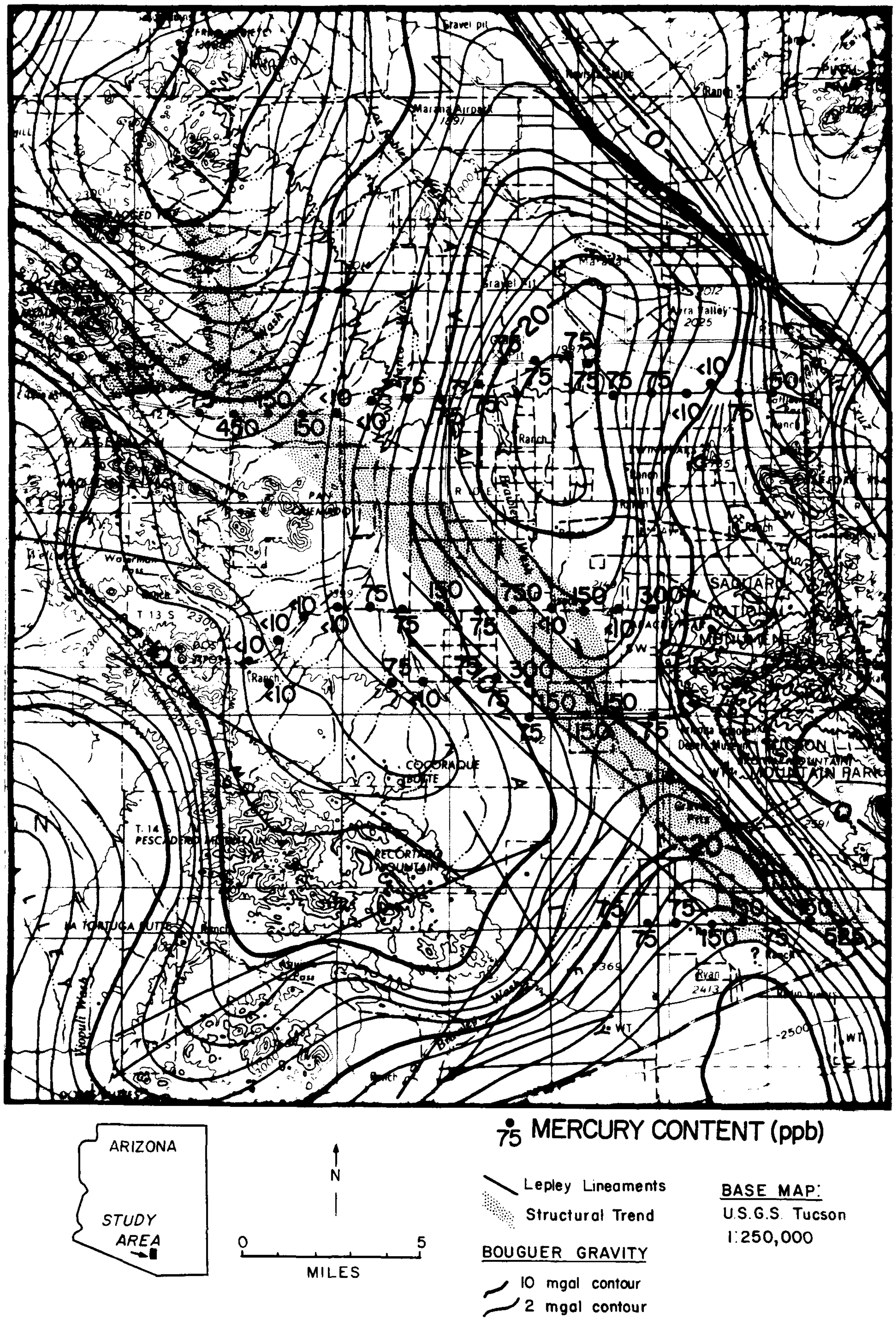


Figure I. Mecury soil survey, Avra Valley, Arizona.

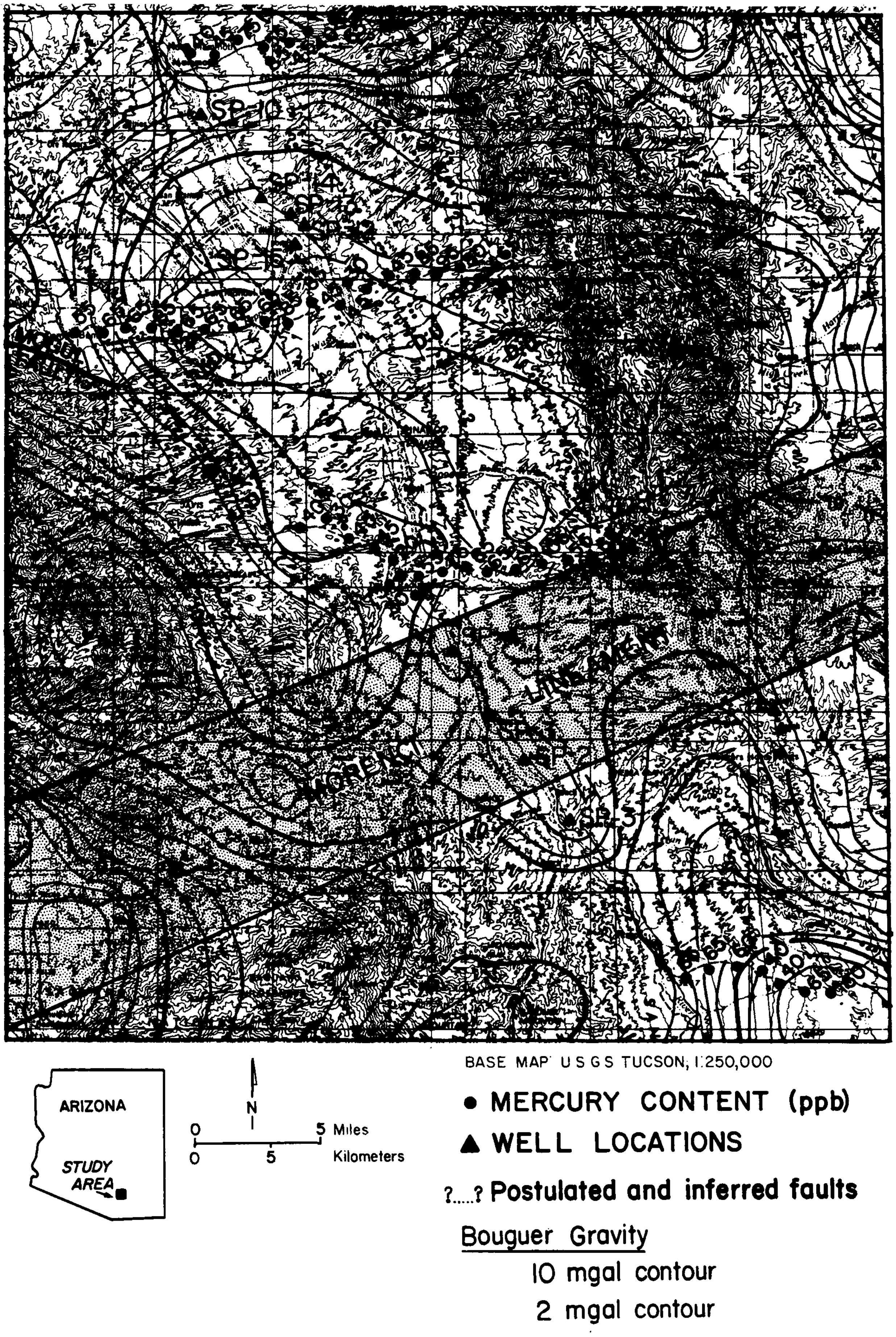
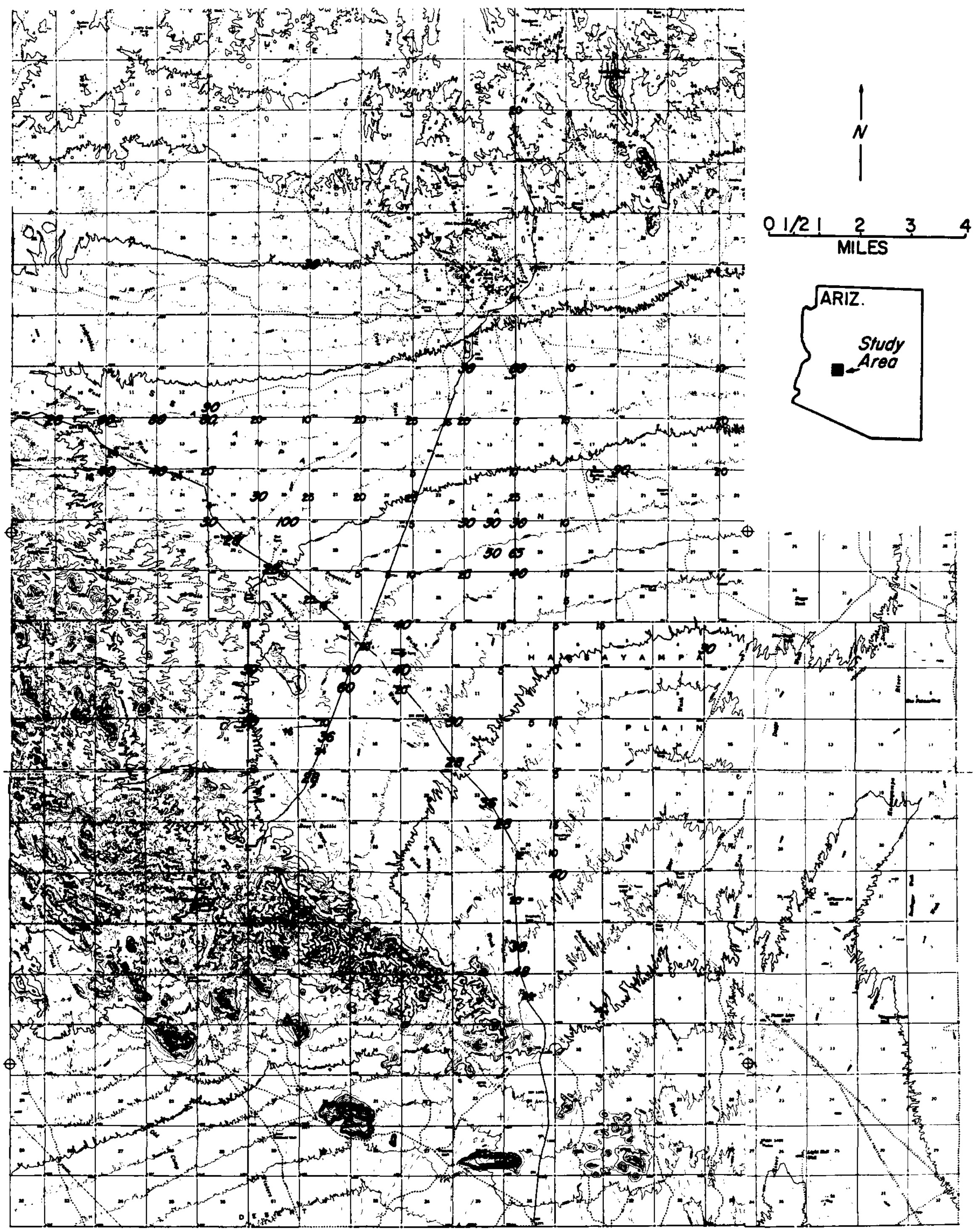


Figure 2. Mecury soil survey, San Pedro Valley, Arizona.



Map showing mercury concentration (ppb) in soil samples

Figure 3. Mecury soil survey, Hassayampa Plain, Arizona.

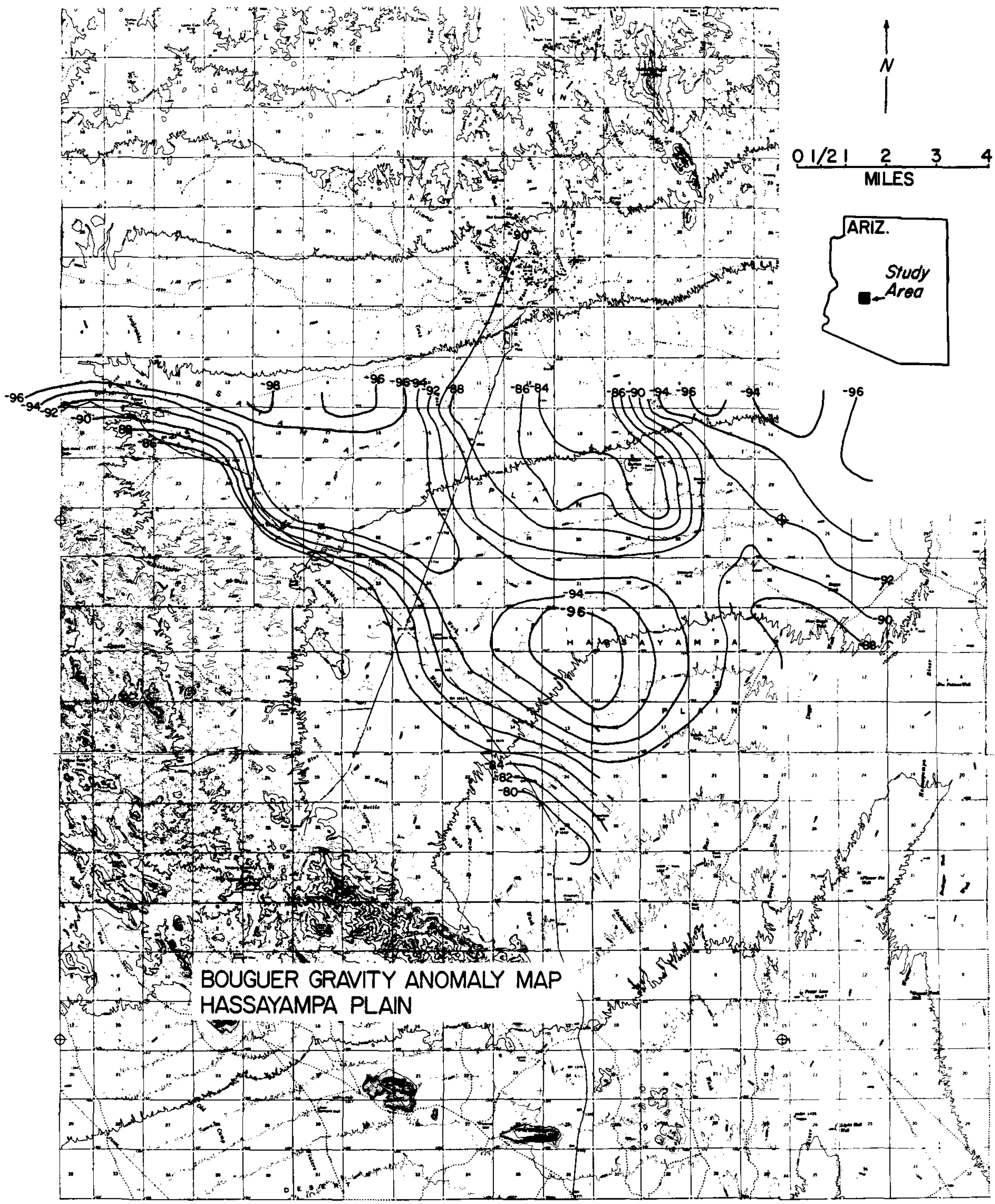


Figure 4. Bouguer gravity anomaly map, Hassayampa Plain, Arizona.

CALIFORNIA DIVISION OF MINES AND GEOLOGY
GEOTHERMAL STUDIES IN LOS ANGELES COUNTY AND IN CALISTOGA, NAPA COUNTY, CALIFORNIA

Two of the most interesting geothermal assessments conducted by the California Division of Mines and Geology, as part of the U.S. Department of Energy's State Coupled Program, include studies of the Calistoga area completed in the 1979-80 project year and of the Los Angeles County area, in progress in the 1980-81 project year. The following are brief summaries and comments on the findings in each of the two areas.

Calistoga Area

The Calistoga, California area contains a hydrogeothermal field exceeding 20°C, covering about 15 square kilometers of valley lands which are geologically evaluated to determine the extent and magnitude of the heat resources. The geological study was conducted by the California Division of Mines and Geology in 1980, with funding provided by the Federal Department of Energy and the State of California.

Geological, geophysicial, and geochemical techniques were used in the area, and these were subsequently supplemented by a small drilling program. Three holes were drilled using the dual tube drilling technique with the deepest hole reaching a final depth of 885 feet. The dual tube method has, as distinct advantages over ordinary drilling techniques, the ability to pinpoint aquifers down to a few inches in thickness and to obtain uncontaminated samples of both water and rock types as drilling proceeds.

The investigations revealed that as many as seven distinct aquifers contribute heated water to this field. The aquifers extend vertically through an average 85 meter thick sequence of permeable alluvial sands and gravels, interbedded with impermeable clay and volcanic ash units. The top of the uppermost thermal aquifer averages about 35 meters below the ground surface. The maximum temperature recorded from the reservoir is 135°C., which supports the estimated maximum temperature of 140°C. that was derived by the Division using geothermometric calculation. Geophysical results indicate the possible location of a heat source for this area, and a possible 15 square kilometer extension of the field, in the upland area to the southwest of the valley.

The Calistoga area contains a small community associated with the wine industry of Napa Valley, and it is noted for its spas and bottled mineral water. Direct heat applications of hydrogeothermal fluids have long been made here for residential and commercial uses. The results of this study will assist in the further efficient development of the field while permitting the maintenance of the geothermal energy source, and will facilitate development of the field with minimal environmental degradation.

Los Angeles County Area

The 73 oil fields of Los Angeles County were studied by the California Division of Mines and Geology under contract to the U.S. Department of Energy to determine if waters produced from oil fields can be considered as sources of thermal energy for non-electrical applications adjacent to the fields. The oil fields lie in two areas, the Los Angeles Basin and the Ventura Basin. The Los Angeles Basin has more potential for applications because of its large population and greater number of fields; most fields are in densely developed commercial-industrial or residential areas which could provide numerous applications.

Data from a canvass of 40 oil field operators in the county indicates that large quantities of warm-to-hot water are produced from fields in the south and central portions of the Los Angeles Basin. Temperatures range from ambient air to boiling, but most are in the 35°-60°C range. The temperatures largely reflect the normal geothermal gradient -- the increase of temperature with depth -- rather than an anomalous heat source under the basin. Production is mainly from depths of about 1000-2000 meters; water quality ranges from about 5,000 to 40,000 mg/liter total dissolved solids.

Many operators believe that extraction of heat from the waters is technologically feasible, although several cite economics and detrimental effects on oil viscosity as possible hindrances. Most operators agree that the most favorable and least disruptive sites for heat extraction would be the water-collection and treatment facilities for each field. The facilities are generally centralized in the fields and could serve as points of distribution for the heat produced.

A SUMMARY OF D.O.E. FUNDED GEOTHERMAL RESOURCE ASSESSMENT EFFORTS IN COLORADO BY THE COLORADO GEOLOGICAL SURVEY

by

Richard Howard Pearl, Ted. G. Zacharakis, and Frank R. Repplier
Colorado Geological Survey
Denver, Colorado.

INTRODUCTION

For the past 3 1/2 years the Colorado Geological Survey has been evaluating the geothermal resources of select thermal areas in the State. Efforts have been directed towards defining the geology, hydrogeology and geothermal characteristics of those thermal areas deemed to have immediate development potential. The areas evaluated were selected in conjuction with the Colorado Commercialization Project, which is also a part of the Colorado Geological Survey offices.

The resource assessment program of the Colo. Geol. Survey has been a fully integrated exploration and assessment program consisting of: geological mapping where necessary but primarily the compilation of existing geological mapping data; hydrogeological mapping data; soil mercury surveys; gradient drilling; reservoir confirmation drilling; and geophysical surveys such as electrical resistivity, telluric, AMT, and seismic. Not all of the above have been run at each area. Following is a brief description of the work performed at each individual area.

AREAS EVALUATED

A) Pagosa Springs, located in southwestern Colorado (Fig. 1). Geological and hydrogeological mapping, seismic and dipole-dipole geophysical surveys, gradient drilling, reservoir confirmation drilling, and soil mercury surveys.

B) San Luis Valley, south central Colorado (Fig. 1). Wide variety of surveys in different parts of the Valley.

1. Shaws Springs area on the west side of the Valley (Fig. 1). Compilation of previous geological mapping; seismic, electrical resistivity, AMT and Telluric geophysical surveys and soil mercury surveys.
2. Central part of the Valley north of Alamosa (Fig. 1). Gradient hole drilling.

C) Canon City Embayment area (Fig. 1). Program was designed to locate a source of thermal waters for the Dept. of Corrections prison complex. Soil Mercury surveys; gradient drilling; compilation of previous geological mapping and seismic, electrical resistivity, AMT and telluric geophysical surveys.

D) Idaho Springs, 30 miles west of Denver (Fig 1). Area evaluated primarily due to efforts of the Commercialization Team. Soil mercury survey; electrical resistivity survey and compilation of previous geological mapping plus reconnaisance geological mapping.

E) Glenwood Springs, located approximately 150 miles west of Denver on the Colorado River (Fig. 1). Seismic and dipole-dipole geophysical surveys.

F) Hartsel Hot Springs in South Park west of Colorado Springs (Fig. 1). Electrical resistivity geophysical survey; soil mercury survey and compilation of previous geological mapping.

G) Ranger Hot Springs north of Gunnison (Fig. 1). Area still under evaluation. Following work either has been done or will be done during the summer of 1981: electrical resistivity geophysical survey; soil mercury survey and geological and hydrogeological mapping.

H) Animas Valley north of Durango in southwestern Colorado (Fig. 1). Soil Mercury; electrical resistivity survey and compilation of existing geological mapping with field checking and revision where necessary.

I) Ouray on north side of San Juan Mountains (Fig. 1). Area evaluated due to efforts of the Commercialization Team. Electrical resistivity surveys; soil mercury surveys and compilation of existing geological mapping.

J) Hot Sulphur Springs in northcentral Colorado (Fig. 1). Area still under evaluation. Soil mercury surveys have been run and geological appraisal completed. Electrical resistivity surveys to be run during summer of 1981.

K) Steamboat Springs in northwest Colorado (Fig. 1). Area still under evaluation. Soil mercury surveys completed. Electrical resistivity survey and a hydrogeological thesis will be done during the summer of 1981.

L) Wagon Wheel Gap on the east side of the San Juan Mountains (Fig. 1). The area is still under evaluation, with electrical resistivity surveys to be run during the summer of 1981.

In addition the following items have been completed or are in preparation.

A) Revision of the Heat Flow Map of Colorado. To be published spring 1981

B) Evaluation of lineament structures of Colorado as noted on satellite imagery and their relation to thermal springs of Colorado.

C) Map of groundwater temperatures in Colorado.

D) Geothermal gradient map of Colorado constructed from maximum temperatures recorded in oil and gas wells.

E) Thermal mine water drainage will be evaluated during summer of 1981. Project being done at request of Commercialization Team.

F) Waunita Hot Springs east of Gunnison. AMAX recently dropped their geothermal leases in the area and released all their exploration maps and reports to the public. The Colorado Resource Assessment Team will publish these maps and reports.

SUCCESSES AND FAILURES

During the course of the investigations listed above a number of successes and failures were encountered. Following is a brief description of them.

Successes

A) Geophysical surveys: In most instances the geophysical surveys ran were successful. However in several important instances serious problems were encountered. They will be discussed in detail later in the paper.

AMT and Telluric geophysical surveys were conducted by the U.S. Geological Survey in the San Luis Valley and the Canon City Embayment areas. The surveys obtained good results and data; However for a complete understanding of the data it has to be used in conjunction with other geophysical surveys and subsurface geological mapping.

Electrical resistivity surveys: The Colorado Geological Survey has a Syntrex RAC-8 electrical resistivity system. This is a good system but it does have some limitations, the primary being that it is limited in the depth that it will penetrate. If this limitation is taken into consideration during the planning of the project then it does not become a serious problem. In most of the areas where the equipment was used, good results were obtained.

B) Gradient measurements: The Colorado Resource Assessment Team now has two deep hole temperature probes available. A Fluid Dynamics System was acquired because of its accuracy and portability. Good results have been obtained using this system. The other system is truck mounted and upon use the probe was found to be not very accurate. Modification of the probe with the assistance of the Wyoming Resource Assessment Team has resulted in measurements of greater accuracy.

C) Bottomhole temperatures from oil and gas wells: Due to the great number of oil and gas test wells drilled in Colorado and the fact that copies of all electric logs run have to be filed with the Colo. Oil and Gas Conservation Commission this program is progressing very well. Data was obtained from all sedimentary basins in the State. While there are obvious inherent errors in using this data it is felt that it will present a good general picture of the geothermal gradients in Colorado.

D) Groundwater temperatures: This is another program that has progressed very well. The Colorado Geological Survey was able to acquire copies of the U.S. Geol. Survey WATSTORE data bank tapes. These tapes, plus other data sources provided temperatures for approximately 7,000 water wells in Colorado. Field verification of anomalous temperatures was required in some instances. This data will be used to construct a map which will be useful for users of groundwater heat pumps.

Failures

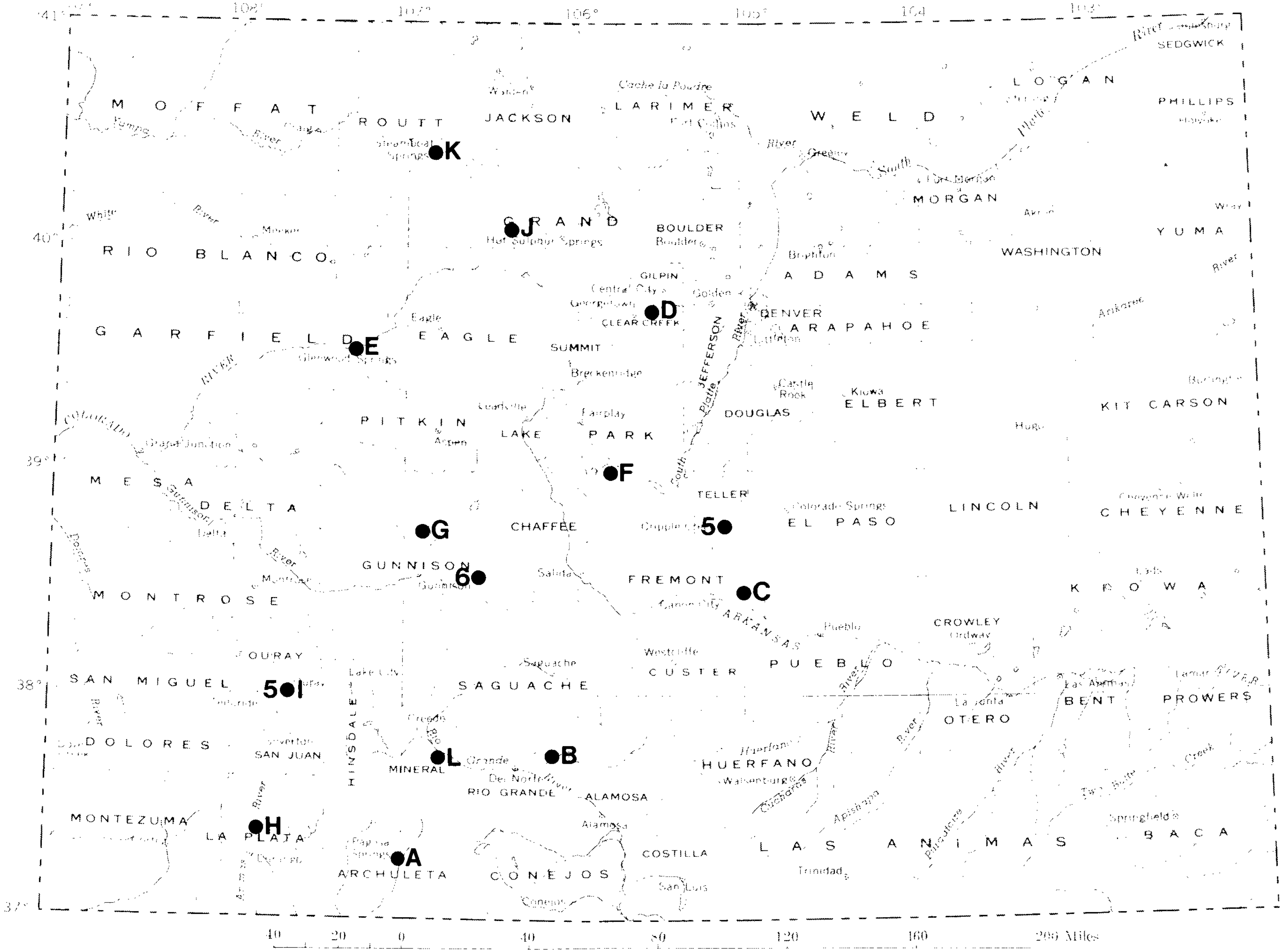
A) Soil Mercury: Probably the single greatest disappointment encountered by the Colo. Assessment Team has been the failure of soil mercury geochemical sampling techniques. Researchers at the Univ. of Utah Research Inst. and Jerome Inst. have been successful in showing that the mercury content of soils in geothermal areas is a viable geothermal exploration technique. Based on the above, the Colorado Resource Assessment Team obtained a Jerome Inst. Gold Film Mercury analyzer in 1979 and initiated a program of sampling in select thermal areas. Using sampling and analysis techniques developed by previous workers, The Colo. Team, working in areas where thermal waters exist and other areas where there were no thermal waters, was able to locate only a couple of geothermal anomalies in two summers of use and these were in close proximity to thermal waters. In attempting to obtain viable results a number of sampling methods were employed; various traverses and grid profiles were used. Soil profiles were run with samples collected based on these profiles, in other instances samples were collected at specified intervals, also random samples collected, or samples were collected from two or more closely spaced holes. No matter what methods were employed in a thermal area, usually only a single anomalous value would be obtained, and that from the vicinity of the hot spring. Quite obviously the method it is not proving to be of any value when only a known thermal spring can be located. After reexamination of previously published data and our results, the Colo. Team came to the conclusion that the geothermal systems in Colorado are either to old or too cold. Perhaps soil mercury surveys are not applicable in the old, cooler systems found in Colorado.

B) Geophysics: Another disappointment encountered was the inability to obtain good seismic geophysical data along the west side of the San Luis Valley. In the San Luis Valley basaltic lava flows and other erruptive materials are close to the surface. A contract was entered into with Geophysics Fund, Inc., the consulting geophysical group from the Colorado School of Mines to run seismic geophysical surveys in the San Luis Valley. Even though Geophysics Funds Inc. has had extensive experience working in a volcanic rock terrain due to the adsorpative nature of the volcanic rocks they were unable to acquire any good quality reflective data. The Colorado Team has extensively discussed this problem and no conclusion thus far has been reached on how this seismic problem could have been resolved. The contractor was chosen for his expertise and it is assumed that proper techniques were employed.

CONCLUSION

Analysis has shown that the geothermal resources of Colorado are primarily low to moderate temperature hydrothermal resources. As such their use will be limited to direct use applications. To aid potential developers of these resources the Colorado Geological Survey, in cooperation with the U.S. Dept. of Energy, in 1977 initiated a program to fully evaluate the geological and hydrogeological environment of those thermal areas in Colorado which have a high development potential. This program has been a fully integrated resource assessment program, consisting of: Geological and hydrogeological mapping, geophysical and geochemical surveys and gradient drilling and measurements.

Even though this program has had mixed results it is believed that valuable information on the occurrence and geological conditions of the geothermal resources of Colorado has been obtained. Upon conclusion of this program in 1982 those areas considered to have high development potential will have been evaluated, to some degree. This information should be of aid to potential developers.



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Hot Sulphur Springs
Glenwood Springs
Grand Junction
Aspen
Leadville
Breckenridge
Fairplay
Central City
Georgetown
Golden
Littleton
Boulder
Brighton
Fort Collins
Greeley
Fort Morgan
Castle Rock
Kiowa
Hugo
Colorado Springs
Salida
Gunnison
Canon City
Pueblo
Westcliffe
Ordway
Saguache
Creede
Del Norte
Alamosa
Walsenburg
Trinidad
San Luis
Conejos
Springfield
La Junta
Las Animas
Lamar
Delta
Montrose
Telluride
Silverton
Lake City
Durango
ARKANSAS
South Platte
Cache la Poudre
White
River
Rio Grande
Apishapa
Purgatoire
40
20
0
40
80
120
160
200 Miles

HAWAII GEOTHERMAL RESOURCE ASSESSMENT PROGRAM:
1980 GEOPHYSICS SUBPROGRAM

Several new studies were begun in 1980; these include a microearthquake survey on Maui and detailed gravity and magnetic coverage of small areas on Maui and Hawaii. The major effort of the geophysics subprogram is still the development and application of electrical resistivity techniques. To this end, modeling and inversion computer programs were either written by us or were obtained from other groups and adapted to our needs. Reports covering the major surveys of Maui and Hawaii, in which electrical methods were used predominantly, have recently been written and are now being reviewed. Also in 1980, two small-scale EM techniques were evaluated for a new application as regional geothermal reconnaissance tools.

Work was not restricted to geophysics. As an aid to those who wish to study large amounts of groundwater chemistry data that may or may not be very precise, some simple statistical discrimination procedures were applied to the Hawaii groundwater data. Results are presently incomplete; however, a set of geothermal discriminators have been identified that both confirm previous ideas and also suggest new relationships.

I. Microearthquake Location Mapping

During the period from July to September 1980, a preliminary microseismic monitoring survey was conducted on the island of Maui. The objectives of this survey were to both conduct a field test of the recently modified microprocessor controlled seismic packages as well as to attempt to locate areas of anomalous seismicity on Maui. A total of eleven seismic instrument packages were deployed on Maui which provided sufficient coverage to detect any event near the island having a magnitude of 2.5 or greater.

All of the seismic packages worked reliably throughout the field test, however, late delivery of high stability oscillators required for accurate determination of arrival times precluded the precise location of more than a few of the detected seismic events. Figure 1 presents a map of the locations of the stations that provided reliable times as well as the locations of three seismic events which occurred during this survey. It is interesting to note that one of the three events is located close to Ukumehame canyon which is also the site of a significant resistivity anomaly. Another significant result may be the absence of seismic activity in the Makena-La Perouse Bay area, which is the location of the most recent (1790) eruption on Maui. However, a longer period of coverage is clearly necessary in both of these areas before it will be possible to draw any significant conclusions concerning their thermal potential.

II. Gravity and Magnetic Mapping

Detailed gravity and magnetic field surveys were conducted on all the rift zones of interest on Maui as well as in the Kawaihae area on the island of Hawaii. Over 450 gravity stations and 10 miles of magnetic profile (at 30-meter intervals) were obtained. Detailed analysis of these data are presently under way, however, several preliminary interpretations can be drawn from the initial results.

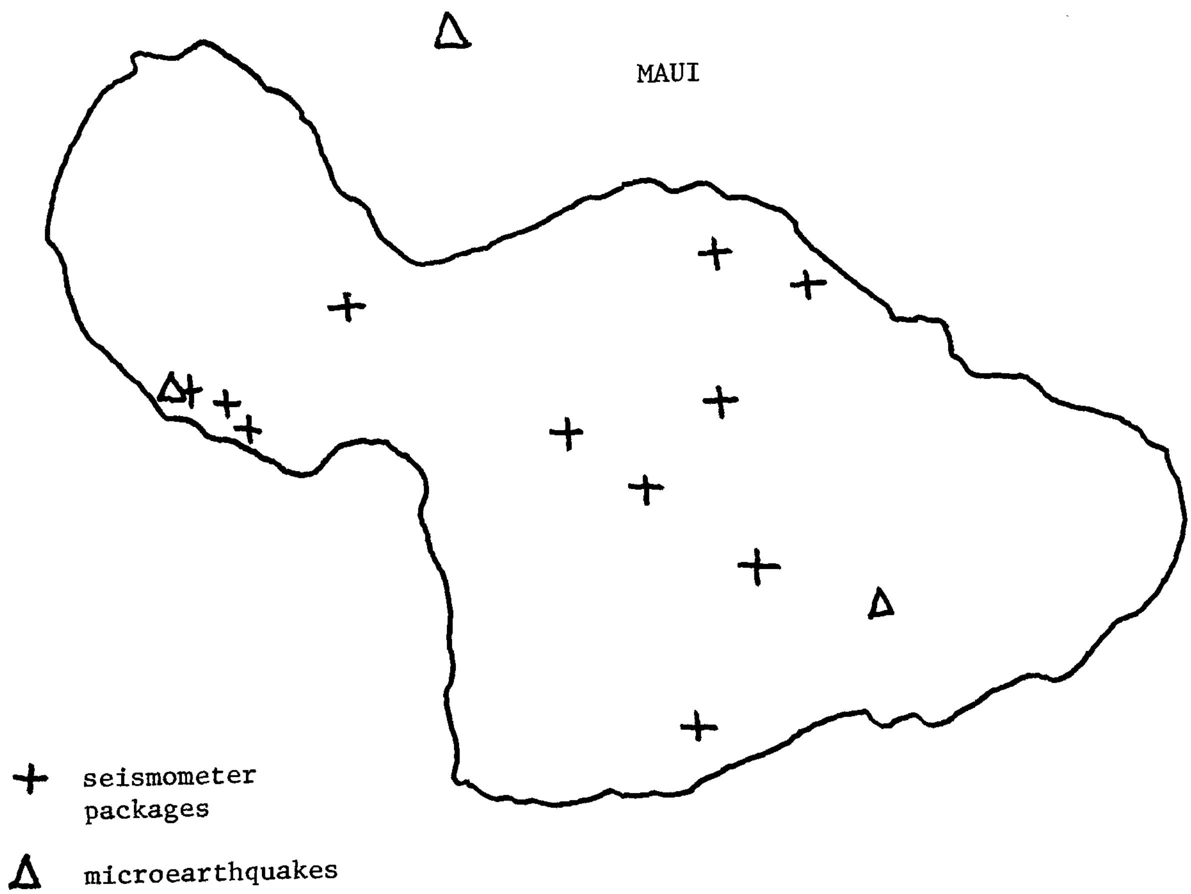


FIGURE 1

The first is that the gravity data obtained are consistent with the results of an earlier, less detailed statewide survey, although the more recent data have also delineated additional small-scale features associated with the rift zones. Modeling of these features should provide valuable information on the subsurface density structure near the rift zones.

Analysis of the magnetic data has been substantially more difficult due to large disturbances in the local magnetic field arising from near surface features such as highly-magnetized boulders or small-scale topographic effects. However, in several instances, signals have wavelengths as long as 1 km have been recognized. Work is in progress to filter out the near surface effects so as to model the deeper structures.

III. Computer Software Development

(1) Three-dimensional Resistivity Program

The program "RES3D" obtained from the U. C. Berkeley group was successfully adapted to run on the department's HARRIS computer. Since this program requires large amounts of CPU time (up to 1 hour/model), the low cost of the HARRIS makes it possible for us to experiment with several models of geothermal interest. Work in 1980 was concentrated on testing convergence criteria using standard one-dimensional models. However, the program is now performing satisfactorily for the purpose of computing Schlumberger apparent resistivities.

(2) One-dimensional Schlumberger Forward Problem

A program was adapted to run on a TI 58 hand calculator, making it possible to calculate apparent resistivities with reasonable accuracy in the field. A BASIC program was also adapted to run on the new APPLE III microcomputer. This program also provides a graphical output, making it extremely useful for checking the results of the different inversions described below.

(3) Schlumberger Inversion

Three separate inversion routines are now running on the department's HARRIS computer. The first, "MARQDCLAG" was written by Walt Anderson of the U. S. Geological Survey. The second, "SLUMB", was obtained from the Utah Geothermal group at UURI. The third program was adapted from a magnetotelluric inversion routine written by B. Lienert at HIG. This last routine has several significant advantages over the two routines already mentioned. The first is that it converges rapidly, even when no starting model is given. Secondly, it gives quantitative information on the amount of resolution which the data is capable of providing. This is extremely important, especially when problems of equivalence are encountered in thin layers.

An example of an inversion of data obtained on one of the Maui transects is given in Figures 2 and 3.

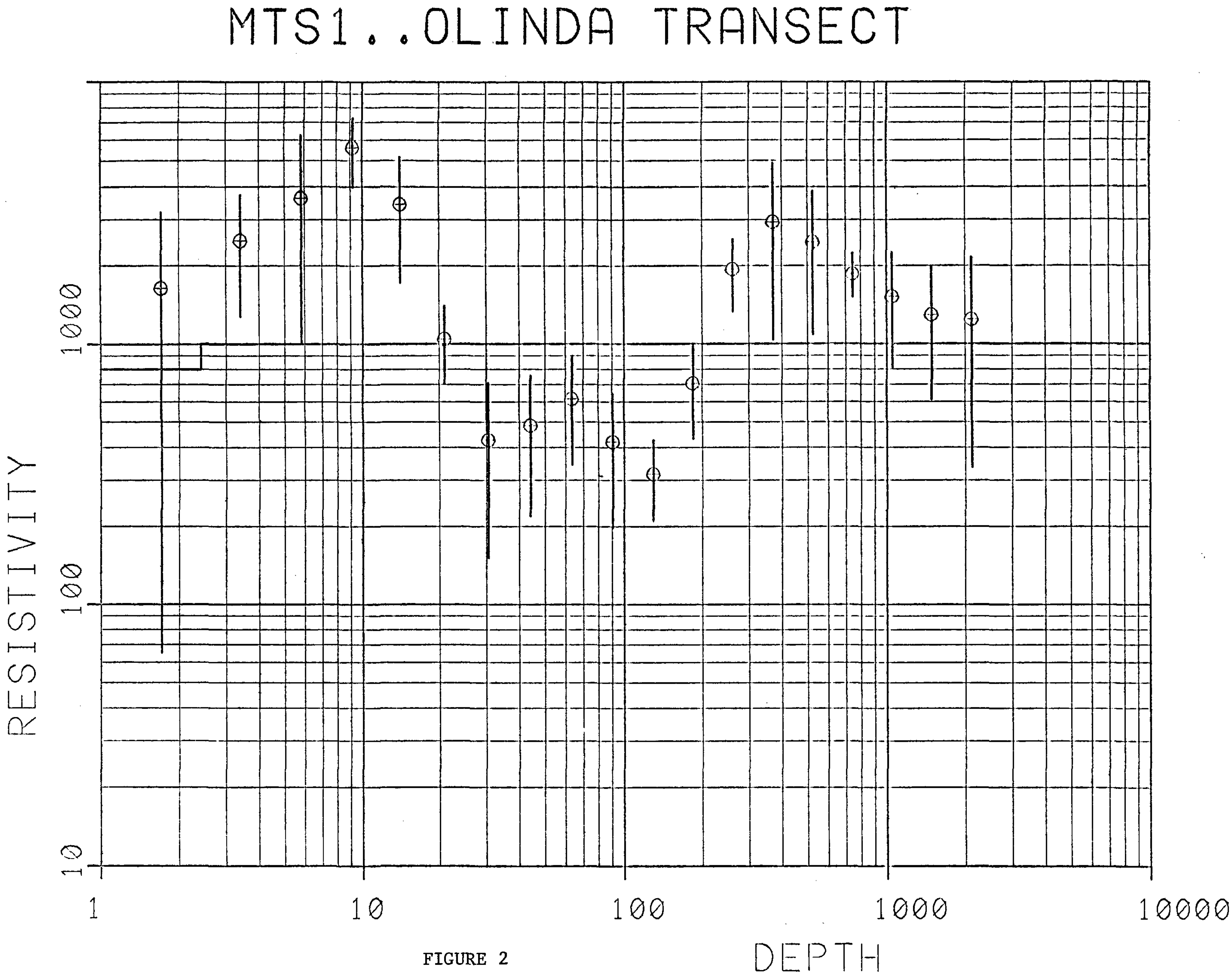

MTS1..OLINDA TRANSECT
RESISTIVITY
1000
100
10
1
10
100
1000
10000
DEPTH

FIGURE 2

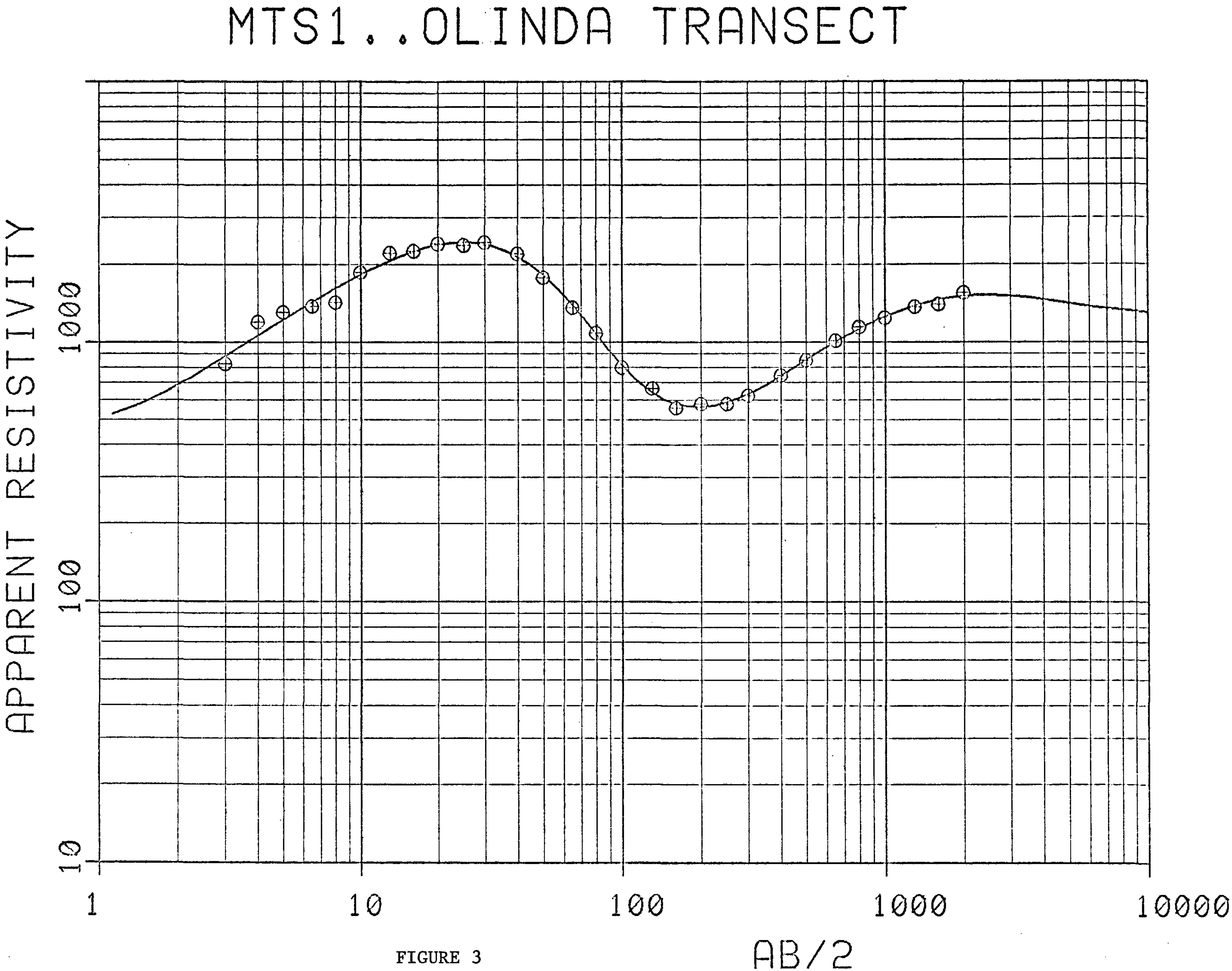

MTS1..OLINDA TRANSECT
APPARENT RESISTIVITY
1000
100
10
1
10
100
1000
10000
AB/2

FIGURE 3

IV. D. C. Resistivity Sounding on Maui

The coverage achieved in 1979 was extended with soundings in the Haiku area on Maui. Two more soundings were also performed in the vicinity of Ukumehame Canyon. The results of the Ukumehame soundings confirmed the low resistivity values obtained in 1979. This area is undoubtedly the most promising one on the basis of the resistivity data.

Late in 1980 two soundings were performed on wilderness transects which were used by the U. S. Fish and Wildlife for a survey of native bird populations. These two soundings formed a good trial for the practicality of performing resistivity soundings in remote areas where vehicular access is impossible. Although the soundings were both performed successfully, we concluded that they would prove extremely expensive and time-consuming as a reconnaissance technique, and should only be attempted when considered absolutely vital. Although neither of the soundings were able to penetrate down to sea level, useful data was obtained on the resistivity structure of the upper kilometer in two areas which are almost certainly representative of large portions of East Maui.

V. Electromagnetic and Resistivity Sounds on Hawaii

A technical report entitled "Geophysical Evaluation of Prospective Geothermal Areas on the Island of Hawaii Using Electrical Methods" has been written and is presently under review. The following is a summary of the findings in six exploration areas on Hawaii:

(1) Kawaihae

The strongest evidence for subsurface heat is the high temperature in water wells in the area. Most are between 26 and 28°C; one well between the towns of Kawaihae and Waimea has a temperature of 37°C. The goal of our investigations was to discover the source of this heat. Previously gathered geologic, electric, and magnetic data pointed to the Kohala volcanics to the north and east of the warm-water well. The youngest lava flow in this area has been dated by K/Ar methods to be 80,000 years old.

Additional electrical soundings around the warm-water well yielded only a little additional information, mostly due to the method's inability to penetrate the water-saturated rocks below sea level; however, one sounding just east of the warm-water well revealed a possibly resistive horizon at depths which should have appeared conductive. This sounding was done over a large, complex magnetic anomaly. Our evidence is only suggestive; however, delineation of an anomalously magnetized, anomalously resistive body in this particular region would essentially be delineation of an intrusive body and a possible source of heat.

Deep EM soundings would answer many questions about the Kawaihae area regarding the location and size of the heat resource. TDEM soundings were tried, but failed because of problems in obtaining large enough currents in a grounded-wire source. The ash covering most of this area is very fine and dry and electrical grounding is very difficult in it. Future EM work should probably be done with an ungrounded source.

(2) Hualalai

The only evidence for subsurface heat here is circumstantial; the volcano has erupted in at least two locations along its northwest rift in the past 200 years. Hualalai is a rather large exploration target - fortunately, previous studies have been completed which effectively rule out the north and southeast rift zones as geothermal resources, at least to depths of 2 km. Our efforts were concentrated on the summit and northwest rift zones; the work consisted of D.C. electric and TDEM soundings.

Moderately low resistivities were discovered beneath the lower portion of the northwest rift at depths between 300 m and 3 km. The actual resistivity values are about twice as high as those found in the Puna KGRA suggesting that this may not be a high temperature resource. Low resistivities were also discovered beneath the summit region, but these may only reflect high-level groundwater that may or may not be warm. Our results are inconclusive on this point.

(3) South Point

Preliminary evidence for heat is only slightly more convincing for this area than it was for Hualalai; the southwest rift zone of Mauna Loa runs through the middle of this area and has erupted numerous times in the past 200 years. The vent locations for these eruptions have generally moved uprift or away from the coast in this time. Aerial infrared surveys have detected warm water flowing into the ocean where the rift intersects the coast and have outlined thermal anomalies on a west-facing fault scarp within the rift.

A previous study to the east of this fault scarp did not detect any anomalously low resistivities. Further information obtained from electric soundings farther uprift also did not detect resistivity anomalies. At this time, prospects for a geothermal resource in this area are not good; however, the area to the west of the fault scarp has not yet been explored.

(4) Southwest Rift of Kilauea

Again, the many eruptions in the past 200 years suggest that residual heat may be present in this area. Warm water is known from only one coastal spring. Previous studies have mapped the boundaries of a high-level water body which borders this rift zone. From this work, it is known that water lies 60 m above sea level to the west of the westernmost edge of the rift zone and that water lies at less than 20 m above sea level to the east of this boundary.

Reevaluation of electrical data taken in a previous field season shows that this edge of the rift structure is the westernmost edge of a very low resistivity area. The resistivity values are comparable to those observed in the Puna KGRA and are probably due to rocks saturated with hot, brackish water. The prospects of a moderate-to-high temperature resource appear good in this area.

(5) Keaau

Extensive coverage of this area by electrical and TDEM soundings show that subsurface resistivities are high. The values are compatible with those for rocks saturated with cold seawater. Prospects for a geothermal resource of any temperature are not good.

(6) East Rift of Kilauea

Although this area has been extensively studied in part years and already has been successfully drilled, a few more electrical soundings were obtained to better delineate the lateral and vertical boundaries of the heat resource. After compilation of all electric and hydrologic data on this KGRA, the area can be separated into three areas of geothermal potential. First, the areas north of the northernmost edge of the rift are expected to yield only normal temperature fluids. The area to the south of the rift's southern edge and the area within the rift downrift of HGP-A test well are expected to yield high temperature fluids at depth; fluid temperatures are expected to exceed 200°C at depths less than 1 km below sea level. Finally, the area within the rift, but uprift of HGP-A, may also have high temperature fluids probably at depths greater than 1 km.

VI. Evaluation of VLF and EM Loop-Loop Profiling as Tools for Rapid Geothermal Reconnaissance in Hawaii

Of all the electrical geophysics techniques used thus far in Hawaii, none is both portable enough for one or two men to operate and carry on foot, and powerful enough to measure water-saturated rock resistivities through several tens of meters of dry rock. Yet, such a technique would be invaluable for covering large areas in enough detail to quickly determine where one might concentrate more expensive exploration methods. The method cannot be one involving direct current (DC) principles; experience and theory have shown that such methods require a minimum of three people and electrode spacings of about five or six times the maximum elevation in volcanic terrain. EM methods are more promising. Therefore, to fill this gap, the VLF (Very Low Frequency) and EM loop-loop (EMGUN or SLINGRAM) methods were evaluated theoretically and subsequently tested in selected field areas.

Low frequency EM energy is much more sensitive to low resistivity materials than those having high resistivities. One can show theoretically that EM techniques are almost totally affected by depth to and the resistivity of a low resistivity rock layer when its overburden is a few hundred times more resistive.

Both methods were used in several areas on the islands of Maui and Hawaii with mixed results. The loop-loop technique never worked well at all; readings were generally noisy which precluded quantitative analysis for saturated-rock resistivities. The VLF technique worked the best, yielding fairly reliable estimates of saturated-rock resistivities at elevations up to 50 m. Use of both techniques on Maui were complicated by the interference of low-resistivity ash interbedded with lava flows. VLF should still be useful in the drier terrain of Maui, although the results will have to be interpreted carefully.

VII. Application of Statistical Analysis to the Determination of Geothermal Indicators: Hawaii Groundwater-Chemistry Data

Both univariate and bivariate statistical procedures have been applied to the 388 well water analyses compiled by HGRAP to deduce the combination of well parameters which would best be used to discriminate between waters that are thermally affected and those that are not. Details of three individual aspects of the study are outlined.

(1) Analysis of Variance (ANOVA) of Data Grouped by Temperature

The 388 sets of data were split into two groups based on present water temperature and ANOVA on 45 variables was computed. The 45 variables include 7 physical parameters, 29 chemical parameters, and 9 chemical ratios. We were testing the null hypothesis that "the average value of each of the 45 variables for groundwaters at temperatures above the threshhold temperature is equal to the average value of the variables for groundwaters at temperatures less than or equal to the threshhold temperature". The following is the list of variables for which the null hypothesis could be rejected at 90% confidence for two values of the threshhold temperature:

Threshhold = 25°C	Threshhold = 30°C
pH	selenium (Se)
alkalinity	silica (SiO_2)
selenium (Se)	sodium (Na)
silica (SiO_2)	chloride (Cl)
zinc (Zn)	potassium (K)
lead (Pb)	calcium (Ca)
calcium (Ca)	copper (Cu)
bicarbonate (HCO_3)	Mg/Cl
copper (Cu)	
Mg/Cl	
K/Cl	
SO_4/Cl	
Na/Cl	
Ca/Cl	
Na/K	

With the exception of pH and all the ratios except Na/K, all variables are higher in waters above threshhold than in waters below threshhold temperatures.

The silica and Mg/Cl relationships had already been suggested on theoretical grounds. Na, Cl, K, and Ca are all major constituents of seawater and their presence in this list is a reflection of the larger amount of seawater observed in most high temperature wells. Cooper and selenium had not been considered before, and may be promising discriminant variables.

This aspect of the study is viewed as successful because it confirms all relationships proposed for thermal discrimination, as well as discovering two new discriminators.

(2) Correlation Coefficient Matrix:

Correlations have been computed for all pairs of the 45 physical, chemical, and ratio variables. The strongest relationships are seen between the major ions in seawater - Na, Cl, SO_4, K, and Mg. Analyses of these coefficients are continuing.

(3) Factor Analysis:

Rotated principal component analysis (R-mode) was performed on the 279 analyses which had non-zero values for the following variables: temperature, HCO_3, SiO_2, Ca, SO_4, Mg, Na, Cl. The three largest factors and their loadings are listed below.

Variable	Factor 1	Factor 2	Factor 3
Cl	0.940		
Na	0.937		
Mg	0.912	0.274	
SO_4	0.889		
Ca	0.853		
SiO_2		0.880	
HCO_3	0.444	0.641	
temperature			0.956
explained variance	61%	12%	11%

Loadings less than 0.25 are not shown. The three factors explained 84% of the total variance. Factor 1 is obviously seawater effects, Factor 3 is temperature effects, and Factor 2 is probably return-irrigation-water effects. At the present stage of analysis, departures from the 3-factor model may be more interesting than the factors themselves, as far as geothermal indicators. Factor analysis isolates the major components of variance; however, a geothermal indicator is probably a very small source of variance to the overall set of data.

GEOTHERMAL EVALUATION OF KANSAS - PRELIMINARY RESULTS

148-164

by

Don W. Steeples

Kansas Geological Survey
The University of Kansas
Lawrence, Kansas 66044

June, 1981

INTRODUCTION

A low-temperature geothermal resource investigation was begun in Kansas in 1979. This paper should be considered a progress report of that investigation, so the results and speculations should be considered preliminary in nature. The individual facets of the study are discussed individually after the outline of the geologic and tectonic framework.

GEOLOGIC AND TECTONIC FRAMEWORK

The Midcontinent portion of North America is the most stable part of the continent, tectonically. The region is relatively aseismic, and there has been no significant deformation of the crust since at least the Late Paleozoic. The major tectonic elements of this region are (1) the southern extension of the Central North American Rift System (Midcontinent Rift; Midcontinent Geophysical Anomaly; Figure 1); (2) the Nemaha Ridge; and (3) the Central Kansas Uplift.

The Central North American Rift System [Ocola and Meyer, 1973; Chase and Gilmer, 1973] can be traced from central Kansas across southeastern Nebraska, Iowa, and Minnesota to its outcrop area in the Great Lakes region. The rift is marked by pronounced gravity and magnetic anomalies [King and Zietz, 1971; Lyons, 1950; Thiel, 1956] and is underlain by mafic igneous rocks, mostly basalt and gabbro, and arkosic sedimentary rocks. The feature is generally regarded as an abortive continental rift which occurred about 1100 m.y. ago [Goldich et al., 1961; Silver and Green, 1963, 1972; Goldich et al., 1966; Chaudhuri and Faure, 1967; Van Schmus, 1971].

The Nemaha Ridge is a striking tectonic feature which was intermittently active during Paleozoic time. It is certainly a major crustal fracture zone, for mylonitized basement rocks have been brought up from within it, and cataclasis is a common feature along its extent from northeastern Kansas into Oklahoma [Bickford et al., 1981]. The fault zone is upthrown on the western side, forming the feature known as the Nemaha Ridge. The eastern flank of the Nemaha Ridge is bounded by the Humboldt Fault Zone [Steeples et al., 1979]. Earthquakes as large as Modified Mercalli Intensity VII have occurred along the Humboldt Fault Zone in historic time [DuBois and Wilson, 1978].

The Central Kansas Uplift (Figure 1) is a broad region in which basement rocks have been moved upward and which is characterized by fault zones and cataclasis. The feature is evidently coextensive with the Cambridge Arch in Nebraska. Although the Central Kansas Uplift was active during the Paleozoic, little is known about its Precambrian history. A relatively high level of microearthquake activity (more than 20 events per year larger than magnitude 1) occurs along this structural trend [Steeples, 1980].

The crystalline crust in the Midcontinent is buried under about 1000 m of sedimentary rocks and is thus mostly known from studies of numerous drill holes [Muehlberger et al., 1966; Goldrich et al., 1966; Lidiak et al., 1981; Kisvarsanyi, 1980]. The crust in this area is notable for its predominantly granitic composition. Mafic rocks are rare, and metamorphic rocks, though present in many places, are not abundant. A major feature of the crystalline crust in the Midcontinent is its division into a northern terrane, consisting of somewhat deformed and sheared granitic rocks and lesser amounts of metamorphic rocks that occur in northern Missouri, northern Kansas, and Nebraska, and a southern terrane totally dominated by silicic volcanic rocks and associated epizonal granitic plutons. The southern terrane can be traced from northern Ohio across Indiana, Illinois, southern Missouri, southern Kansas, and Oklahoma into the Texas Panhandle. Geochronological studies [Bickford et al., 1981; Denison et al., 1981] indicate that the northern terrane is generally older, with many rocks yielding ages of 1640 m.y. (U/Pb, zircon) to 1740 m.y. (Rb-Sr), whereas the southern terrane varies in age from about 1475 m.y. in the St. Francois Mountains of southeastern Missouri [Bickford and Mose, 1975] to about 1380 m.y. in southwestern Missouri, southeastern Kansas, and Oklahoma [Bickford and Lewis, 1979; Bickford et al., 1981].

Lying upon the crystalline crust in the Midcontinent region is a section of sedimentary rocks ranging from about 150 m in thickness over parts of the buried Nemaha Ridge to as great as 2 to 3 km thick in basins such as the Hugoton Basin of southwestern Kansas and northwestern Oklahoma. The average thickness of the sedimentary rock section in eastern Kansas where our drilling projects were done is about 1 km. The rocks range in age from Late Cambrian to Pennsylvanian or Permian in eastern Kansas, but there is a thick Cretaceous section in central Kansas, and rocks of Tertiary age occur on the western plains. Paleozoic rocks in the Midcontinent region are mostly marine in origin and are dominated by carbonate units and shale.

PIGGYBACK DRILLING

Normal exploration procedure for most resources involves drilling as a culmination of geologic and geophysical investigations. It would seem that drilling four holes in an area the size of Kansas at the outset of a regional geothermal evaluation is a reverse approach.

However, the drilling was done as a "piggyback" operation at a relatively low cost compared to the total project. The data obtained from the boreholes has enabled us to provide high quality heat-flow data for Kansas and to better evaluate thermal information available from the petroleum drill holes.

THE DRILLING OPPORTUNITY

In 1976 a cooperative program between the U.S. Geological Survey and the Kansas Geological Survey was begun. The purpose of the study was to determine the regional geohydrologic characteristics of the Arbuckle Group to include definition of flow patterns in the Arbuckle and in relation to other overlying units, determination of hydraulic parameters in the various units, and determination of regional chemical quality. Test data procured from oil exploration companies are being analyzed to accomplish this objective.

Additional funding to the Kansas Geological Survey became available in FY 1979 and FY 1980 for the purpose of test drilling and installation of deep monitor wells. This funding was matched and increased by the U.S. Geological Survey, the U.S. Army Corps of Engineers, and the Kansas Department of Health and Environment; the project was expanded to include determination of hydrologic properties of the Arbuckle and other units by drilling at specific locations.

We were not involved professionally in any aspect of the Arbuckle project when it was originally funded. However, we realized that valuable petrologic and geophysical data could be obtained if these holes could be deepened by about 100 m to penetrate the crystalline basement. Four drilling sites were selected. Three of the four sites were selected so that, in addition to the hydrologic study, basement rock samples and geophysical data could be obtained from the same holes. Two of the sites were located above intense, circular magnetic anomalies; the third was located in a region where sparse well control indicated a terrane of silicic volcanic rocks in the basement.

We received funding from LASL and the U.S. Department of Energy (DOE) to deepen the first two holes, recover core from the basement, and perform high-quality heat flow measurements in the holes. Funding was also received from a separate LASL contract to deepen the third hole; however, severe circulation problems developed within the Arbuckle Formation before the "piggyback" experiment could begin. All of the money from that contract was returned to LASL.

Cost of Drilling Operations

A contract was awarded in September 1979 in the amount of $444,701 for work to be performed at the first three sites. A second contract, in the amount of $129,341 for work to be performed at the fourth site, was awarded in April 1980. Seventeen drill-stem tests were completed for the purpose of determining the hydraulic relationships between the Arbuckle and other major overlying aquifers. Three cores of the Arbuckle were collected in addition to the basement cores, and these are currently being analyzed under a third contract. A complete suite of geophysical logs, including vertical flow determinations, were completed at each site. Also, acoustical televiewer photographs were taken over selected intervals at three sites. Water samples have been collected at three sites and are undergoing complete chemical analyses, including determinations for age dating with lithium, bromine, strontium, deuterium, carbon 14, and tritium.

The increased incremental cost of approximately $9,000 per hole for coring in the Precambrian section of two holes is included in the above figures. Most of the preliminary scientific results reported here are a direct result of this small additional expenditure, thus demonstrating the value of "piggybacking." Less than $30,000 of the total geothermal funding was used for the heat flow studies and the "piggyback" drilling.

SCIENTIFIC RESULTS FROM DRILLING

The author had significant input as to the location of the holes, and their sites were chosen to maximize potential information from the basement, subject only to the general suitability of the location to the primary mission of the drilling project, i.e., the hydrologic study of the Arbuckle. The legal descriptions and locations of holes drilled are given in Table 1.

Drilling at the first hole (Miami County) was completed on December 10, 1979. Approximately 8 m of 6.7-cm-diameter core of fresh granite were recovered from a depth of 658 to 666 m. This hole was located on a sharp 1000-γ circular aeromagnetic high, shown as locality 1 on Figures 1 and 2.

The second hole (Douglas County; locality 2 on Figures 1 and 2) was also located on a circular magnetic high with an amplitude of about 1100-γ; drilling was completed on March 19, 1980. Three meters of 10-cm-diameter core of fresh granite were recovered from a depth of 905 to 908 m. The 3 meters represent only 58 percent recovery of the 5.2 meters cored. We were very fortunate not to lose all of the core, as it started slipping out of the core barrel during the trip up the hole. The core catcher barely hooked the core again and prevented disaster. We were not charged for the core that was lost.

Two additional holes (localities 3 and 4 on Figures 1 and 2) were drilled to depths of 1117 m and 554 m, respectively. Severe lost-circulation problems developed on both of these holes within parts of the Arbuckle Formation, and drilling was halted at that depth because the primary objective of the drilling had been met. Penetration of Precambrian basement at sites 3 and 4 would have cost an additional (possibly very large) undetermined amount of money.

The scientific data we ultimately expect to obtain from the drill core and from the geophysical measurements include the following: age, petrography, major and trace element chemical composition, density, and remanent magnetism of the rocks encountered; heat flow; and heat production of the rock material. The holes into basement can be made suitable for hydrofracturing experiments to measure in situ stress, provided future funding becomes available. The holes will be available to other scientists for other experiments within 2 years. Interested individuals should contact the author of this report.

Some preliminary data are available on age, thermal gradient, and heat flow. The age of the Precambrian cores is about 1350 m.y. (U/Pb of zircons) [Steeples and Bickford, 1981] indicating that the circular magnetic anomalies represent a suite of intrusions younger than the "normal" 1650 m.y. age for the crust in the area.

TABLE 1. Legal Description of Drill-Hole Locations in Kansas

	Location	Total Depth
Douglas County	SE 1/4 NW 1/4 NW 1/4 Sec..13, T12S, R17E	908 m
Labette County	Center of SE 1/4 Sec.22, T31S, R20E	553 m
Miami County	SE 1/4 SW 1/4 SE 1/4 Sec.18, T18S, R23E	666 m
Saline County	SW 1/4 SW 1/4 SW 1/4 Sec.32, T13S, R2W	1117 m

Geothermal Gradients

Preliminary thermal logging has been performed on all four holes by personnel from David Blackwell's laboratory at Southern Methodist University. The thermal logging equipment was not capable of reaching the bottom of the holes, so these data should be considered preliminary, pending results from deeper logging. Samples of core or well-cuttings have been sent to Blackwell's laboratory for thermal conductivity measurements. The following geothermal gradients have been measured to date in the four holes drilled on this project:

Location	Gradient	Depth Logged
Douglas County	30.3°C/km	565 m
Labette County	28.5°C/km	520 m
Miami County	36.0°C/km	395 m
Saline County	30.7°C/km	565 m

Preliminary data from Blackwell indicate an unusually high rate of radioactive heat generation, about 11 heat generation units, in the core from the Miami County hole, compared with the 5 to 6 heat generation units for typical granites. The heat generation from the core obtained from the Douglas County hole was not anomalously high.

High quality heat flow measurements have been made in the four drill holes [Blackwell et al., 1981]. The average heat flow in the holes was 54 ± 5 milliwatts/square meter (1.3 ± 0.1 heat flow units). These heat flow values are typical for areas of stable Precambrian-aged crust. Additional results will be available after the holes are logged to total depth.

OTHER STUDIES IN PROGRESS

The aeromagnetic map of Kansas (Figure 2) was funded in part with geothermal money. A preliminary interpretation of that map has been published by Yarger (1981). Gravity data are being gathered to supplement the interpretation of the aeromagnetic map. A gravity map contoured at one milligal intervals will be available for the eastern half of Kansas by mid-1982.

We are continuing to measure geothermal gradients statewide and will complete that phase of the field work in time to publish a geothermal gradient map in 1982. Preliminary indications are that geothermal gradients state wide are consistently in the range of 30 to 35°C/km.

Geothermal Speculations for Kansas

Figure 3 shows structural contours on the Precambrian surface relative to sea level. There are areas on this map that bear discussion with respect to low grade geothermal prospects. They will be discussed in counter-clockwise order around Kansas starting with the southeast corner of the state.

1. In the southeastern portion of the state, dip is toward the northwest away from the Ozark Uplift. Sedimentary cover is roughly 500 meters thick with relatively good quality water in much of the sedimentary section. The Arbuckle group of early Ordovician age is used as a fresh water source by several towns. The water temperature is in the range of 25° to 35°C, warm enough to be used for heat pump applications for space heating. This area is the most likely portion of Kansas to have any geothermal applications in the next 5 to 10 years. The geothermal gradient below depths of 500 meters is quite low - possibly in the range of 15 to 20°C/km.

2. The Nemaha Ridge is a buried granitic mountain range that reaches to within 200 meters of the surface in northeastern Kansas. Geothermal gradients in the sedimentary section exceed 50°C/km, but geothermal prospects are poor because the sedimentary section is so thin. The high geothermal gradient is not thought to persist below the bottom of the sediments.

3. To the west of the Nemaha Ridge is the Midcontinent Geophysical Anomaly (Figure 1). In Figure 3 the MGA is more or less bounded on the southeastern flank by six kimberlite intrusions of Cretaceous age. The northwest flank of the MGA is bounded by a zone of microearthquakes that have occurred since 1978. These boundary features along opposite sides of the MGA indicate structural zones that may allow circulation and convection of water deep into the crust.

   The MGA itself is caused by mafic intrusive and extrusive rocks of late Precambrian age as discussed earlier. Surrounding the MGA is an arkosic Precambrian age sandstone (Rice Formation) that developed during the later stages of rifting and subsidence. The thickness of the Rice Formation is unknown, but probably substantial. Modeling of the gravity and magnetics by Yarger (1981) indicates that the edges of the Rice Formation are probably fault-bounded. It is possible that the Rice Formation is several kilometers thick, based on differential arrival times at two seismograph stations in the vicinity of the MGA.

   A deep seismic reflection profile experiment is being performed across the MGA by the Consortium for Continental Reflection Profiling (COCORP) during 1981. The results of this experiment will probably allow calculation of the thickness of the Rice Formation.

   If there are, indeed, several kilometers of Rice Formation present, geothermal prospects for production of large volumes of warm water would be excellent. The quality of any such water is unknown.

4. The Central Kansas Uplift (Figure 1) is much the same as the Nemaha Ridge from a geothermal standpoint. However, thousands of oil wells produce water at temperatures of about 40 to 45°C in conjunction with oil production. The heat from this water is not purposefully extracted prior to reinjection in salt water disposal wells. Other than this by-product warm water, the geothermal prospects are not good.

5. In northwestern Kansas, the Dakota Formation may be locally useful for heat-pump applications. Will Gosnold (1981 personal communication) has discovered that the Dakota waters in western Nebraska are unusually warm, apparently as a result of convection updip to the east from the Denver-Julesburg Basin. It is not yet clear whether this same effect is present in Kansas, but it will be investigated in the near future.

6. In southwestern and south-central Kansas, the sedimentary section is 2 to 3 km thick. Waters produced in conjunction with petroleum are at temperatures of 60 to 65°C. Water quality is generally poor and the prospects for geothermal development are not bright because of sparse population density and lack of industry in the area.

FIGURE CAPTIONS

FIGURE 1. Principal positive structural features in Kansas. Major basins include the Forest City Basin in northeast Kansas, the Cherokee Basin in southeast Kansas, the Salina Basin in north-central Kansas, and the Hugoton embayment of the Anadarko Basin in southwestern Kansas. Drill sites for heat flow measurements are shown chronologically by numbers 1, 2, 3, and 4.

FIGURE 2. Aeromagnetic map of Kansas. Contour interval is 50 gammas. Drill sites from Figure 1 are shown as 1, 2, 3, and 4.

FIGURE 3. Precambrian structural contours of Kansas with 1000 foot contour interval relative to sea level. Stars in northeast Kansas denote kimberlite locations. Faults and microearthquakes show locations where crustal fractures are present.

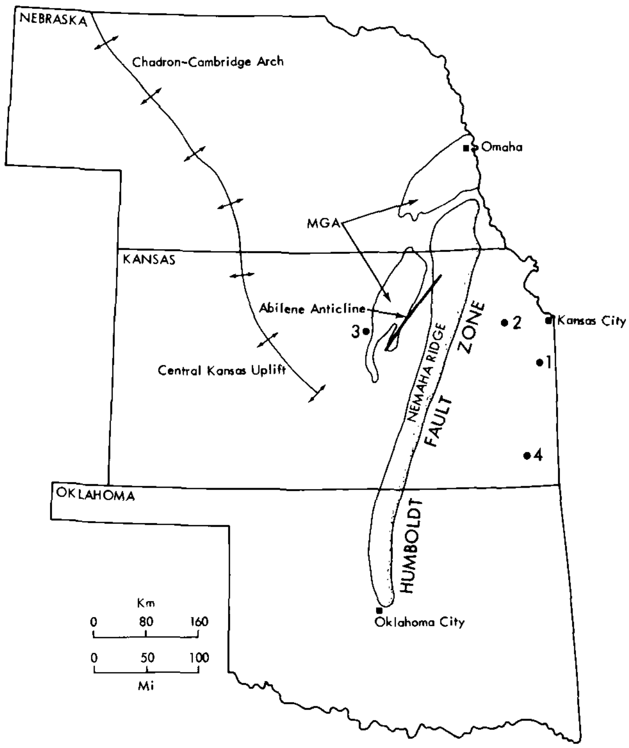

NEBRASKA
Chadron-Cambridge Arch
Omaha
MGA
KANSAS
Abilene Anticline
3
2
Kansas City
1
Central Kansas Uplift
NEMAHA RIDGE
ZONE
FAULT
4
OKLAHOMA
HUMBOLDT
Km
0
80
160
0
50
100
Mi
Oklahoma City

FIGURE 1.

AEROMAGNETIC MAP OF KANSAS

H. Yarger, R. Robertson, J. Martin, K. Ng, R. Sooby and R. Wentland

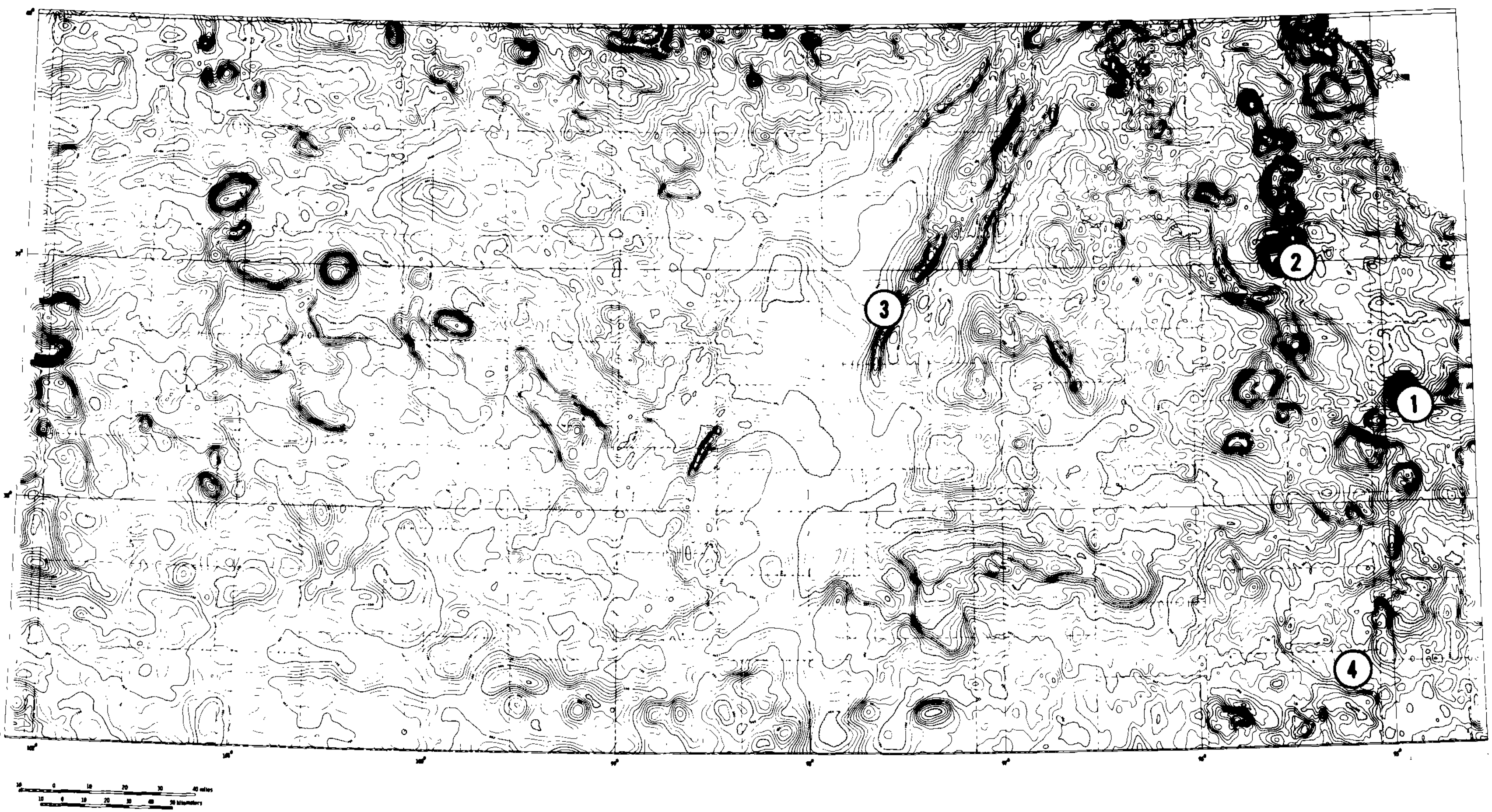
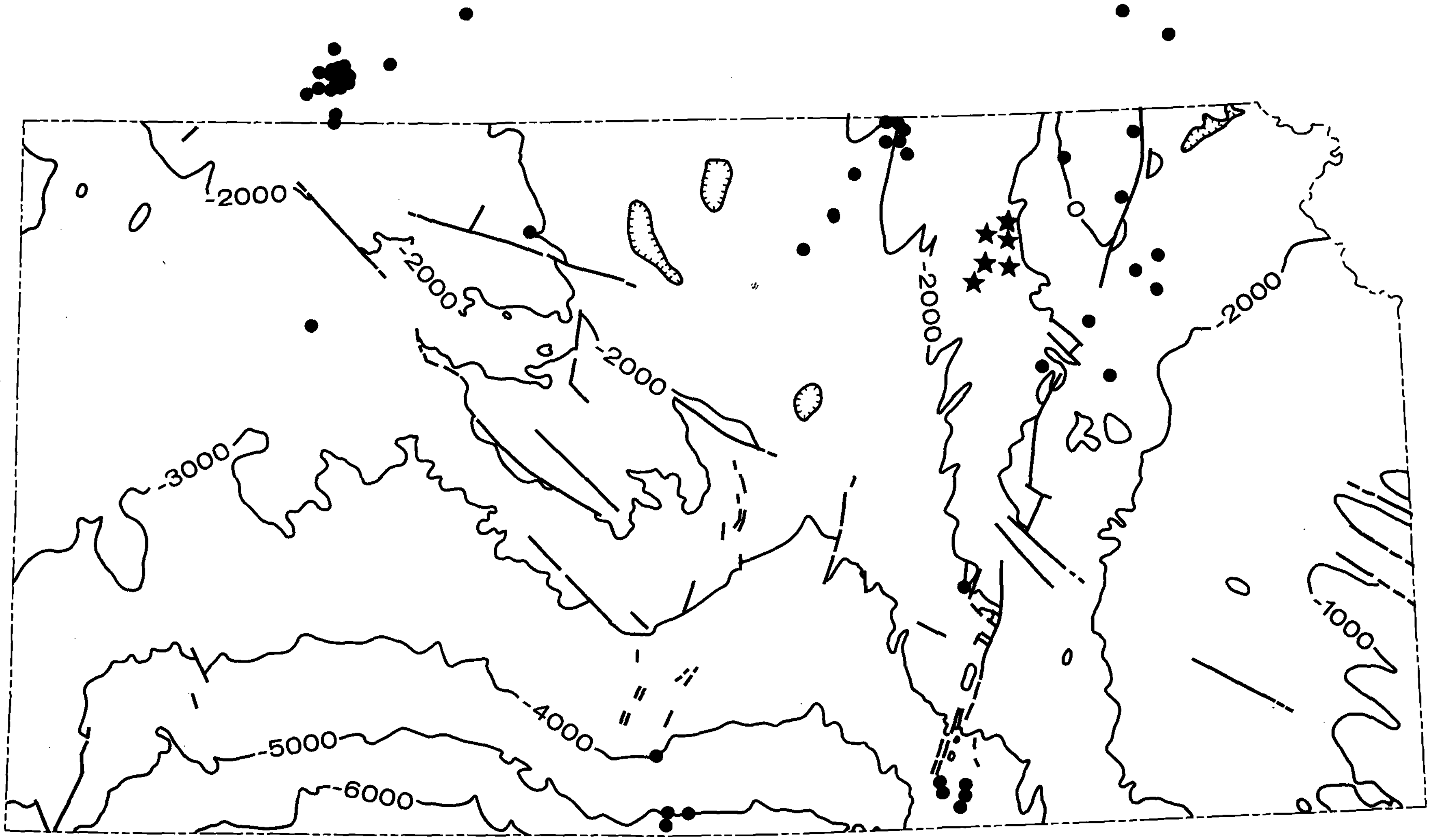


FIGURE 2.



/ Faults
• Microearthquakes
S Precambrian structural contours

FIGURE 3.

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GEOTHERMAL RESOURCES IN MONTANA

J. L. Sonderegger and F. A. Schmidt
Montana Bureau of Mines and Geology, Montana College
of Mineral Science and Technology, Butte, Montana 59701

Abstract --A list of persons and groups doing geothermal research in Montana is presented. A revised list of springs and wells with their flow and temperature values is shown with the heat value, in billions of British Thermal Units (Btu's) per year, for reference temperatures related to low temperature uses. The Boulder and Hunters springs are the foremost hot spring resources, while the Madison Limestone related springs around the Little Rocky Mountains, and Brooks spring north of Lewistown provide the major low temperature resources capable of large development utilizing heat pump technology. The water chemistry of almost all springs is suitable for direct application. A discussion of drilling activities around spring sites and the relative success (or lack thereof) provides some factors to consider. In an attempt to delineate areas with ground-water temperatures suitable for heat pump use, a 10°C (50°F) temperature cutoff was used. Urban area data is suspect; inadequate pumping time may yield spuriously warm temperatures.

The purpose of this paper is to summarize the work done to date, and to report on some recent results relating to Montana's geothermal resources.

Interest in surface occurrences of thermal water as something other than scientific or "medical" curiosity did not become prominent until the early 1970's when predictions of energy shortfalls began appearing. In Montana, previous work consisted of cataloguing by G. A. Waring (23), and "while passing through" studies by S. L. Groff (results summarized in 3); also, Balster (2) compiled a map using bottom-hole temperatures in the Madison Group.

Recent research was initiated by the U.S. Geological Survey in the early seventies from their Menlo park regional office. The formation of first the U.S. Energy Research and Development Agency (ERDA) and then the U.S. Department of Energy (DOE) broadened the federal research base and provided funding for state and private research projects. The following list includes most of the Montana-based groups performing geothermal research (either in resource assessment or in engineering applications):

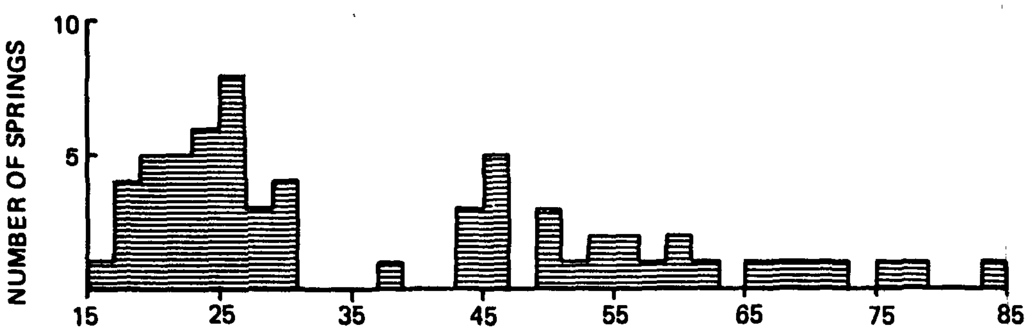
1. U.S. Geological Survey, Montana WRD Office, Helena, Montana: Robert Leonard--resource evaluation.
2. Department of Natural Resources and Conservation, Division of Renewable Energy, Helena, Montana: Michael Chapman--user assistance and grants.

3. Montana University System
   a. University of Montana, Missoula: Tony Quamar--resource evaluation
   b. Montana State University, Bozeman: Robert Chadwick--resource evaluation
   c. Montana College of Mineral Science and Technology, Butte: John Sonderegger and Charles Wideman--resource evaluation
4. Fort Peck Tribal Research Program, Poplar, Montana: Carl Fourstar--resource definition and application (near Poplar)
5. Montana Energy Research and Development Institute, Butte, Montana: Karen Barclay--resource definition and application (Warm Springs State Hospital)

THERMAL SPRINGS

Because warm and hot springs represent an expression of a geothermal system at depth, an inventory of such springs has traditionally been the first step in evaluating the resource potential. One of the problems recognized in the mid 1970's was that adequate measurements of spring discharge and temperature were not always available (at a given temperature, the energy available is directly proportional to the spring discharge) normally because of poor discharge numbers which often varied by as much as 400 percent. In the fall of 1975, Robert Leonard was assigned to the USGS Montana district; after reviewing the Montana Bureau of Mines and Geology (MBMG) spring data files, Leonard decided to restrict his work to occurrences of waters hotter than 100°F in the southwestern portion of the state. Later, the MBMG instituted a statewide study of low temperature occurrences partially funded by ERDA and DOE.

Figure 1 is a histogram of thermal spring temperatures in Montana. The large block of springs representing temperatures of 30°C or less is, in the majority of cases, related to springs issuing from the Madison Group. Most geologic parameters tend toward normal or lognormal distribution. Ground-water temperatures appear to have a lognormal distribution; in Montana, the average ground-water temperature is between 7 and 9°C depending upon the area of the state under discussion. Figure 2 is an approximation of the type of distribution one would expect for thermal spring temperatures; from Figure 2 we infer that the data presented in Figure 1 are grossly biased, i.e., that we have only included those springs with temperatures of less than 25°C which have high discharges. If the temperature of a spring is greater than 25°C, it is usually safe to assume (in western Montana) that even in the summer a body of ponded spring water loses more heat than it gains. At temperatures less than 25°C and low spring discharge quantities (less than 50 gpm), it is possible for solar and biological factors to increase the measured temperature enough to cause a spuriously anomalous spring temperature.



TEMPERATURE OF SPRINGS IN 2°C INCREMENTS; FIRST BLOCK IS 15-16°C.

FIGURE 1. HISTOGRAM DEPICTING THE FREQUENCY OF THERMAL SPRING TEMPERATURES.

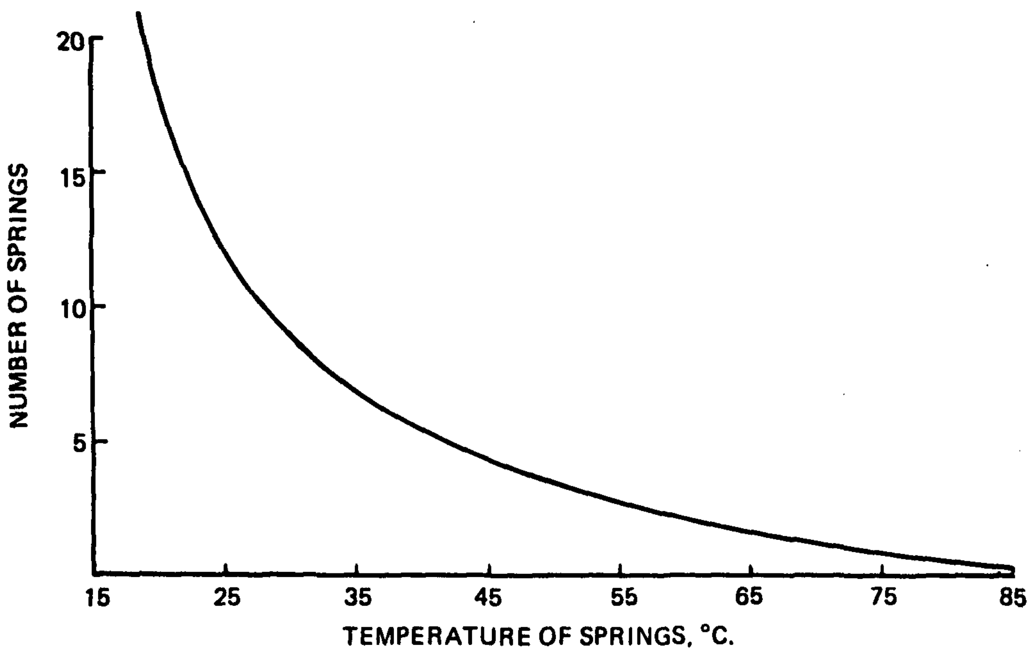


FIGURE 2. EXPECTED LOG-NORMAL DISTRIBUTION OF SPRING TEMPERATURES.

Also, our investigations into mine-water drainage, which is usually of fairly shallow origin, showed that the smaller the discharge value the greater the annual variation in water temperature (14). The smallest discharge reported in the MBMG spring data list for springs in the 15 to 20°C range is 130 gallons per minute, and only two of the springs have discharges of less than 1000 gpm (4). By comparison, only two of the seven springs with temperatures of 65°C or greater heave discharges greater than 100 gpm (Hunters Hot Springs and Boulder Hot Springs).

Obviously, we have erred on the side of being conservative in our past work. However, Table 1 (condensed and updated from references 4 and 21; the former includes location information and some water quality data) shows that when available heat energy is calculated to bottom-use temperatures of 25, 18, and 10°C, only the high discharge/low temperature springs constitute a significant resource. An alternate way of viewing these data is with respect to heat pump usage. For a domestic dwelling of 2500 square feet, the generally available heat pumps now being produced would require 10 to 15 gpm of 15°C water for typical Montana winter weather conditions. Thus, a 15°C spring with a proven 150 gpm yield could only heat ten domestic dwellings. By comparison, even without the use of a heat pump, 150 gpm of 60°C water will heat 60 to 75 domestic dwellings using modern design practices. It is for these practical reasons that only large volume springs were initially emphasized in our studies.

Figure 3 depicts the locations of the springs listed in Table 1. Most of these springs are in western Montana, with the largest concentration in southwestern Montana. At present, there are no known instances of magmatic heating of these thermal waters (6). Dates on the age of igneous rocks in Montana range from very ancient to 0.11 million years before present (9). Known rocks younger than 2.0 million years are very few, extrusive, and of very limited extent in western Montana; consequently, they are not believed to represent a significant thermal resource. The known geothermal systems in eastern Montana are believed to result solely from deep circulation of meteoric ground water with fracture control of spring locations (21).

The best summary to date of all available water chemistry is by Leonard *et al* (15) from 24 springs and 3 wells, which is essentially for the southwestern portion of the state. By the time this article appears, the MBMG will have published a preliminary map of the geothermal resources of Montana, which will include the most representative chemical data for at least 70 springs and wells. Also, an annotated bibliography of geothermal studies in Montana, current through January of 1980, has just been published (20), and NOAA has published a thermal spring list for the United States (5).

Geophysical studies at hot spring sites have been conducted by the U.S. Geological Survey and the three units of the University System listed previously. All of these results have emphasized the importance of faults and fractures controlling the

Table 1. Heat value of water from selected springs and flowing wells.

Name	Temp. (°C)	Flow (gpm)	H_1 (25°C) (10^9 Btu/yr)	H_2 (18°C) (10^9 Btu/yr)	H_3 (10°C) (10^9 Btu/yr)
SPRINGS					
Alhambra	56.5	100	24.9	30.4	36.7
Anaconda	21.7	3.2		0.09	0.30
Andersons	25	75		4.15	8.89
Andersons Pasture	26	900	7.11	56.9	114
Apex	25	750		41.5	88.9
Avon	25.5	24	0.09	1.42	2.94
Bear Creek	24	10		0.47	1.11
Bearmouth	20	1100		17.4	86.9
Beaverhead Rock	27	100	1.58	7.11	13.4
Bedford	23.6	1500		66.4	161
Blue Joint	29	200	6.32	17.4	30.0
Boulder	76	590	238	270	308
Bozeman	54.6	75	17.5	21.7	26.4
Bridger Canyon	20.2	150		2.61	12.1
Broadwater	62	12	3.51	4.17	4.93
Brooks	19.9	72000		1080	5630
Browns	23.7	1100		49.5	119
Camas	45	24	3.79	5.12	6.64
Carter Bridge[1]	26.5	1500	17.8	101	196
Chico	45	320	50.6	68.3	88.5
Deer Lodge Prison	26	100	0.79	6.32	12.6
Durfee Creek	21.1	2300		56.3	202
Elkhorn	48.5	30	5.57	7.23	9.12
Ennis	83.2	15	6.90	7.73	8.67
Gallogly (Lost Trail)	38	100(?)	10.3	15.8	22.1
Garrison	25	54		2.99	6.40
Granite	51	100	20.5	26.1	32.4
Green	26	80+	0.63	5.06	10.1
Gregson (Fairmont)	70	40	14.2	16.4	19.0
Greyson	17.9	900			56.2
Hunsaker[2]	24.5	110		5.65	12.6
Hunters	59	1300	349	421	503
Jackson	58	260	67.8	82.2	98.6
Kimpton[2]	18	300			19.0
La Duke	65	130	41.1	48.3	56.5
Landusky	21	3100		73.5	269
Landusky Plunge	24	2900		137	321
Little Warm	22	5000		160	474
Lodgepole	30	2700	107	256	427
Lolo	44	180	27.0	37.0	48.3
Lovells	19.4	3500		38.7	260
McMenomey Ranch	19	7300		57.7	519
Medicine	46	100	16.6	22.1	28.4
New Biltmore	53	26-	5.75	7.19	8.83

Name	Temp. (C)	Flow (gpm)	H_1 (25°C) (10^9 Btu/yr)	H_2 (18°C) (10^9 Btu/yr)	H_3 (10°C) (10^9 Btu/yr)
Nimrod	20.5	3200		63.2	265
Norris	52.5	106	23.0	28.9	35.6
Paradise	43.4	17	2.47	3.41	4.49
Pipestone	57	250	63.2	77.0	92.8
Plunkets	16.5	4000			205
Potosi,3	38	17	1.75	2.69	3.76
Pullers	44.4	50	7.66	10.4	13.6
Renova	50	40	7.90	10.1	12.6
Silver Star	71.5	40	14.7	16.9	19.4
Sleeping Child	45	530(?)	83.7	113	147
Sloan Cow Camp	29.5	350	12.4	31.8	53.9
Staudenmeyer	28	1800	42.7	142	256
Sun River	30.4	710	30.3	69.5	114
Targhee Sulfur[2]	18	55			3.46
Toston	15.2	20000			822
Trudau	22.7	175		6.50	17.6
Vigilante	23.5	2200		95.6	235
W.S. State Hosp.	77	60	24.6	28.0	31.6
Warner	18	130			8.22
West Fork S.H.	26	500	3.95	31.6	63.2
White Sulphur[3]	46	400+	66.4	88.5	114
Wolf Creek	68	53	18.0	20.9	24.3
WELLS					
Camp Aqua	50	330+	65.2	83.4	104
Colstrip[4]	96	230	129	142	156
Lucas	42.2	100	13.6	19.1	25.4
Ringling	48	800	145	190	240
Symes	40	100	11.8	17.4	23.7
White Sulphur-dug	58	350	91.2	111	133

[1]Average temperature with mixing factors deleted.

[2]Added after Figure 1 was drafted.

[3]Replaced by well.

[4]Cemented and abandoned.

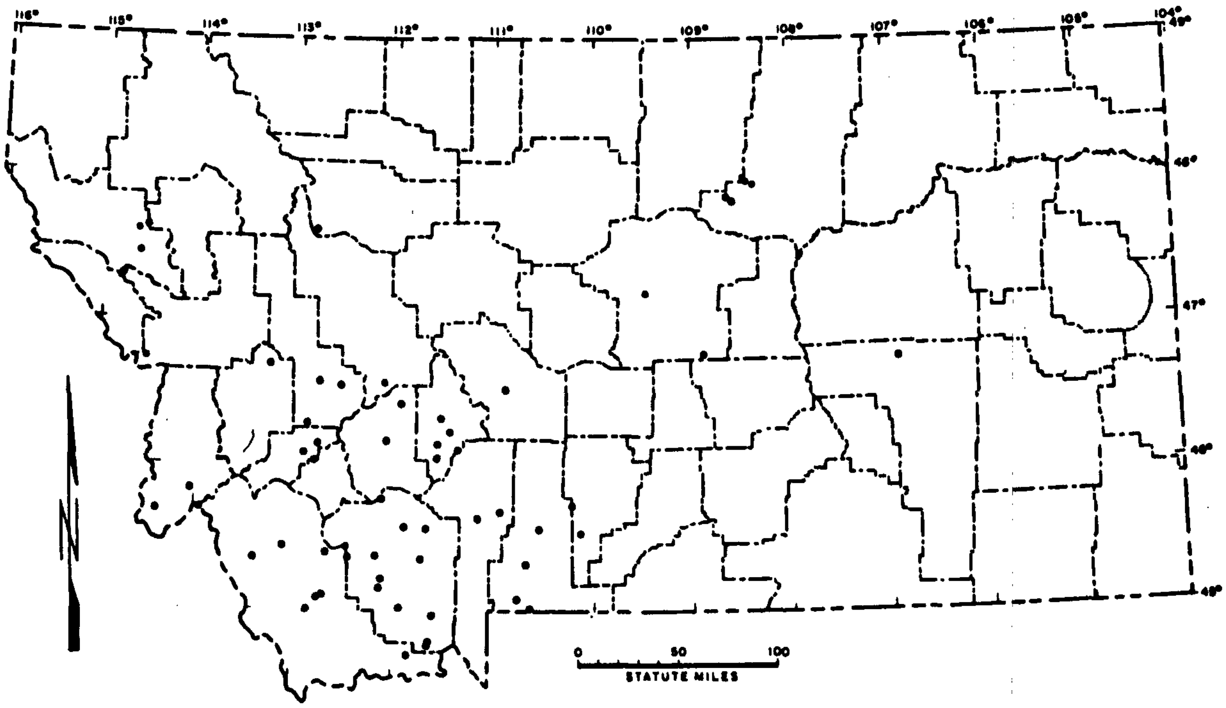


FIGURE 3. LOCATION OF THERMAL SPRINGS IN MONTANA.

occurrence of the hot spring systems that have been studied. The Ennis hot spring has the highest surface temperature (83°C) of all springs in the state, and has been the object of detailed study by the USGS and the Montana Tech Geophysics Department. At the Ennis (Thexton) hot spring, gravity, seismic, telluric, and audio-magnetotelluric investigations have shown that block faults parallel and nearly normal to the valley trend have controlled the discharge point of the thermal system (8, 17, 18). Studies at other sites such as: (1) Warm Springs State Hospital (12); (2) Silver Star (1, 16); (3) Norris and Hunters hot springs (7); and the Little Bitterroot Valley (Camas area, work in progress, 10, 13) show structural factors as having a significant effect on the location of the thermal system discharge point(s).

WARM AND HOT WELLS

Thermal wells can be divided into two basic categories: (1) those wells drilled with the express intention of obtaining hot water or hot dry rock; and (2) wells drilled for hydrocarbons or water which incidentally encountered hot water. The boundary between these two classes is sometimes vague, representing water wells drilled near a hot spring with the hope that hot water might be encountered.

Wells have been drilled expressly for geothermal purposes at the Bozeman, Broadwater, Ennis, Fairmont, Warm Spring State

Hospital, and White Sulphur Springs hot spring areas and at the Marysville heat flow anomaly. Results to date have not been highly encouraging. The best results have occurred at the Broadwater hot spring where Frank Gruber is reported to have obtained about 350 gpm of water at approximately the spring temperature, 62°C or 144°F (R. B. Leonard, pers. comm.). The results and duration of pump testing at Broadwater have not been made public, so we have no way of evaluating whether this system will provide a sustained yield at the tested discharge rate and temperature.

At White Sulphur Springs, Dave Grove has promoted the development and utilization of geothermal energy. The first attempt was to drill a deep well to heat the new bank building. The well was drilled in 1978 to a depth of 875 feet. Temperature logging of this well showed that the hottest zone encountered was between depths of 100 to 200 feet; the pump test data provided a calculated transmissibility of 103,000 gallons per day per foot of drawdown (gpd/ft) and an estimated safe yield of 50 gpm of 118°F (48°C) on a continuous use basis (D. E. Dunn, pers. comm.). The second project was to improve the spring area by cleaning it out and installing a cement culvert (equivalent to the procedure used for dug and bored wells). This system is reported to be producing 350 gpm of 136°F (58°C) water (Lloyd Donovan, pers. comm.). The latter approach is an excellent example of successful inexpensive development; previously reported temperatures for the spring range from 95 to 125°F, with the "best" value being 115°F. It appears that in the process of improving the spring, shallow ground water mixing was reduced, producing the higher temperature.

Other spring operators have not been as fortunate. At Fairmont (Gregson) hot springs, several wells were drilled in an attempt to increase the amount of hot water available. All of these wells produced cold water. Experience at the Bozeman hot spring has been mixed. The present "spring" is actually a shallow well adjacent to the spring discharge point. A recent attempt to obtain more hot water resulted in a well which could not be held open and which did not produce enough water to warrant installing a pump; reworking of this well has improved its yield.

The Marysville "hot dry rock" well was drilled because of very high heat flow values in that area. Unfortunately, the 6790 foot deep well encountered water bearing zones with a maximum temperature of 204°F (96°C) (19).

By comparison, the 540 foot well drilled last summer at Ennis, while originally scheduled as a test well, had smaller diameter pipe used for heat flow testing. The well hit bedrock at approximately 540 feet and had a bottom hole temperature of 95°C (203°F). With the bottom open it was flowing 2.5 gpm with a surface temperature of 93°C (199°F) (R. B. Leonard, pers. comm.). At present there is an obstruction in the well and attempts to fish it out have so far been unsuccessful.

At Warm Springs State Hospital, a 1498 foot production/test well was drilled in the fall of 1979. The driller's pump broke down during development, so no pump testing was conducted. At the time the pump failed, it was reported that the discharge was about 140 gpm, with 975+ feet of drawdown, which yields a maximum transmissibility coefficient (T) of 200 gpd/ft. A flange, pressure gauge, and additional valve were recently installed by the shopmen at the hospital. We conducted a short, 65 minute, shut-in test on 9 April 1980 which proved interconnection between the well and spring, and provided T values of 34 gpd/ft before the spring responded and 70 gpd/ft after spring flow started increasing. The shut-in pressure at the end of the test was 138 pounds per square inch (psi). Based upon the data available, we estimate that the well has a maximum safe yield of 70 gpm of 78 to 80°C water. The difference in T values between the development work following drilling and the shut-in test may be because slotted casing was used instead of well screen and there may be some very large well losses. The Montana Energy Research and Development Institute has scheduled additional development and testing for this well and it is hoped that the well performance can be improved.

In the category of wells which incidentally encountered hot water, the best documented case is the Western Energy well at Colstrip. The well was drilled to a depth of 9200 feet; the majority of the hot water is believed to have come from the Mission Canyon Limestone at a depth of 7700 feet. Well tests by Van Voast yielded a transmissibility of 650 gpd/ft, and a storage coefficient of 2×10^{-4}; under test conditions, the well flowed 230 gpm of 207°F (97°C) water with a 16 psi confining pressure. A petroleum laboratory analysis of the water yielded a total dissolved solids content of about 1500 milligrams per liter. The pH value reported was 6.3, which is not very acidic; but, the water was sufficiently corrosive to cause casing leaks in a period of about five years. The well has since been cemented and abandoned.

Old petroleum test wells that produce warm or hot water frequently produce this water from the Madison Group. The Ringling and Lucas wells near White Sulphur Springs produce 800 and 100 gpm of 48°C (118°F) and 42°C (108°F) water from Mississippian age rocks (15). The Saco well, now used by the Sleeping Buffalo Resort produces a reported 290 gpm of 49°C (106°F) water from this same strata.

A recent study by P.R.C. Toups, Inc. for the Fort Peck Indian Reservation has proven a valuable resource is available in the water separated from the crude oil produced on the Poplar Dome. Also, they suspect that hot water may be available at relatively shallow depths north and east of Poplar along the trace of the Brockton-Froid fault zone (22).

HEAT PUMP APPLICATION

The present heat pump technology calls for "heavy duty" pumps and compressors in order to utilize typical Montana ground water in the temperature range of 42 to 47°F (6 to 8°C). Figure 4 shows six areas which appear to have ground-water temperatures above 10°C, and many be suitable for use with normal heat pump systems. A word of caution is needed with respect to these data. Temperature is one of the most easily altered characteristics of ground water due to failure to pump a well long enough for all aspects of the delivery system to come to thermal equilibrium, either due to the problem of disposing of the water or low well yield. Most inventory work is done during the summer months, which commonly means that any error in the temperature measurements validity will be biased towards a higher temperature.

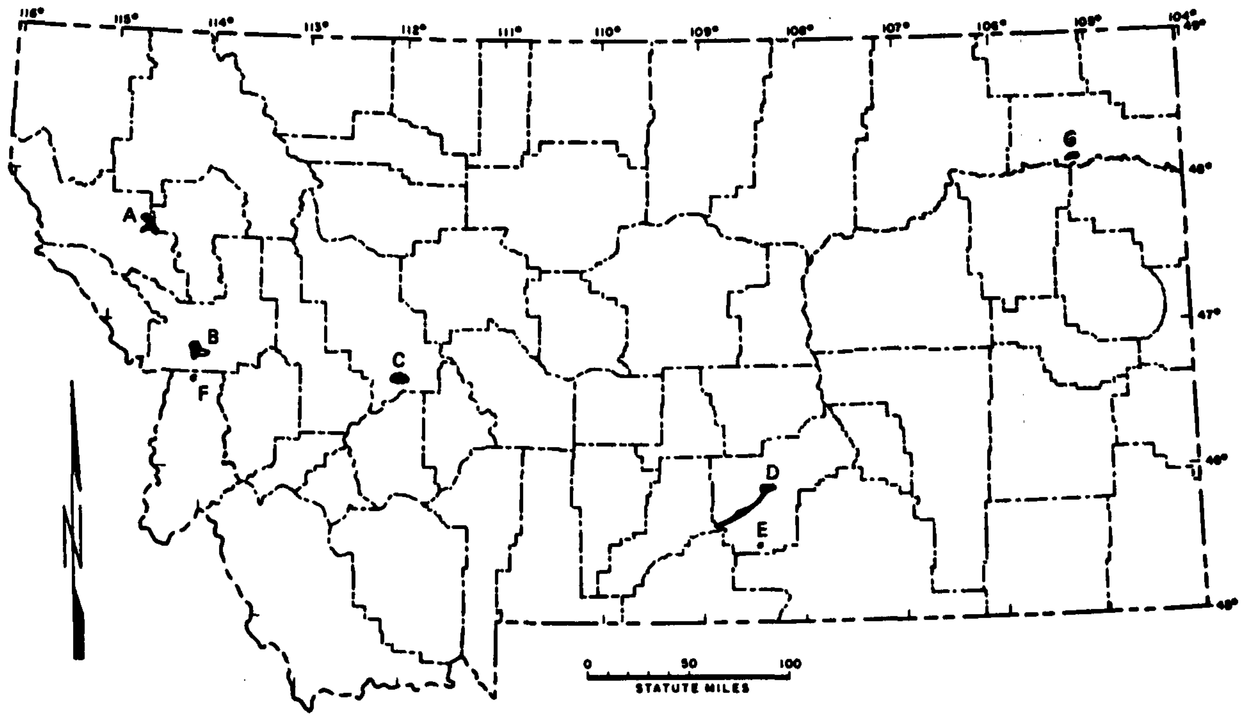


FIGURE 4. LOCATION OF AREAS FAVORABLE FOR HEAT PUMP USAGE. SEE TEXT CONCERNING SPECIFIC AREAS.

Favorable areas B, C and D are in suburban areas of Missoula, Helena, and Billings, where problems of water disposal are greater. The reported "warm" temperatures for these areas contribute a smaller percentage of the total number of temperatures in these areas, and may be related to failure to achieve thermal equilibrium. The water is almost entirely from shallow (< 300 feet deep) wells and may show considerable seasonal variation. In these three areas, it is recommended that the water temperature be measured during the winter season after the well has been pumped steadily for at least two hours. If the temperature and yield are satisfactory under these conditions, the well should be permitted to recover and a three-day continuous pumping recording the water level in the well should be conducted to ensure an adequate yield. Most people in the field believe that a sustained yield of 20 gpm is required (11).

Other areas depicted on Figure 4 have greater certainty of the temperature data. The Little Bitterroot Valley (area A) has an extensive gravel aquifer in the valley fill sediments. Temperatures of well water produced from this zone generally range from 10 to 51°C. The area is still under investigation by Joe Donovan and a final report will be issued by MBMG in 1981.

Area E, northeast of Pryor, is tentative at this time. A drilling report for one water well indicates that wells drilled into the Kootenai Formation should be abnormally warm in this area.

Area F is provisional at present, being based upon the temperature from one well. The Bureau recently drilled a 400 foot municipal test well outside of Florence. Flow testing of this well was brief (120 minutes at 10 gallons per minute); however, the well produced water at a temperature of 64°F (17.3°C). Even if increased preduction from this zone lowered the temperature because of pumping-induced vertical movement of cooler water from above, the production temperatures should still be adequate for heat pump use.

Area G, just off the Poplar Dome, is the site of ground temperature surveys conducted by Joe Birman of Geothermal Surveys Inc. Temperatures were measured at a depth of ten feet below land surface and temperatures greater than 10°C (50 F) were encountered along several linear trends (22). Bedrock is the Bearpaw Shale in this area and it may be necessary to drill fairly deep to obtain sufficient water for heat pump use. The investigators hope to find a secondary zone of hot water at a depth of roughly 500 feet, just below the Bearpaw Shale.

SUMMARY

Good data are available for most of the thermal springs in Montana. The quality of data for thermal wells varies greatly and part of our current effort is to improve this data base. Data presented show heat content for various reference temperatures related to low temperature use. Drilling results are variable in the vicinity of hot springs; development of the springs is recommended prior to drilling. Heat pump utilization will increase, with the greatest potential being in the Little Bitterroot Valley.

ACKNOWLEDGMENTS

This paper utilizes data collected by many workers. The authors thank their colleagues at the Montana Bureau of Mines and Geology, R. N. Bergantino, J. J. Donovan, S. M. Kovacich, D. C. Lawson, J. J. McDermott, M. R. Miller, T. W. Patton, K. S. Thompson, and W. A. Van Voast, their associates at Montana Tech, J. W. Halvorson and C. W. Wideman, and Glenn Wyatt from the Earth Sciences Department at Montana State University. Data from the U.S. Geological Survey, and the cooperation of R. B. Leonard, is greatly appreciated.

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Addendum

The first issue of Geothermics for 1981 (v. 10, no. 1) arrived after submission of this manuscript. This issue includes an article entitled "Sodium/Lithium Ratio in Water Applied to Geothermometry of Geothermal Reservoirs" by Christian Fouillac and Gil Michard (p. 55-70). They present the following two empirical equations for reservoir temperature calculation:

(1) $\log_{10}(m_{Na}/m_{Li}) = 1000/T - 0.38$, for $Cl^{-} < 0.2M$, and

(2) $\log_{10}(m_{Li}) = -2258/T + 1.44$, for $Cl^{-} < 0.2M$;

the reader is referred to Fouillac and Michard for details on the deviation of the equations.

Using these equations with the Camp Aqua well data results in the highest calculated reservoir temperatures. Equation (1) yields a reservoir temperature of 49°C, slightly below the observed temperature at the wellhead. Equation (2) yields a reservoir temperature of 83°C, slightly greater than our source temperature using the chalcedony curve on the SiO_2 - Enthalpy plot (figure 6). While these calculations are subject to the concerns about dilution and ion-exchange processes, these data provide additional support for use of the chalcedony curve on SiO_2 - Enthalpy plots for low-temperature geothermal systems.

Geophysical Investigations of Certain Montana Geothermal Areas

Charles J. Wideman, Lester Dye,
James Halvorson, and Mark McRae

Selected hot springs areas of Montana have been investigated by a variety of geophysical techniques. Resistivity, gravity, seismic, and magnetic methods have been applied during investigations near the hot springs. Because the geology is extremely varied at the locations of the investigations, several geophysical techniques have usually been applied at each site.

Figure 1, an illustration of the generalized geology of Montana, is used to illustrate the extreme complexity of the geology of the study areas. As can be seen from the figure, southwestern Montana is a region of Basin and Range type features superimposed on a mosaic of Batholiths, volcanics, and other features. Northwest Montana is an area of thrust faulting and folding, followed by normal faulting; the normal faults frequently control the orientation and width of the valleys. Along the southern boundary of the state, in the Yellowstone region, the geologic framework is complex, with areas of faulting, folding, and relatively young volcanics.

From the geologic background shown in Figure 1, several hot springs areas have been chosen for investigation. The locations of some of the hot springs are shown in Figure 2. Special emphasis has been placed on the West Yellowstone area for this report because of the unique method of interpretation of gravity data for this area. The interpretation will be discussed in detail later. Table I is a listing of the hot springs which have been investigated by Montana Tech researchers using geophysical techniques. Comparison of Figures 1 and 2 shows that most of the areas studied have been along valley margins in the southwestern portion of the state.

The West Yellowstone research area is unique in terms of geologic setting. It is also unique in that it is immediately adjacent to a National Park. The study area, a region approximately six miles by nine miles in extent, is northwest of the town of West Yellowstone, Montana. In this area, there is geologic evidence of folding, thrust faulting, rhyolite flows, and modern day block faulting caused by uplift and extension. The results of a gravity survey of the area have been combined with gravity data obtained from the Geophysical Data Base maintained by N.O.A.A. in Boulder, Colorado. The combined gravity data were used to obtain a Bouguer Anomaly map which is shown in Figure 3. The geologic complexity of the area is reflected in the Bouguer map, and variations of several milligals over short distances are common.

Interpretation of the gravity map has been facilitated by making use of theoretical work done by Yeatts (1973). His work predicts the displacements expected near faults of various attitudes and displacements and, in particular, his map showing expected vertical displacements near a thrust fault with 15 degree dip is illustrated in Figure 4. The pattern of displacements shows remarkable similarity to the portion of the gravity map near Horse Butte, located in the northwest portion of the survey. Therefore, the Horse Butte area is interpreted to be an expression of thrust faulting. The area is suspected of having topographic features at the time of thrust faulting closely resembling those predicted by the map of Yeatts. Inspection of Figure 4 shows that the paleo-surface associated with thrust faulting should have regions of relative low elevations behind the fault. In addition, there should be a region at the sides of the fault which are predicted to be broad lows. Immediately in front of the fault, downwarping occur as on the over-ridden side; but further to the front a moderate high is predicted. The region of relatively high elevations occurs because of folding in front of the thrust fault. Areas which were topographically low have since been filled by rhyolite flows and other material derived from the surrounding high areas. Areas of modern day faulting tend to occur where the fill material undergoes thickness vaiations. To the best of our knowledge, this is the first time that gravitation analysis has been aided by the use of predictions based upon dislocation theory. For the case presented here, the analysis seems to have facilitated the gravity interpretation of an extremely complex area.

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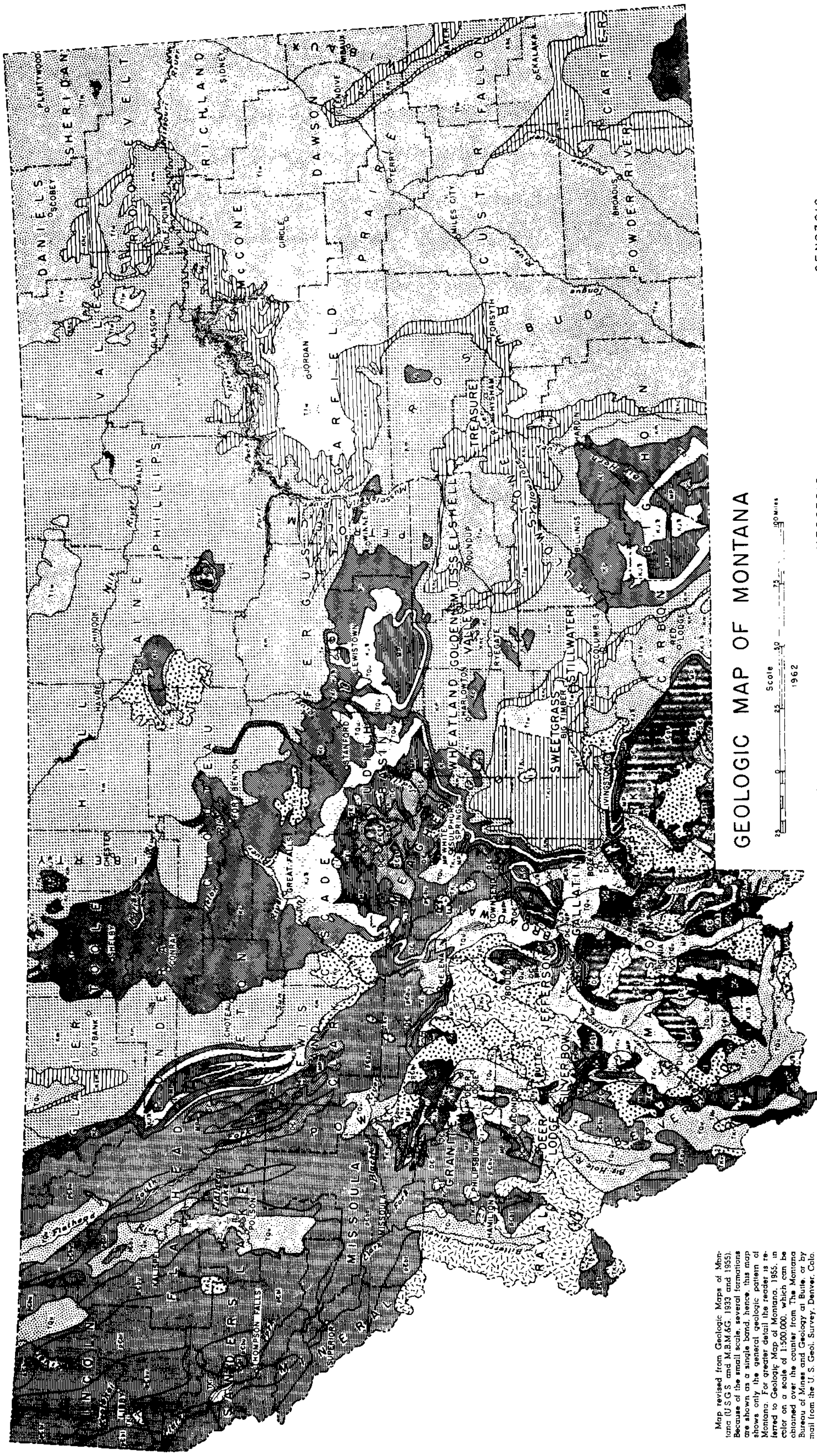
Table 1

Geophysical Investigation Areas

Area Name	Investigation Method
Centennial Valley	Gravity, Magnetic
Deer Lodge Valley	Resistivity, Gravity
Helena Valley	Gravity, Seismic, Resistivity, Magnetic
Hot Springs-Camp Aqua	Gravity, Seismic
Gregson (Fairmont) Hot Springs	Gravity
Pipestone Hot Springs	Resistivity, Gravity, Seismic
Silver Star (Barkells Hot Springs)	Resistivity, Gravity
Ennis (Thexton Hot Springs)	Seismic, Gravity
Warm Springs	Gravity, Resistivity
West Yellowstone	Gravity



EXPLANATION

SEDIMENTARY ROCKS

CENOZOIC

TQu, Tfw

TQu, Tertiary sediments, Willow Cr. fm, glacial deposits, and alluvium in western Montana.
Tfw, Fort Union, Flaxville, Wasatch, and White River fms in central and eastern Montana.

TKl

TKl, Livingston fm of Crazy Mts area; includes Montana group above Eagle ss, Hell Cr., and Fort Union fms.

MESOZOIC

Ksm, Khc

Ksm, St. Mary River formation of north-central Montana.
Khc, Hell Cr fm, central and eastern Montana.

Km

Km, Montana group: Telegraph Creek, Eagle, Claggett, Judith River, and Bearpaw fms over most of state; Fox Hills ss and Pierre sh. in extreme east.

Kc

Kc, Colorado group: Thermopolis, Mowry, Belle Fourche, Greenhorn, Carlile, and Niobrara formations in south-central and eastern Montana, and the Blackleaf fm and Marias River shale of north-central Montana

KJR̄

KJR̄, Lower Mesozoic Ellis group, Morrison formation, and Kootenai fm. Includes Triassic Dinwoody where present.

PALEOZOIC

MP

MP, Mississippian, Pennsylvanian, and Permian fms: Madison group, Big Snowy group, Amsden, Quadrant-Tensleep, and Phosphoria formations.

D€

D€, Devonian and Cambrian formations: Flathead, Wolsey, Meagher, Park, Pilgrim, Red Lion, Maywood, Grove Cr., Jefferson, and Three Forks formations.

PRECAMBRIAN

p€b, p€bu, p€bl

p€b, Belt formations undifferentiated.
p€bu, Upper Belt formations of the Piegan and Missoula groups
p€bl, Lower Belt formations of the Ravalli group and the Prichard formation.

p€gs

p€gs, Pre-Belt gneisses, schists, marbles, and associated rocks of the Cherry Creek series, pre-Cherry Cr. fms, and the Stillwater igneous complex.

IGNEOUS ROCKS

TKv

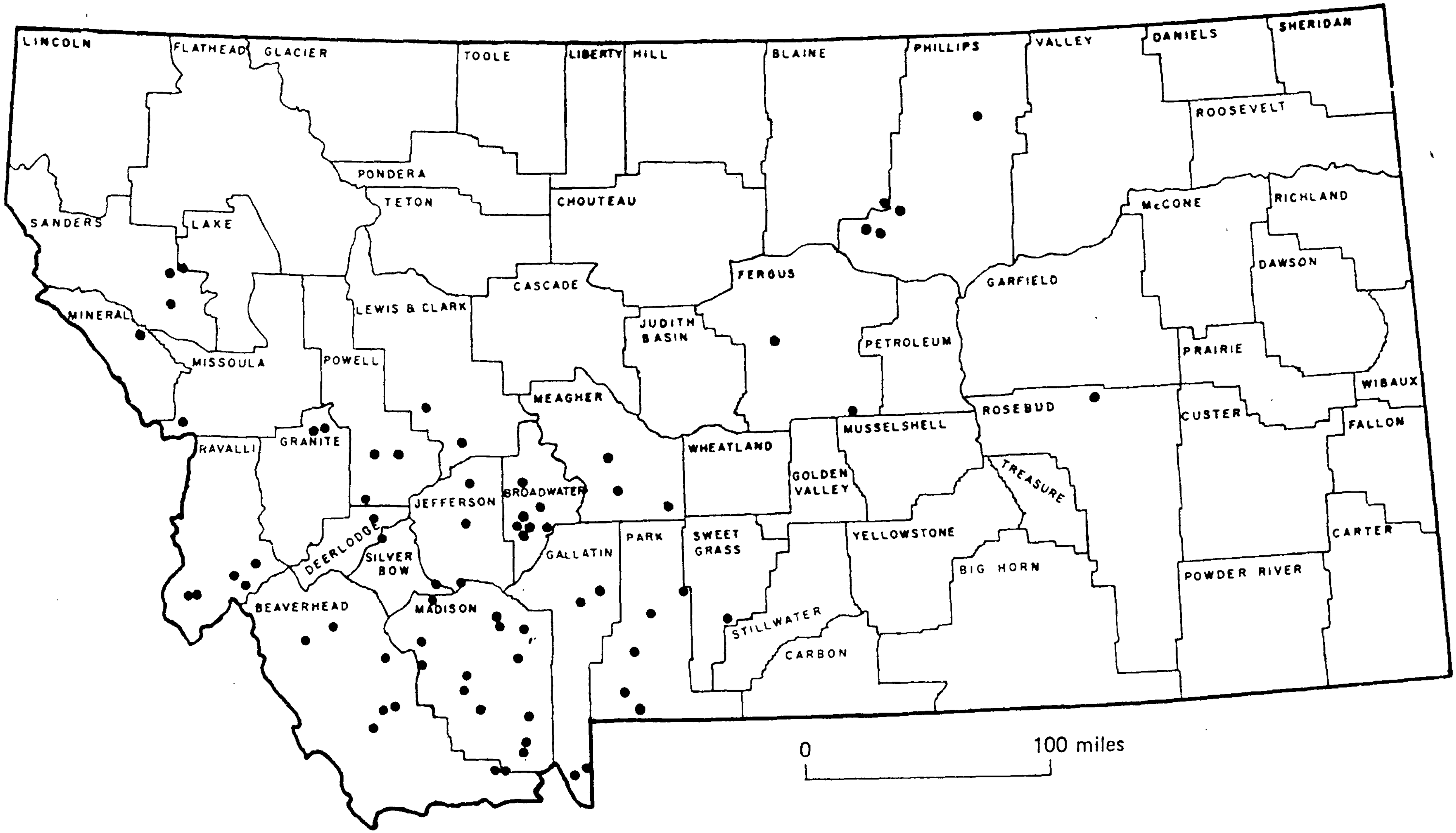
TKv, Volcanic Rocks

TKi

TKi, Intrusive Rocks

Map revised from Geologic Maps of Montana (U.S.G.S. and M.B.M.&G. 1933 and 1955). Because of the small scale, several formations are shown as a single band, hence, this map shows only the general geologic pattern of Montana. For greater detail the reader is referred to Geologic Map of Montana, 1955, in color on a scale of 1:500,000, which can be obtained over the counter from The Montana Bureau of Mines and Geology at Butte or by mail from the U.S. Geol. Survey, Denver, Colo. (Price $3.50)

Figure 1



Locations of hot springs and hot wells.

Figure 2

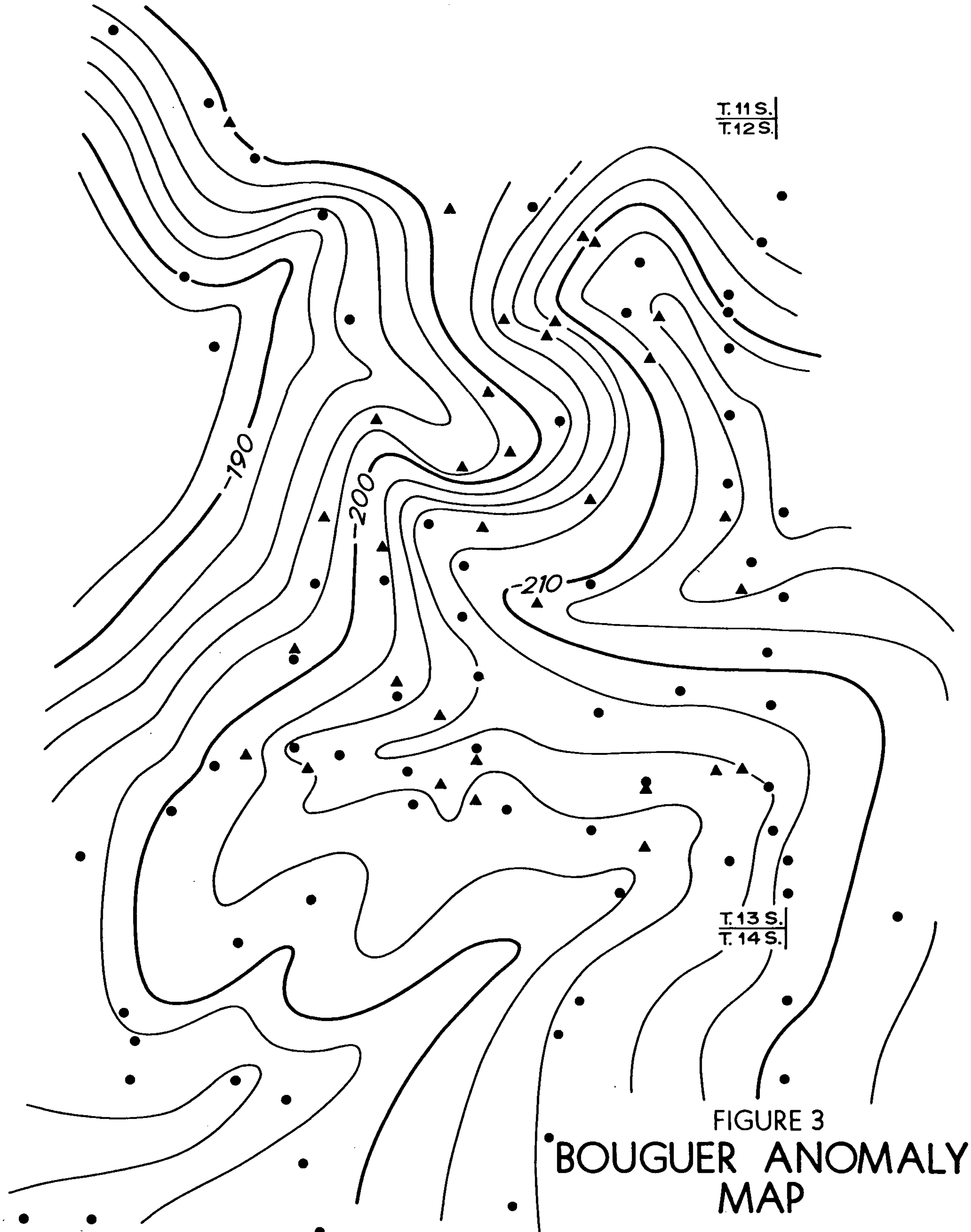


FIGURE 3
BOUGUER ANOMALY MAP
WEST YELLOWSTONE
MONTANA

Vertical Displacements near a Volterra dislocation. Adapted from Yeatts.(1973). Dip angle 15°

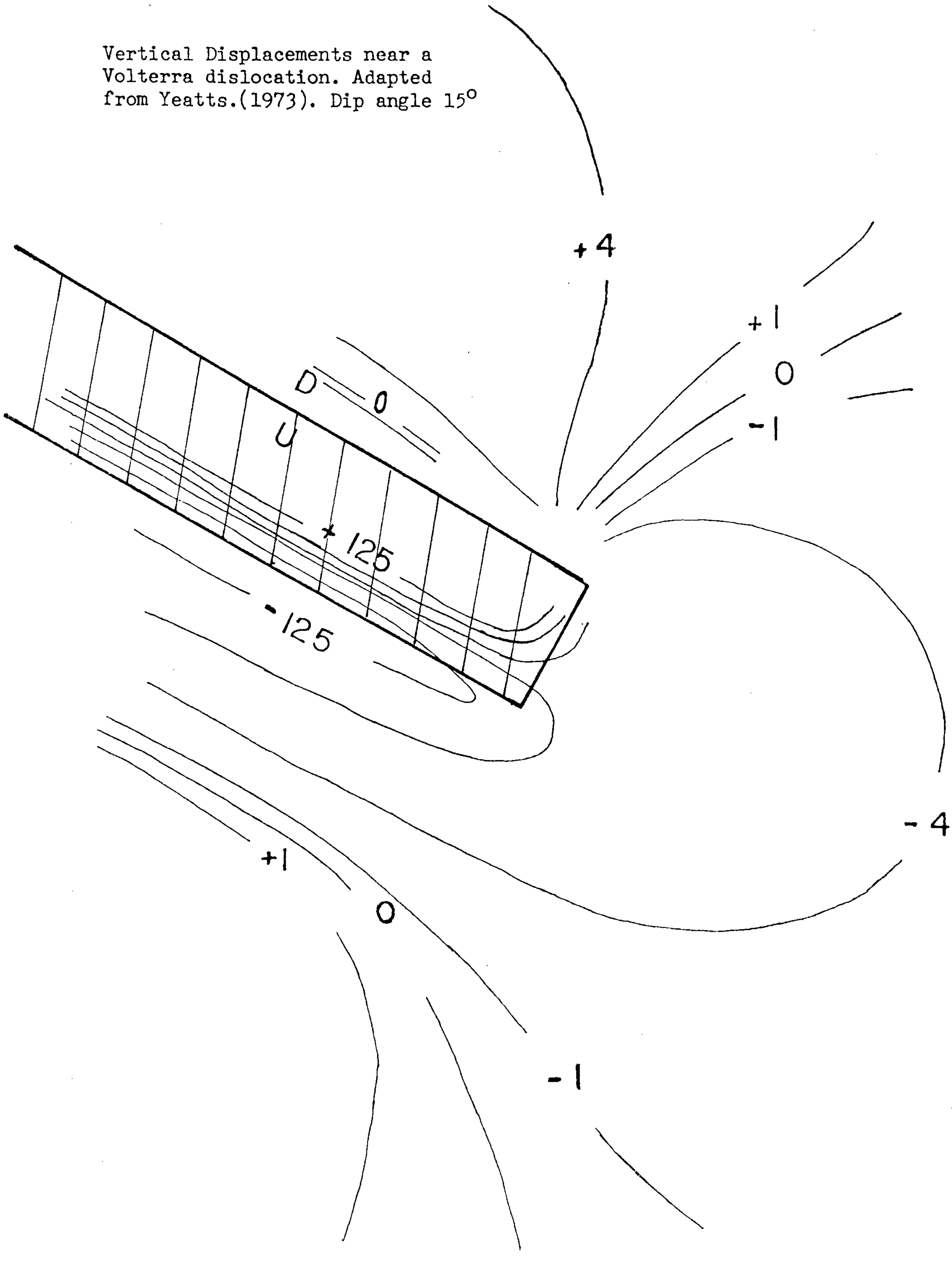


FIGURE 4

187-204

GEOTHERMAL INVESTIGATIONS IN NEBRASKA: METHODS AND RESULTS

William D. Gosnold, Jr.
University of Nebraska at Omaha

Duane A. Eversoll
Marvin P. Carlson

Conservation and Survey Division
University of Nebraska-Lincoln

GEOTHERMAL INVESTIGATIONS IN NEBRASKA: METHODS AND RESULTS

Introduction.

At the inception of the geothermal resource assessment program in Nebraska there was some skepticism about the existence of any geothermal resources within the state. Now after two years of study and collaboration with other workers in the geothermal field we find that about two-thirds of the state has access to a potential low-temperature resource. The nature of the resource is warm water in laterally extensive aquifers which are overlain by thick (> 1 km) sections of low thermal conductivity sediments. For most of the resource area the high temperatures in the aquifers result from high temperature gradients in the overlying shales. However, in the northcentral and far western parts of the state there is evidence for convective heat flow due to updip water flow in the aquifers. The success of the program has resulted from the synthesis of heat flow and temperature gradient measurements with stratigraphic and lithologic data. This paper describes the methods used and the results obtained during the study.

Methodology.

The general plan of the resource assessment has been to acquire heat flow and subsurface temperature data and to synthesize these data with other geological information. Heat flow sites (Figure 1) were selected on the basis of our interpretation of existing data on the thermal regime of Nebraska. The published literature are summarized by Gosnold(1980 a) and are reproduced here for reference (Figure 2). In addition to published literature data from temperature logs made in a number of water table observation wells (Figure 1) were also used in the selection of heat flow sites. Thirteen of the heat flow sites are located in areas indicated as having anomalous subsurface temperatures by the AAPG-USGS geothermal gradient map of North America (AAPG-USGS, 1976). Seven sites are on or near the Nemaha Ridge and ten sites are on or near the Chadron-Cambridge Arch. Three sites also include an area suspected of having anomalous subsurface temperatures by inference from a geothermal gradient map of South Dakota(Schoon and McGregor, 1974).

The geological setting of Nebraska is that of a stable continental platform. The relatively flat-lying sedimentary veneer ranges in thickness from about 300 m in the northeast to greater than 3000 m in the west, and the stratigraphy is relatively well known (Condra and Reed, 1959). The known structural features cannot cause widespread convective heat transfer, thus conductive heat flow is considered to be the primary factor in the thermal structure of the upper crust beneath Nebraska.

In a conductive regime subsurface temperatures are determined by the heat flow and the thermal conductivities of the lithologic units present. Thus knowledge of the heat flow, stratigraphy, and thermal conductivities allows calculation of subsurface temperatures and provides a means for estimating the geothermal resource potential. The general scheme is shown in Figure 3 where heat flow determinations at two sites are used to estimate subsurface temperatures in the regions between and below the sites. The practice of projecting temperature gradients beyond measured depths is theoretically valid in conductive regimes if the thermal conductivities of the stratigraphic section are known. Nevertheless it is best to verify temperature gradient projections with equilibrium temperature measurements in deep wells. An essential component of our investigation is the measurement of deep-well temperature gradients especially those near populated regions where the geothermal resource may be exploited.

Projection of a potential geothermal resource by this method requires using the heat flow, thermal conductivity, and stratigraphic data to produce a subsurface temperature map. Then the temperature contours are superimposed on structure contours of the aquifers, and those regions which satisfy the criteria for a low-temperature resource are defined by the intersecting contour lines.

Results.

A total of 28 wells were completed for heat flow and temperatures were recorded to the nearest 0.01 K at 5 m intervals with a thermistor probe. Bulk conductivities of drill cuttings from nine of the wells were measured at the Southern Methodist University Geothermal Laboratory, and porous rock conductivities were calculated using the method of Sass et al., (1971 a). The remainder of the drill cutting samples are being processed for measurement later. Estimates of thermal conductivities in the remaining wells were made on the basis of lithology and known conductivities to allow preliminary heat flow calculations (Table 1) for the resource assessment. Some of the preliminary heat flow values have been reduced from previous estimates (Gosnold, 1980a, 1980b) to conform with new data on the thermal conductivity of shales in the Midcontinent (Blackwell et al., 1981).

Heat flow values for most of the state range from 38 mWm^{-2} to 67 mWm^{-2} and fall within expected values for a stable platform with only conductive heat flow. However large areas of anomalously high heat flow appear to exist in the north central section and in the panhandle west of the Chadron-Cambridge Arch. These high heat flow areas are interpreted to be due to convective heat flow within deep aquifers. Two separate convective systems are postulated to account for the heat flow anomalies.

One system underlies the north central part of Nebraska and the south central part of South Dakota. The warm water in this system may enter the Dakota Group through a subcrop connection with the Madison aquifer in South Dakota and flow within the Dakota Group through the high heat flow zone. Warm waters are known in numerous wells penetrating the Madison and the Dakota in South Dakota(Schoon and McGregor, 1974) and 12 water wells in Boyd County Nebraska produce warm water from the Dakota Group(Souders, 1976). A flowing well at Lynch Nebraska produces water at 28°C at about 570 l min^{-1} and was formerly used to fill the city swimming pool. Temperature gradients and heat flow increase from east to west in the high heat flow zone suggesting that the source area for the warm water may lie to the west.

A separate convective system is postulated to account for the high heat flow west of the Chadron-Cambridge Arch. Figure 4 is a structure contour map on top of the Dakota from Volk (1972) and shows a configuration that could cause an extensive, convective heat flow anomaly between the arch and the Denver-Julesburg Basin. Structural cross sections (Figures 5 a &b) in western Nebraska from Condra, Reed, and Scherer (1950) indicate that sub-surface temperatures should be high due to great thicknesses of low-conductivity shales. The coupled effect of the thick shale units and updip water flow probably account for the subsurface temperature patterns in the area (Figures 6 a,b, & c). The results of a finite-difference heat flow model are shown in Figure 6 d. An updip flow of water in the Dakota at a rate of 1 m yr^{-1} gives heat flow and subsurface temperature profiles that are consistent with the existing data. The results of the heat flow data do not contradict the predictions of high heat flow by Swanberg and Morgan(1979), in fact the results provide an explanation for the silica geothermometry anomaly.

Both zones which show evidence of convective heat flow will be included in our scheme of projecting subsurface temperatures on the basis of a conductive heat flow model. We can do this because the convecting zones are the aquifers underlying the Cretaceous shales and we see no problem with projecting temperature gradients down to the tops of the aquifers.

The data have been synthesized to produce a temperature contour map for a depth of 1 km (Figure 7). This map is our first attempt to define geothermal resources in Nebraska, and it does delineate regions which overlie potential low temperature thermal waters. A future version of the map will have structure contours for the warm-water-bearing aquifers and temperature contours for the aquifers. We believe that the second version will clearly delineate potential resource areas by showing three pieces of information, i.e., the locality of the resource, the depth to the resource, and the temperature of the resource. We suggest that this approach is a significant improvement over recent attempts to represent geothermal resources in the Midcontinent.

Examples of comparisons between shallow-well temperature gradients and bottom hole temperature data are shown in Figures 8 & 9. The broad scatter in the BHT data is a ubiquitous phenomenon and casts doubt on the usefulness of those data. However, as a large data set, the BHT data are useful for identifyting areas which may have anomalous temperatures.

Concluding Remarks.

In a stable continental interior the synthesis of heat flow data with stratigraphic, and thermal conductivity data is highly effective in exploring for low temperature geothermal resources on a regional scale. An effective method of presentation of the resource in map form is to delineate the resource area with shading, indicate the depth to the resource with contour lines, and indicate the temperature of the resource with another set of contour lines.

Acknowledgements.

This study was performed under U.S. Department of Energy Contract No. DE-AS07-79ET2705.

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Region	Locality Name	Latitude North	Longitude West	Gradient (K km^{-1})	Depth Interval (m)	Conductivity (Wm^{-1}K^{-1})	Rock Type	Heat Flow (mWm^{-2})
Southeast & Nemaha Ridge	Bennet	40° 38.4'	96° 30.8'	28	10-150	2.4	Ls	67
	Elk Creek	40° 15.9'	96° 11.0'	32	200-240	3.3	Do	106
	Liberty	40° 3.0'	96° 27.5'	30	120-145	1.7	Ls+Sh	51
	Stella	40° 11.2'	95° 50.0'	30	60-80	1.8	Ls+Sh	54
	Table Rock	40° 9.1'	96° 4.6'	18	140-155	3.0	Gr	54
	Union North	40° 51.4'	95° 48.9'	20	120-125	2.9	Ls	58
	Union South	40° 46.2'	95° 59.6'	23	75-80	2.4	Ls	55
Northeast	Fremont	41° 29.5'	96° 33.4'	15	60-115	4.2*	Ss	63
	Oakland	41° 49.5'	96° 27.3'	9	75-140	4.2*	Ss	38
	O'Neill	42° 26.2'	98° 39.0'	50	105-150	1.1*	Sh	55
	Wayne	42° 13.8'	97° 2.5'	62	65-115	1.1*	Sh	68
North Central	Naper	42° 58.8'	99° 1.4'	86	10-155	1.1*	Sh	95
	Springview	42° 57.8'	99° 42.4'	109	10-145	1.1*	Sh	120
	Valentine	42° 54.1'	100° 30.3'	64	25-150	2.2*	Sa+Si	145
Chadron Arch: North Platte Area	Box Elder Canyon	40° 57.6'	100° 34.4'	27	45-225	2.2*	Sa+Si	59
	Gothenburg	40° 47.2'	100° 20.5'	30	10-235	2.2*	Sa+Si	66
	Cross Ranch	41° 36.4'	101° 48.2'	54	200-570	1.1*	Sh	59
	Rothwell Ranch	41° 46.7'	101° 40.9'	47	270-570	1.1*	Sh	52
	Milldale Ranch	41° 39.7'	101° 28.7'	49	240-470	1.1*	Sh	54
Chadron Arch: Northern Panhandle	Gordon	42° 54.9'	102° 12.3'	48	10-185	1.7*	Si+Cl	82
	Rushville	42° 36.7'	102° 12.3'	38	120-200	1.7*	Si+Cl	65
	White Clay	42° 47.4'	102° 39.4'	48	10-35	1.7	Si+Cl	82
	Hay Springs	42° 34.2'	102° 38.9'	45	10-235	1.7	Si+Cl	77
	Whitney	42° 45.3'	103° 16.4'	66	10-153	1.7*	Si+Cl	112
Southern Panhandle	Bayard	41° 49.7'	103° 17.0'	60	90-153	2.2*	Sa+Si	132
	Lisco	41° 25.1'	102° 33.5'	47	10-189	2.2*	Sa+Si	103
	Sidney	41° 8.2'	102° 56.1'	52	20-180	2.2*	Sa+Si	114
	Big Springs	41° 4.2'	102° 5.9'	59	10-135	2.2*	Sa+Si	130

Table 1. Preliminary heat flow data in Nebraska. Estimated conductivities are indicated by (*). Rock type key: Ls = limestone, Do = dolomite, Sh = shale, Ss = sandstone, Sa = sand, Si = silt, Cl = clay, Gr = granite.

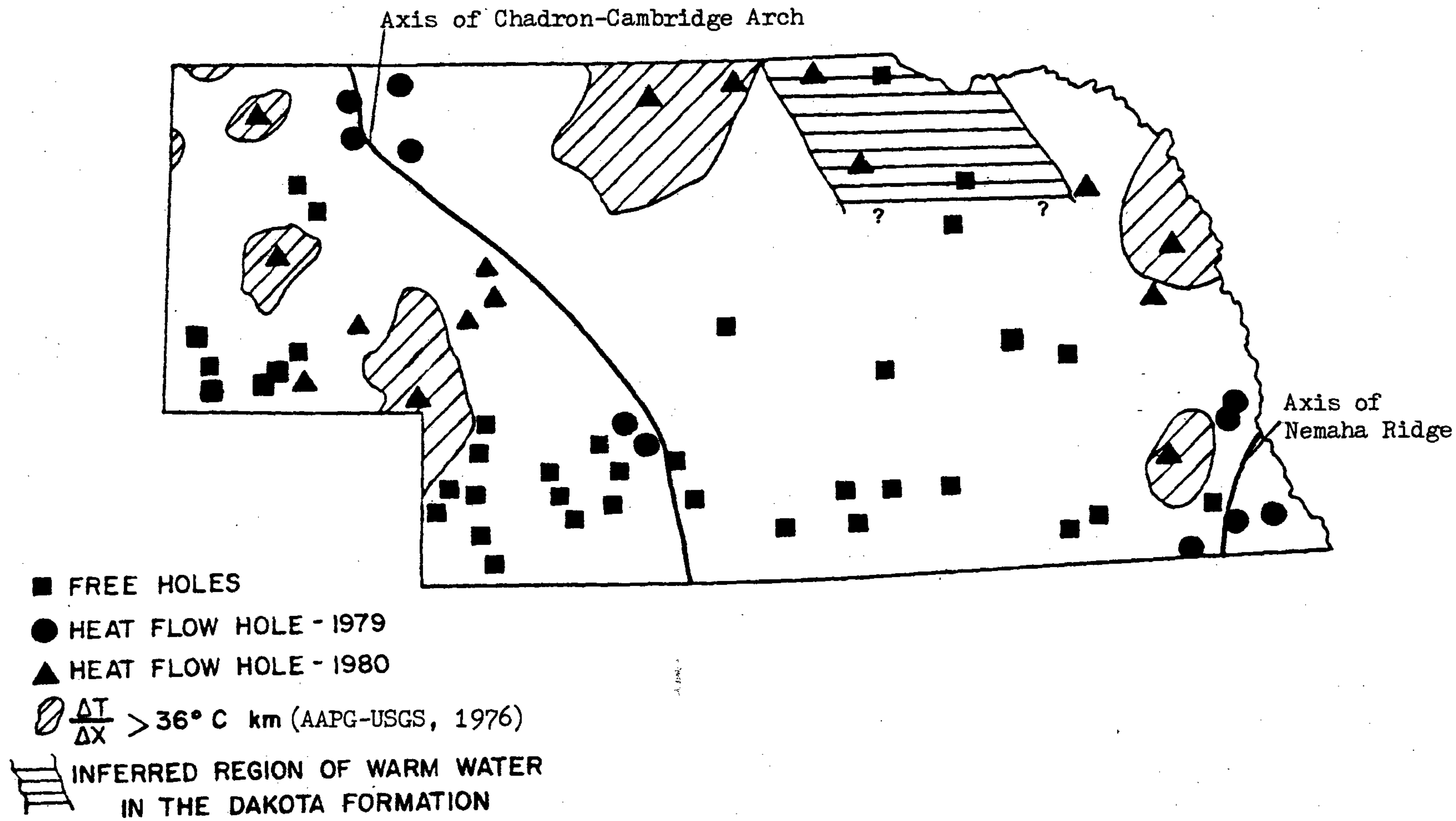


Figure 1. Locations of heat flow sites and other wells where temperature gradients have been measured.

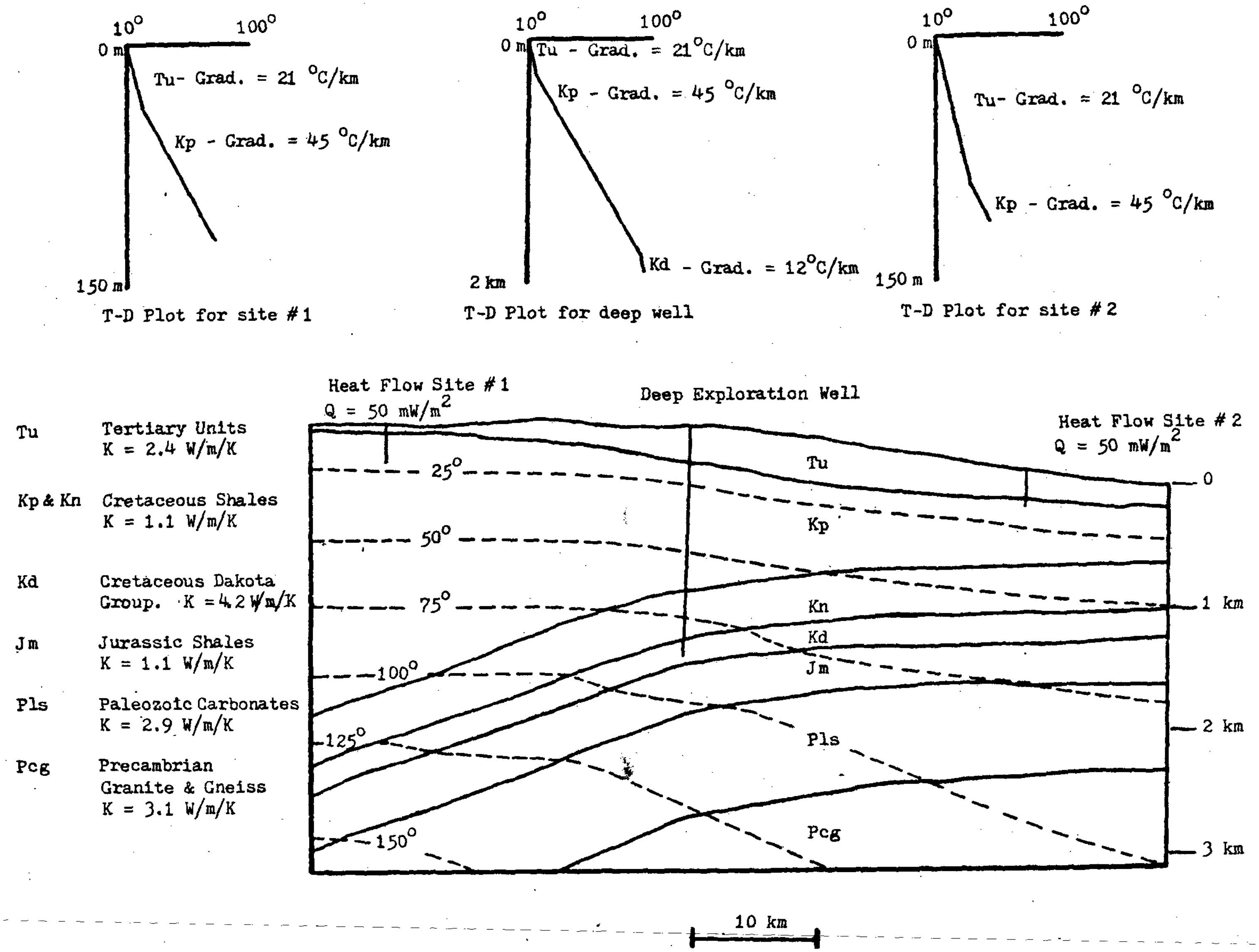


Figure 3. Temperature profile in a conductive thermal regime. Isotherms are contoured on the basis of known heat flow, stratigraphy, and thermal conductivity. The section is typical of western Nebraska where the potential resource is warm water in the sandstone aquifers of the Cretaceous Dakota Group.

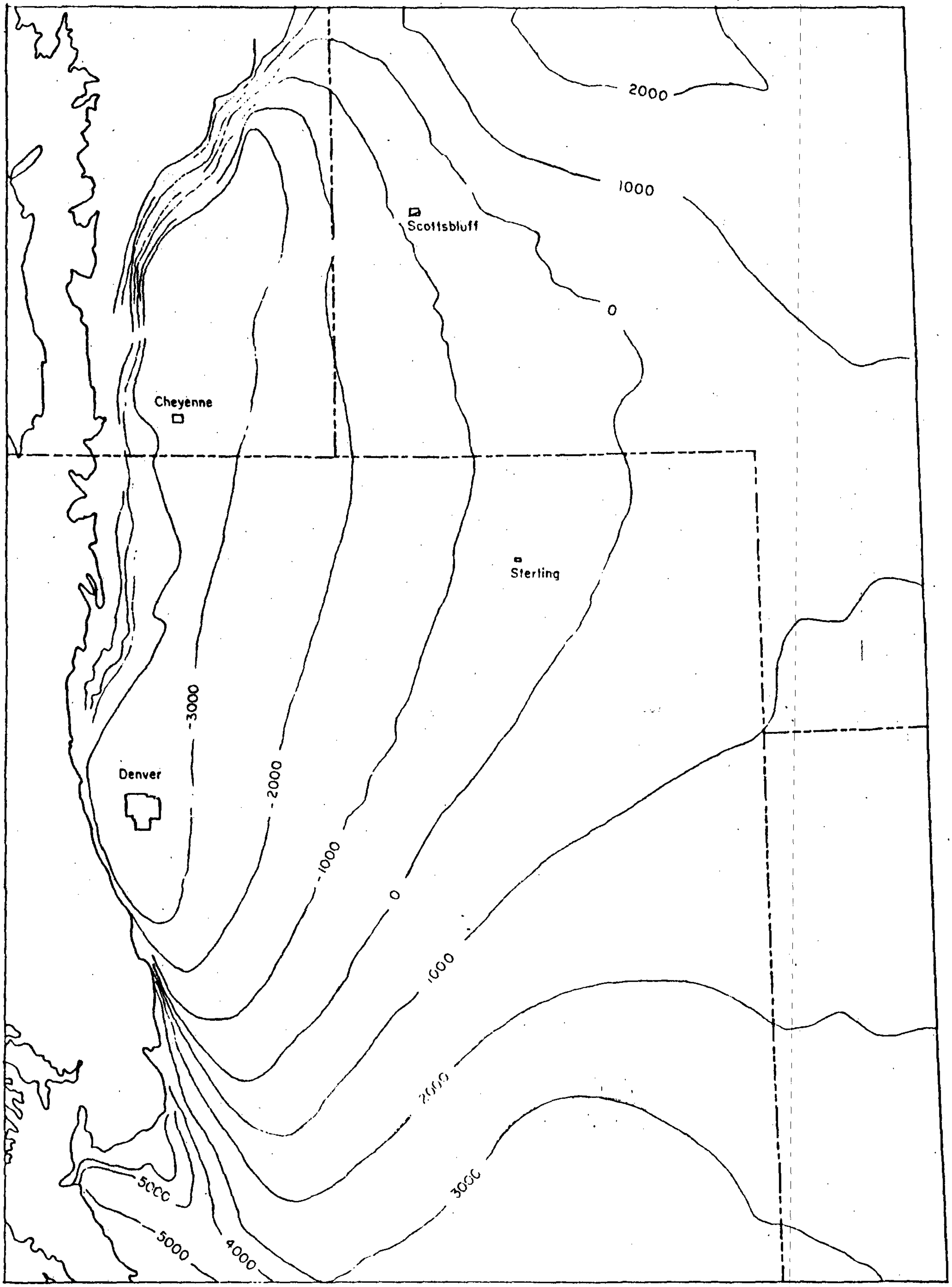


Figure 4. Structure contour map of the Dakota group from Volk, (1972). Datum is sea level.

Figure 5 a.

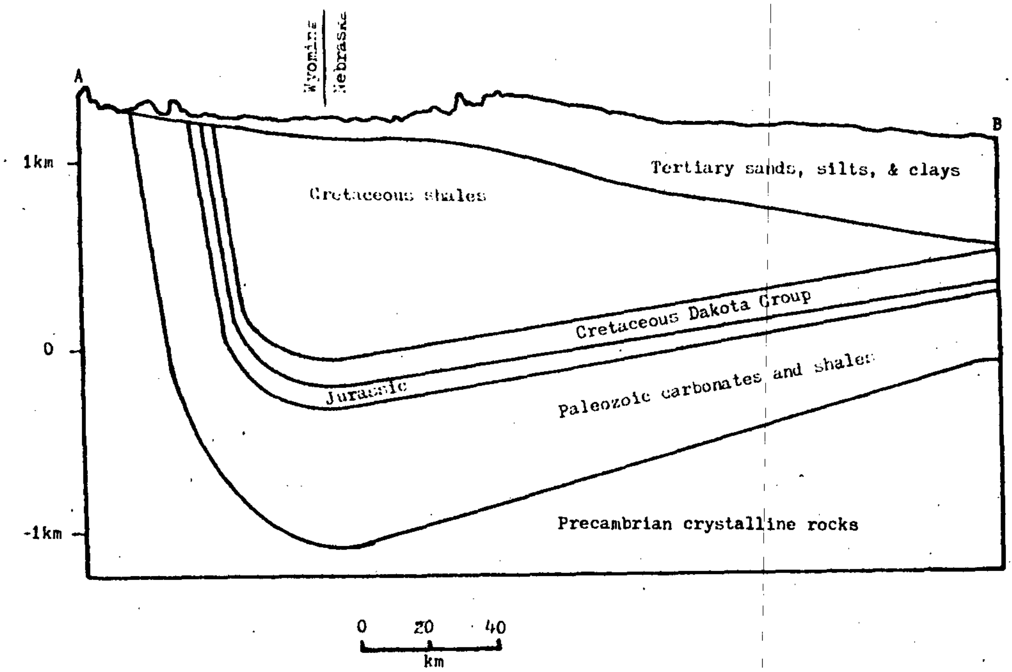


Figure 5 b.

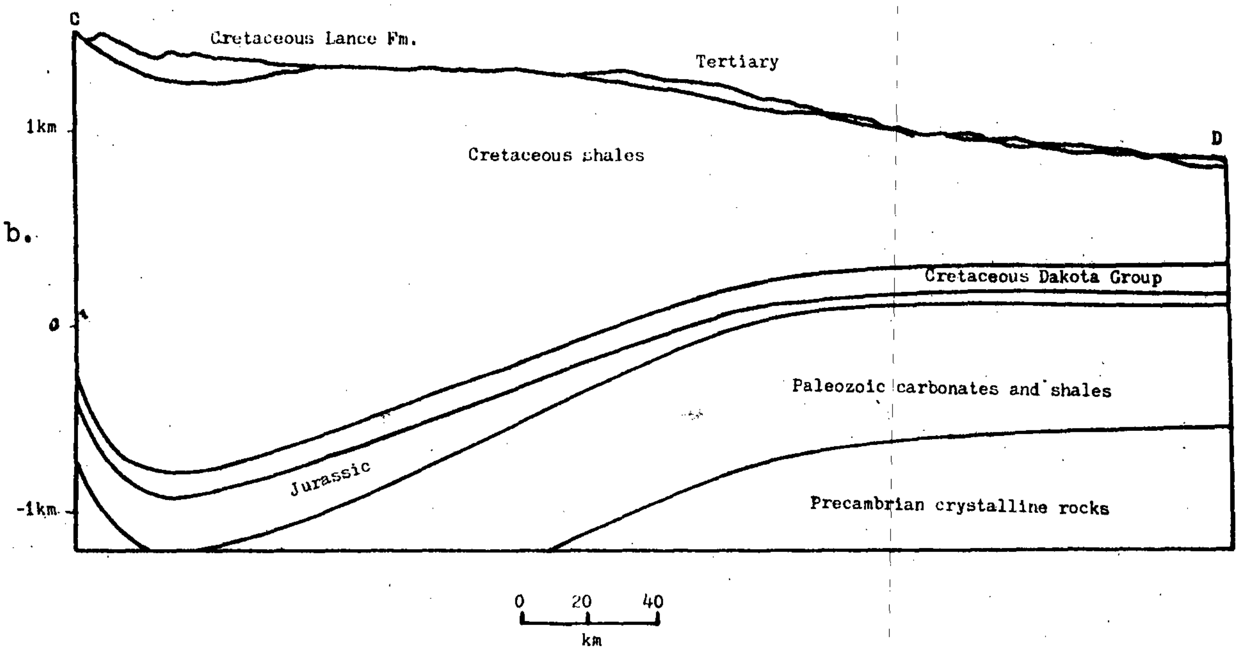
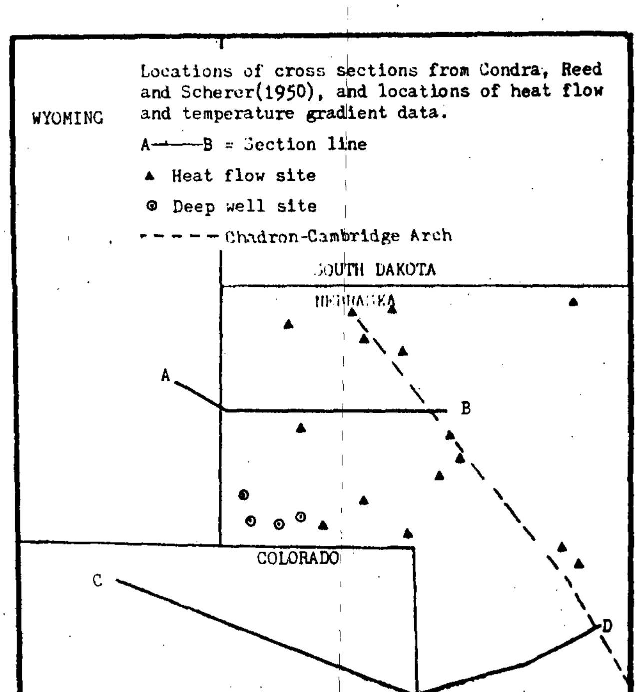


Figure 5. Structural cross sections in western Nebraska from Condra, Reed, and Scherer (1950).



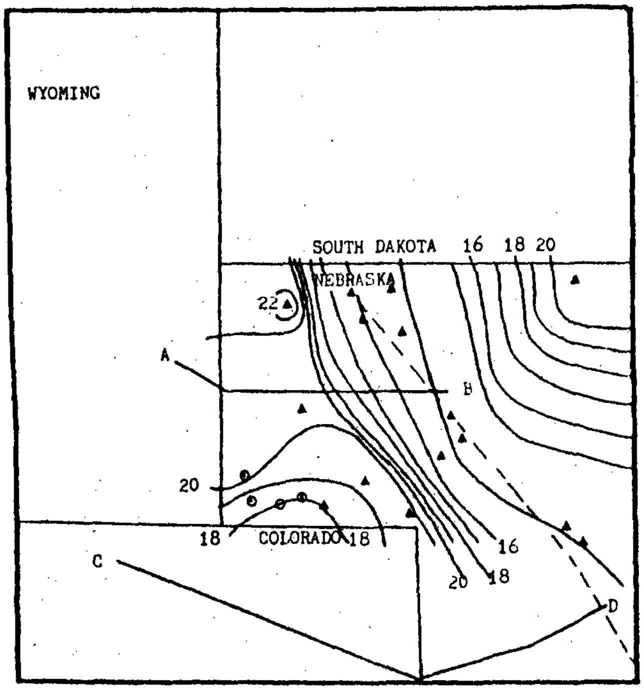


Figure 6 a. Hand contours of temperatures recorded at a depth of 150 m.

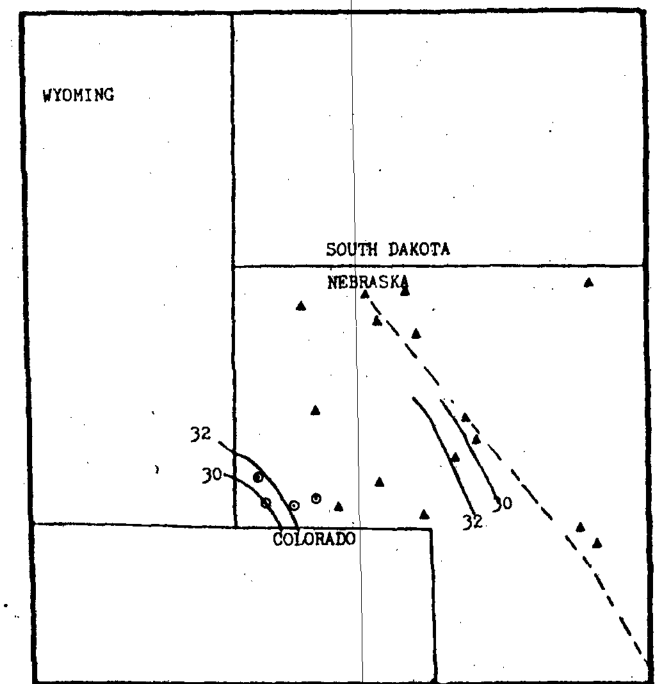


Figure 6 b. Hand contours of temperatures recorded at a depth of 500 m.

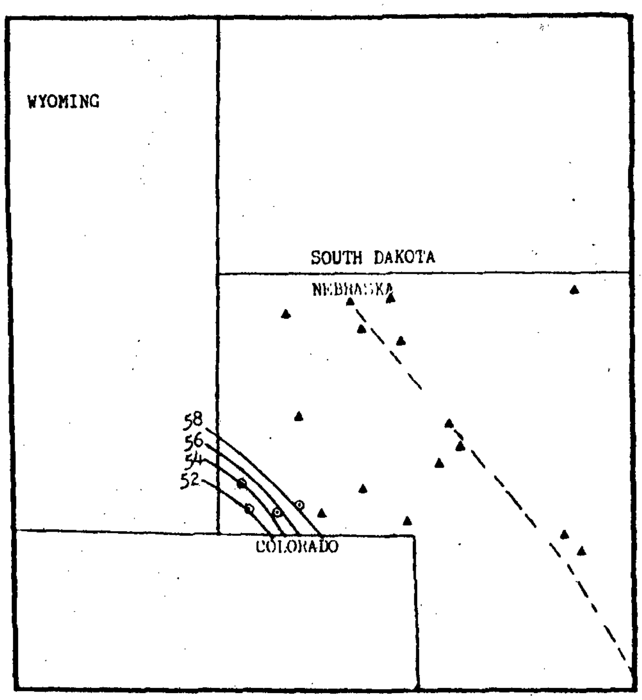


Figure 6 c. Hand contours of temperatures recorded at a depth of 1000 m.

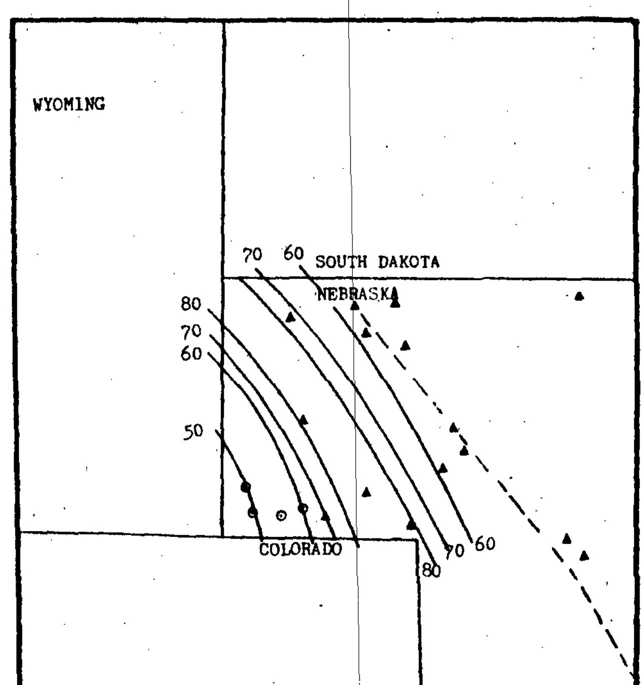


Figure 6 d. Hand contours of temperatures at a depth of 1000 m predicted by a finite difference model of heat flow with updip convection at 1 m yr^{-1} in the Dakota Group.

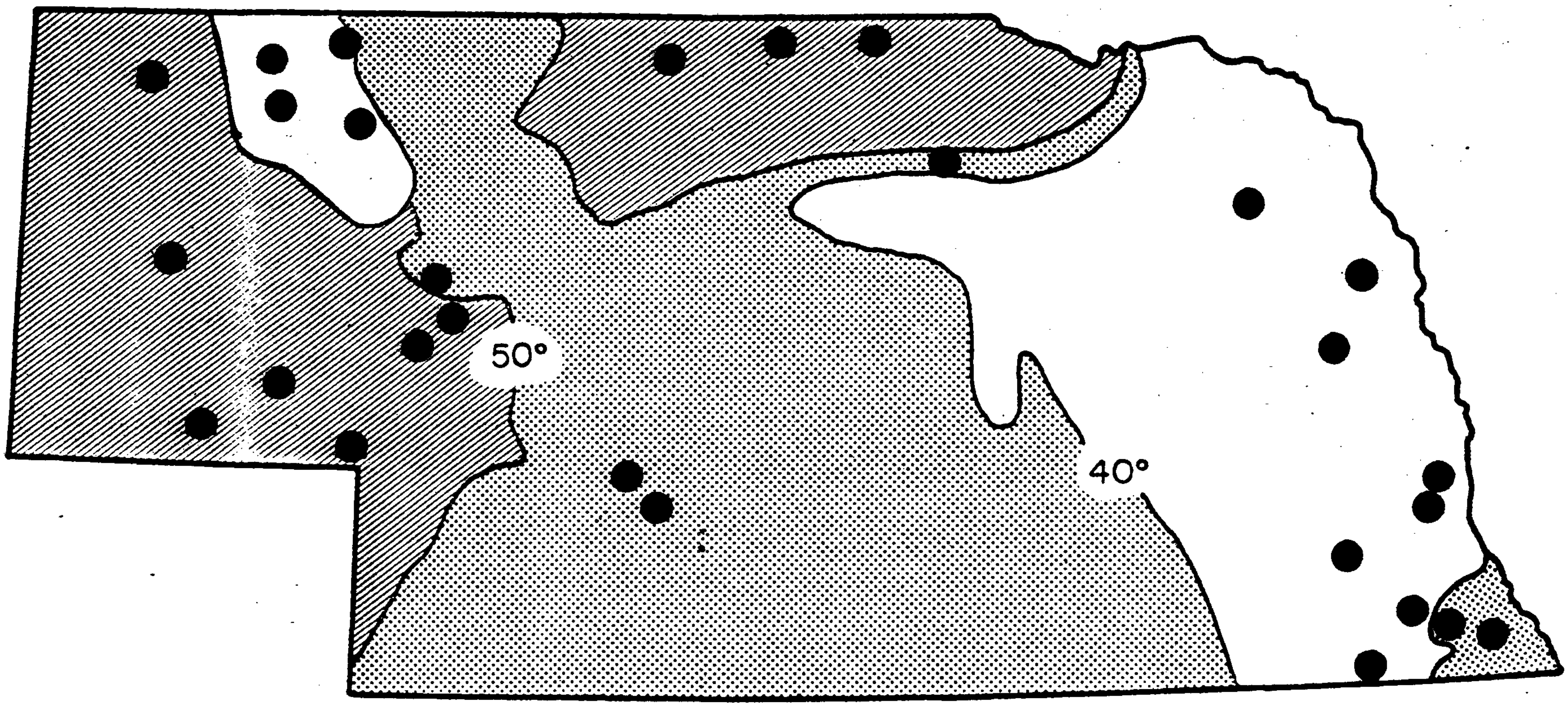


Figure 7. Temperature contours at a depth of 1 km as inferred from a synthesis heat flow data with stratigraphic and thermal conductivity data. The dots are heat flow sites.

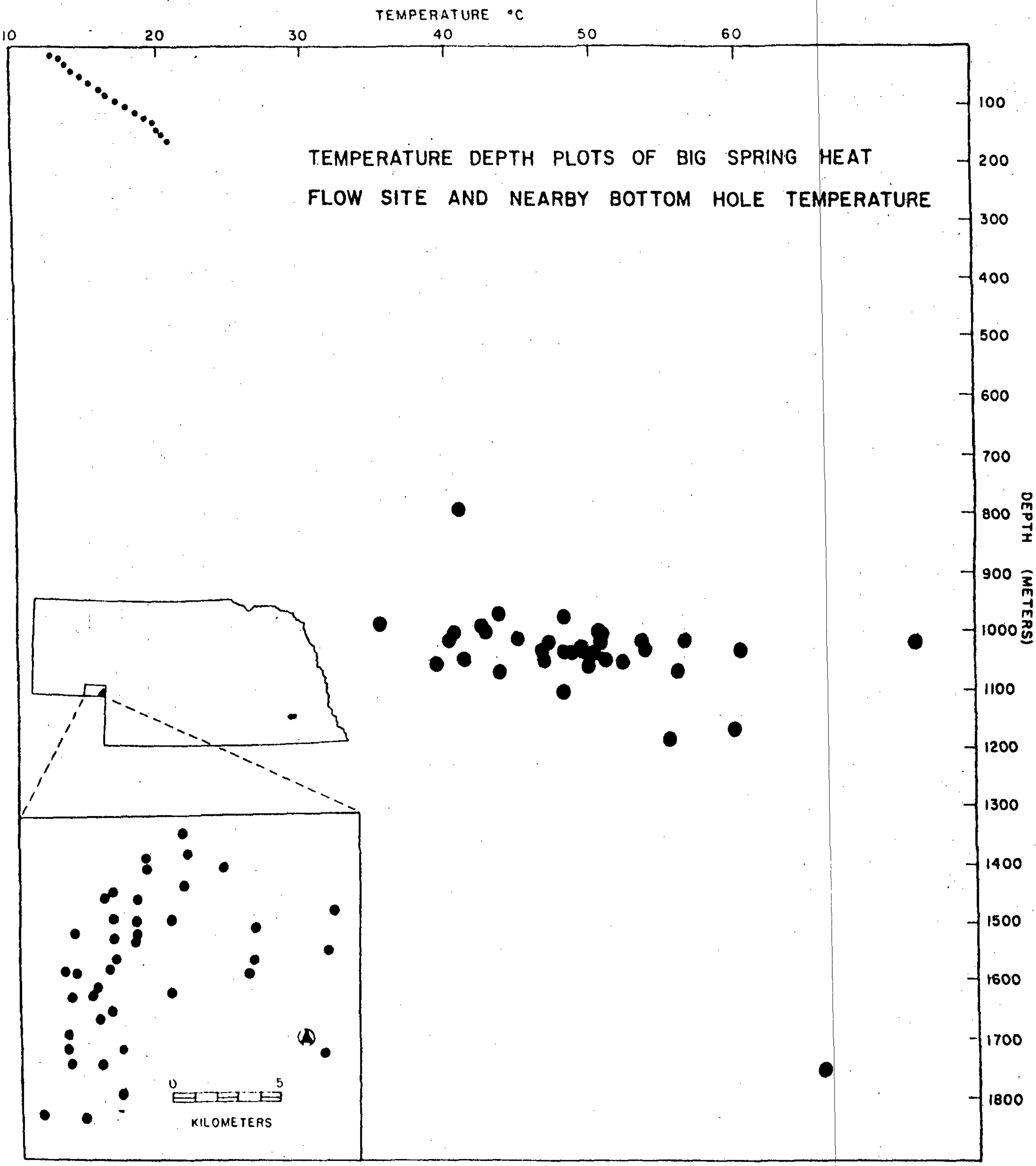


Figure 8. Comparison between the temperature gradient in a heat flow hole and bottom hole temperatures in Deuel County.

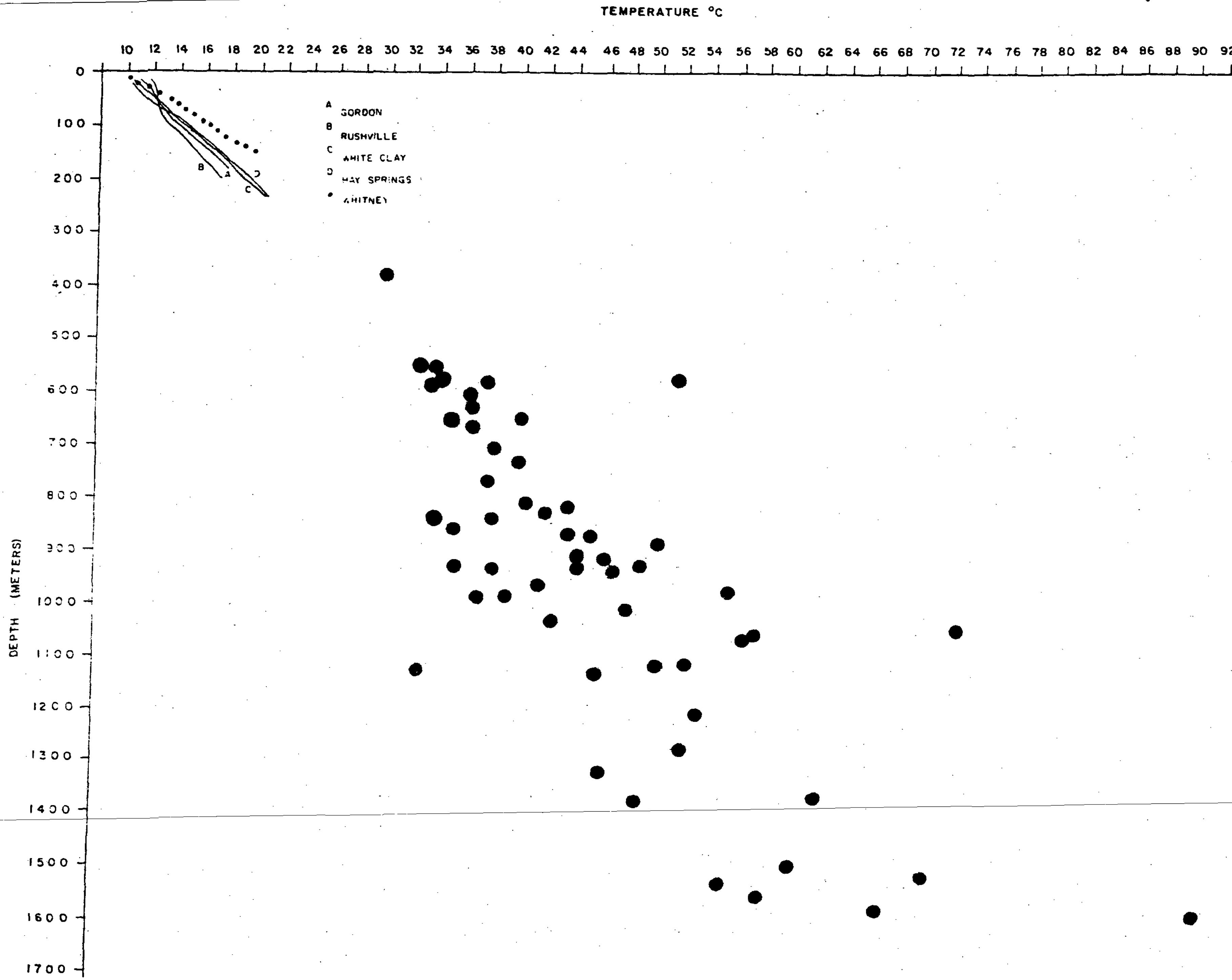


Figure 9. Comparison between gradients in heat flow holes and bottom hole temperatures in western Nebraska.

NEVADA RESOURCE ASSESSMENT PROGRAM - 1980

Presented at the State Assessment Meeting,
Glenwood Springs, Colorado, May 1981

by D.T. Trexler, T. Flynn,
B.A. Koenig, and J.L. Bruce

Introduction

During the past year the Nevada Resource Assessment Team has been working in three areas of Nevada: the first is a potential industrial heat application site - Golconda; the second area has potential for space heating - Hawthorne; and the third area has applications for space heating at a Naval Air Station - Fallon. Several exploration techniques have been employed during the term of the present contract including: chemical analyses of fluids, hydrogen and oxygen stable light isotope analyses, low sun-angle photography interpretation, micro-gravity surveys, two-meter temperature probe surveys, LANDSAT imagery analysis, and geologic reconnaissance.

Several of these techniques are discussed and the positive and negative aspects are addressed as they pertain to particular areas of investigation. The areas of investigation are shown in Figure 1.

Golconda

The Golconda study area is located in north-central Nevada and encompasses 1800 square kilometers. The study area is typical of the Basin and Range province of the western United States, characterized by a broad north-trending valley bounded by mountain ranges.

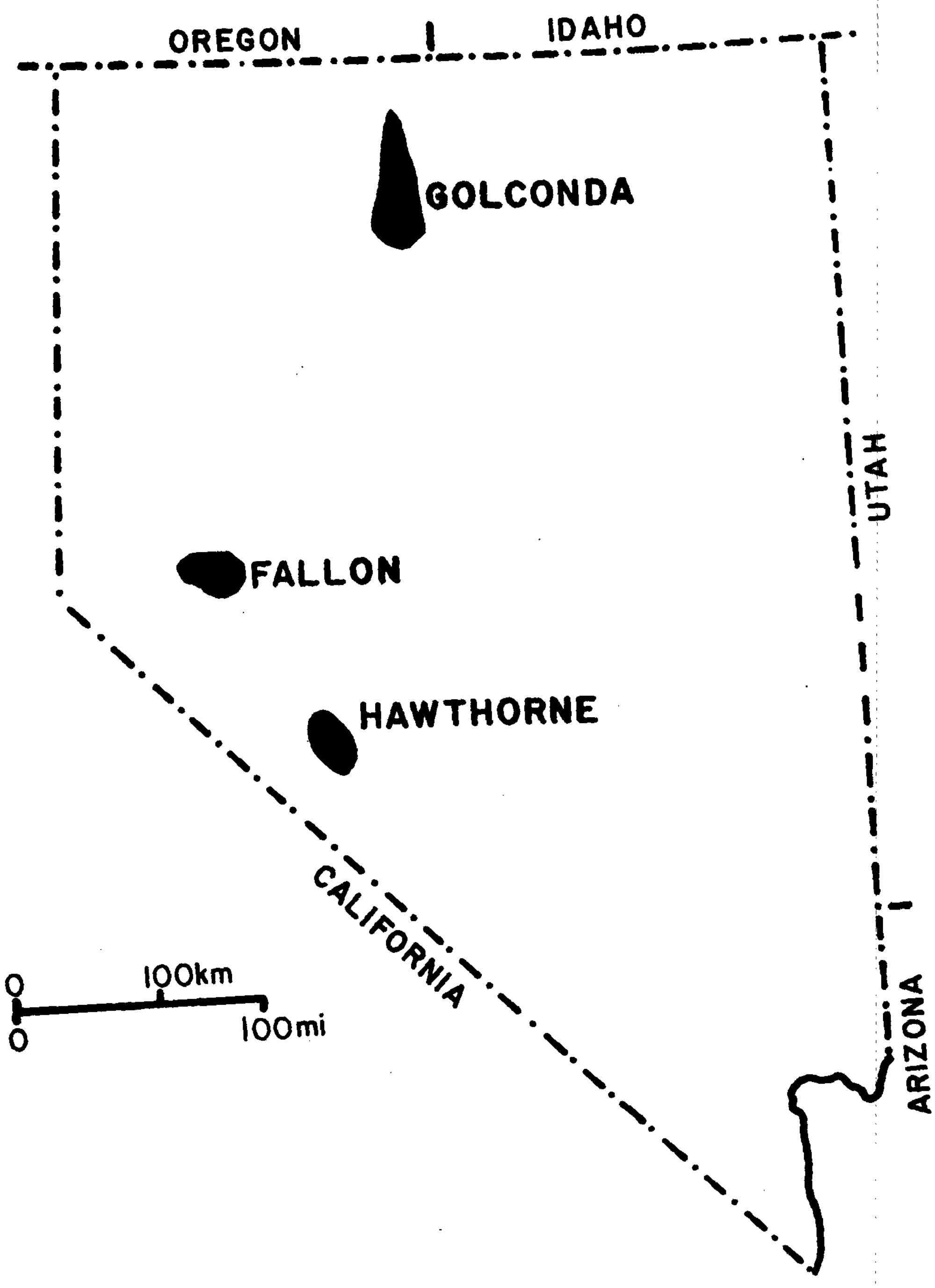


FIGURE 1. AREAS OF INVESTIGATION.

Figure 2 shows variations in the chemical compositions of thermal and non-thermal fluids within the study area. Thermal fluids tend to be richer in sodium-barcarbonate than non-thermal fluids. The difference in chemical composition is further exemplified in Figure 3, which indicates that the non-thermal fluids have a wide range in composition while the thermal fluids plot in two discrete fields. The large compositional variations for the non-thermal fluids is due in part to the effects of the local geologic environment.

Variations in temperature and the major cation and anion constituents for three hot springs and one thermal well for the past 35 years are shown in Figure 4. The more complete data sets indicate that within average analytical error the thermal fluids do not change drastically for the major cation and anion constituents. The minor variations which occur may be due to the different laboratories that provided the analyses rather than absolute chemical variations.

Trace element analyses of both thermal and non-thermal waters from the Golconda study area (fig. 5) indicate that lithium and barium are better indicators of thermal fluids than strontium or boron. The dashed lines in this figure represent non-thermal fluids. Analysis No. 2 (second from the left in each diagram) has a measured temperature of 13°C yet has trace element concentrations very similar to thermal fluids, in fact, this concentration of boron for this sample is higher than several of the thermal fluids.

The results of stable light isotope analyses of 18 samples

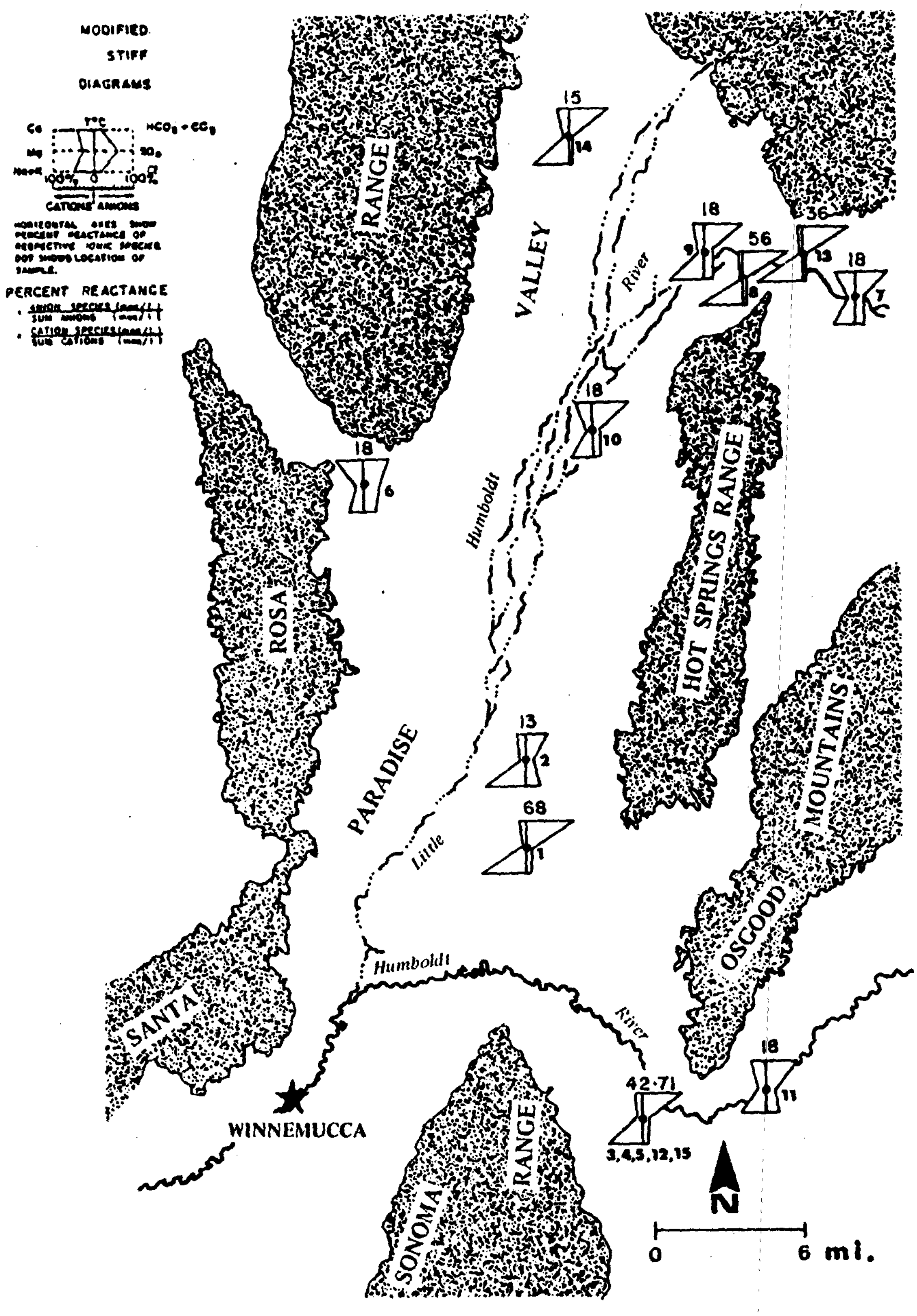


Figure 2. Chemical variations in fluids sampled throughout the Paradise Valley study area.

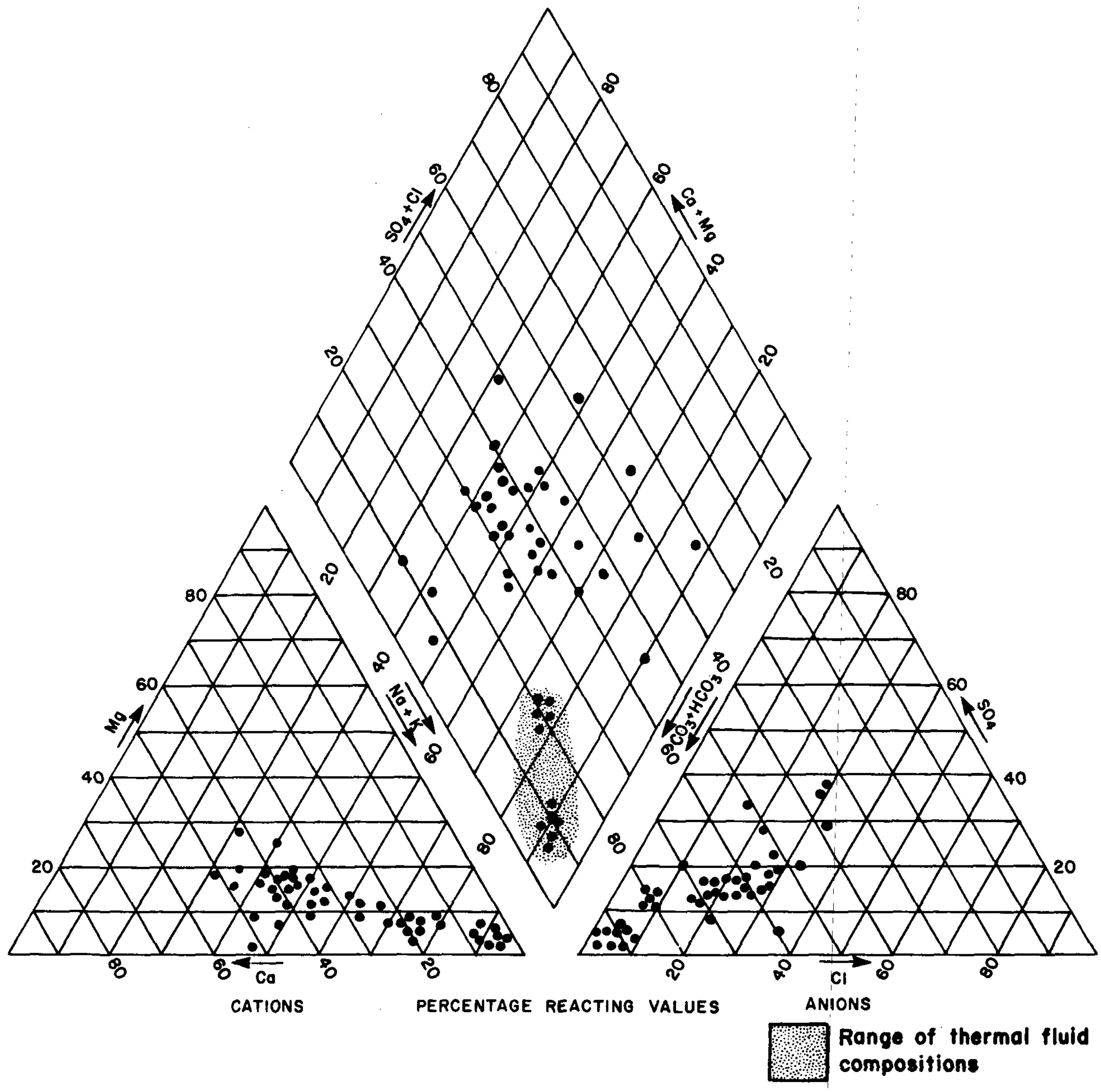


Figure 3 . Chemical characteristics of thermal and non-thermal fluids in the Paradise Valley study area.

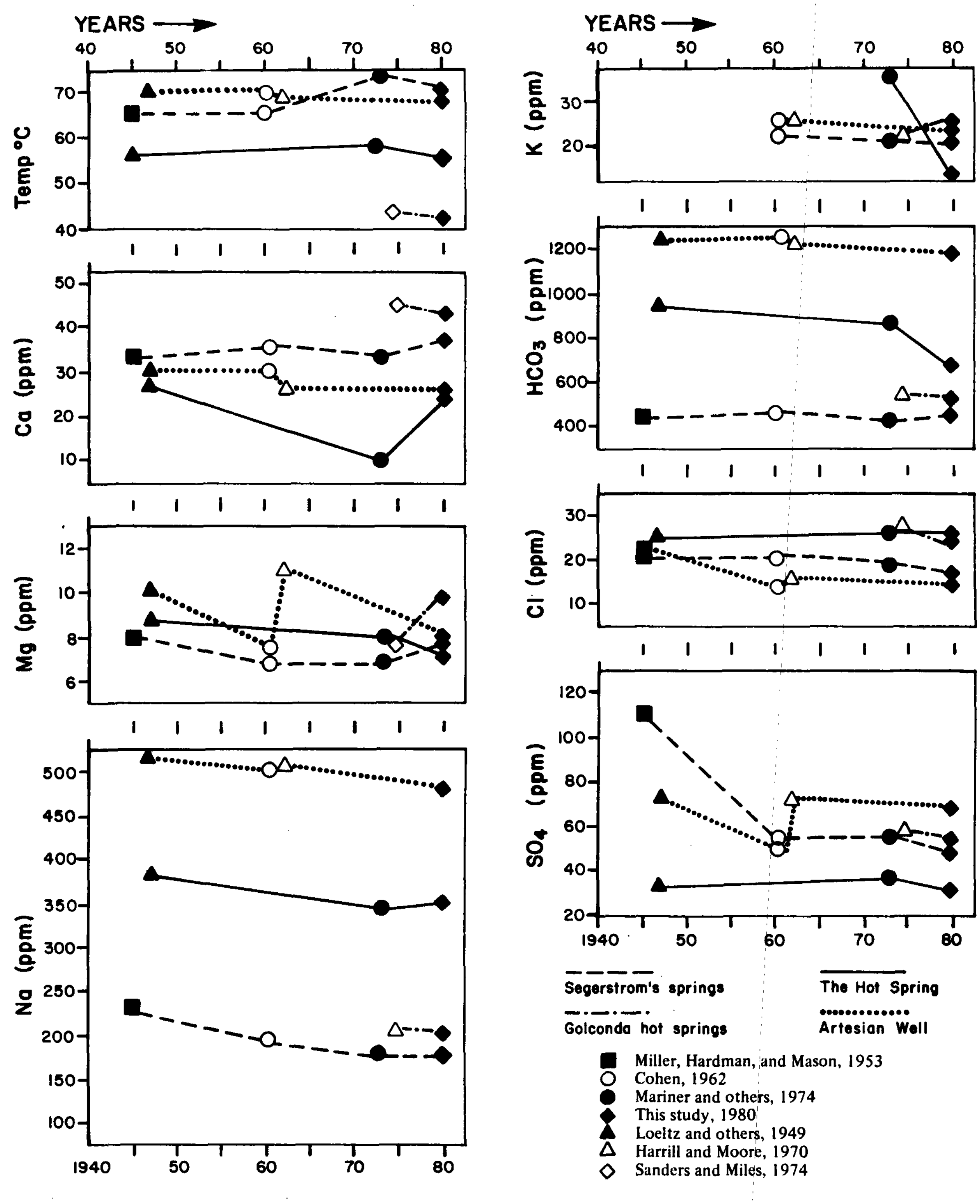


Figure 4. Chemical composition of selected thermal fluids in the Golconda study area, 1945-1980.

of both thermal and non-thermal fluids are presented in Figure 6. The non-thermal fluids are, in general, distinct from the thermal fluids. One exception is the well at Dutch Flat (A, in fig. 6), the same sample that displayed trace element compositions similar to thermal fluids. The similar isotopic characteristics may indicate that these fluids were at one time heated and have been cooled by conduction during transport.

Low sun-angle photography was flown over the Golconda study area to aid in the detection of subtle surface faulting. Several faults were mapped along the west side of the Hot Spring Mountains as a result of interpretation of this photography (fig. 7). Three sites selected for drilling 130 m (400 ft) test wells are also shown. All three holes encountered valley fill alluvium to total depth.

Figure 8 is a temperature gradient profile of well No. 3. A maximum temperature of 19°C was recorded at 130 m and the calculated maximum gradient is 40°C/100 m.

Hawthorne

The Hawthorne study area is located at the south end of Walker Lake in west-central Nevada. There are no surface manifestations of geothermal activity within 30 km of the town of Hawthorne; however, shallow to intermediate depth wells (100-150 m) have temperatures ranging from 24-37°C. A year ago the El Capitan Hotel and Casino drilled a well to irrigate a planned golf course, the well encountered water with a temperature of 96°C at 550 feet. The hole was completed to a T.D. of 750 feet

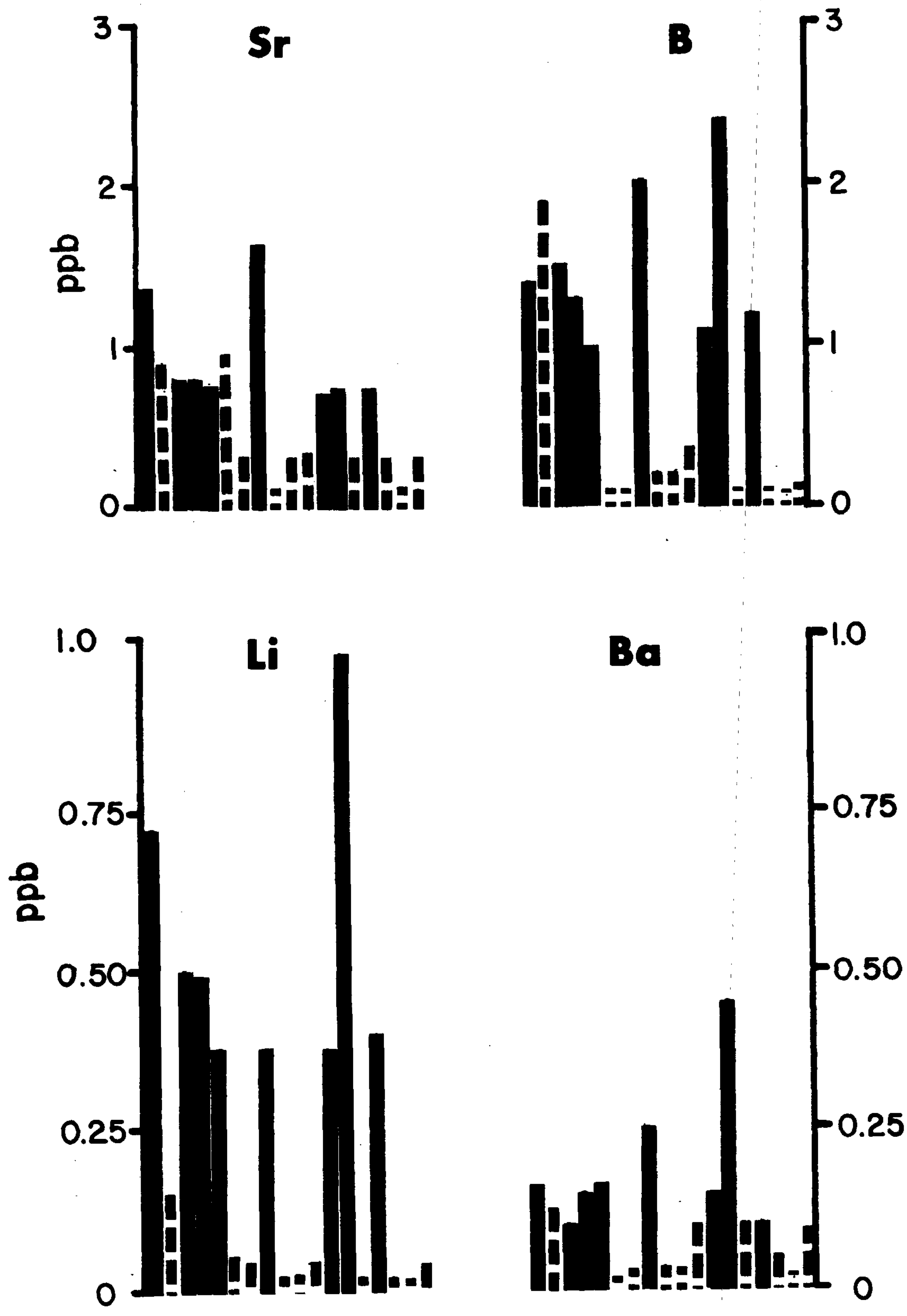


FIGURE 5. TRACE ELEMENT ANALYSES OF THERMAL (SOLID) AND NON-THERMAL (DASHED) FLUIDS IN THE GOLOGONDA STUDY AREA.

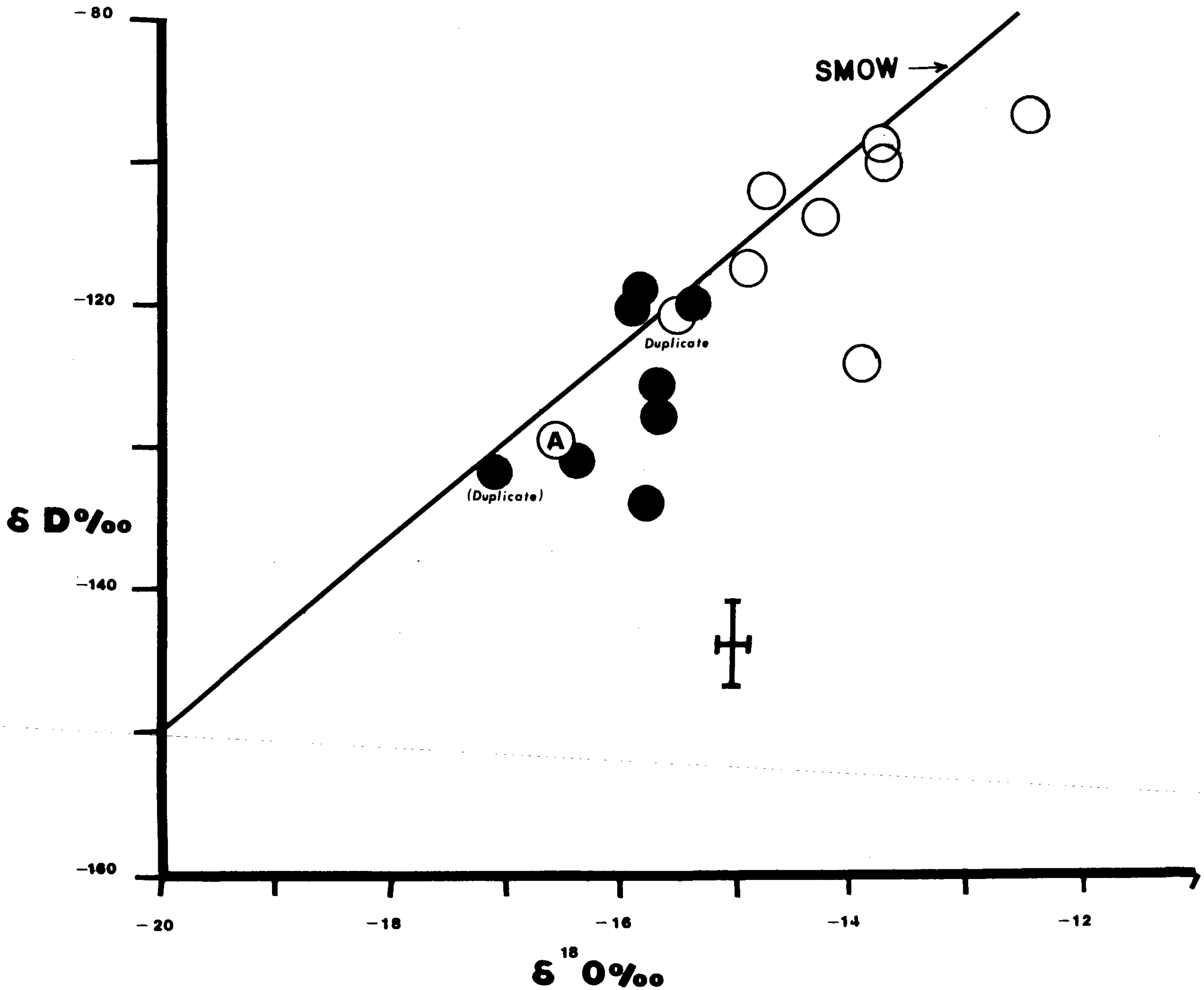


FIGURE 6. STABLE LIGHT ISOTOPE ANALYSES OF 18 SAMPLES .

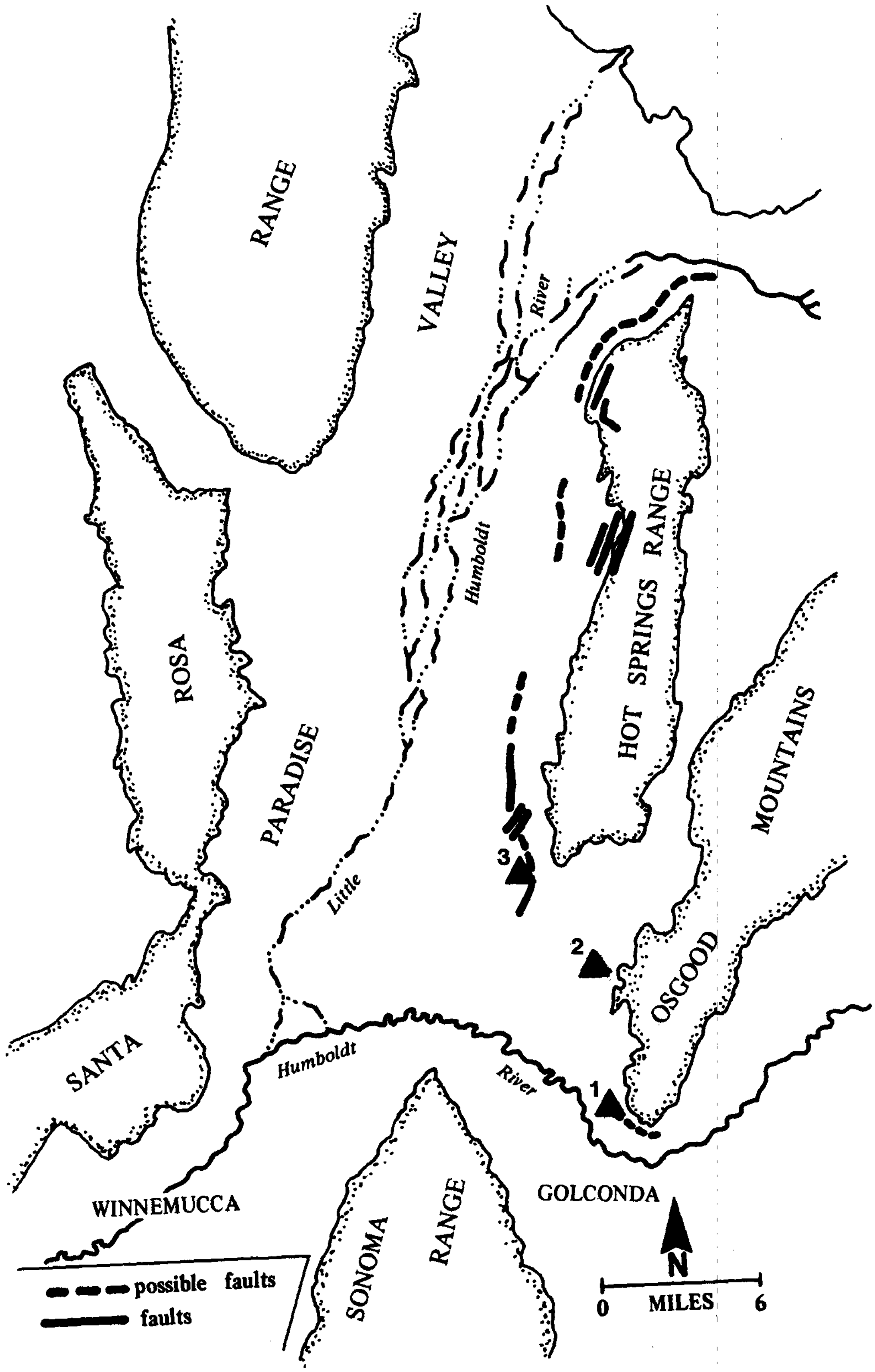


Figure 7 . Linear and curvilinear features interpreted from low sun-angle photography in the Paradise Valley study area.

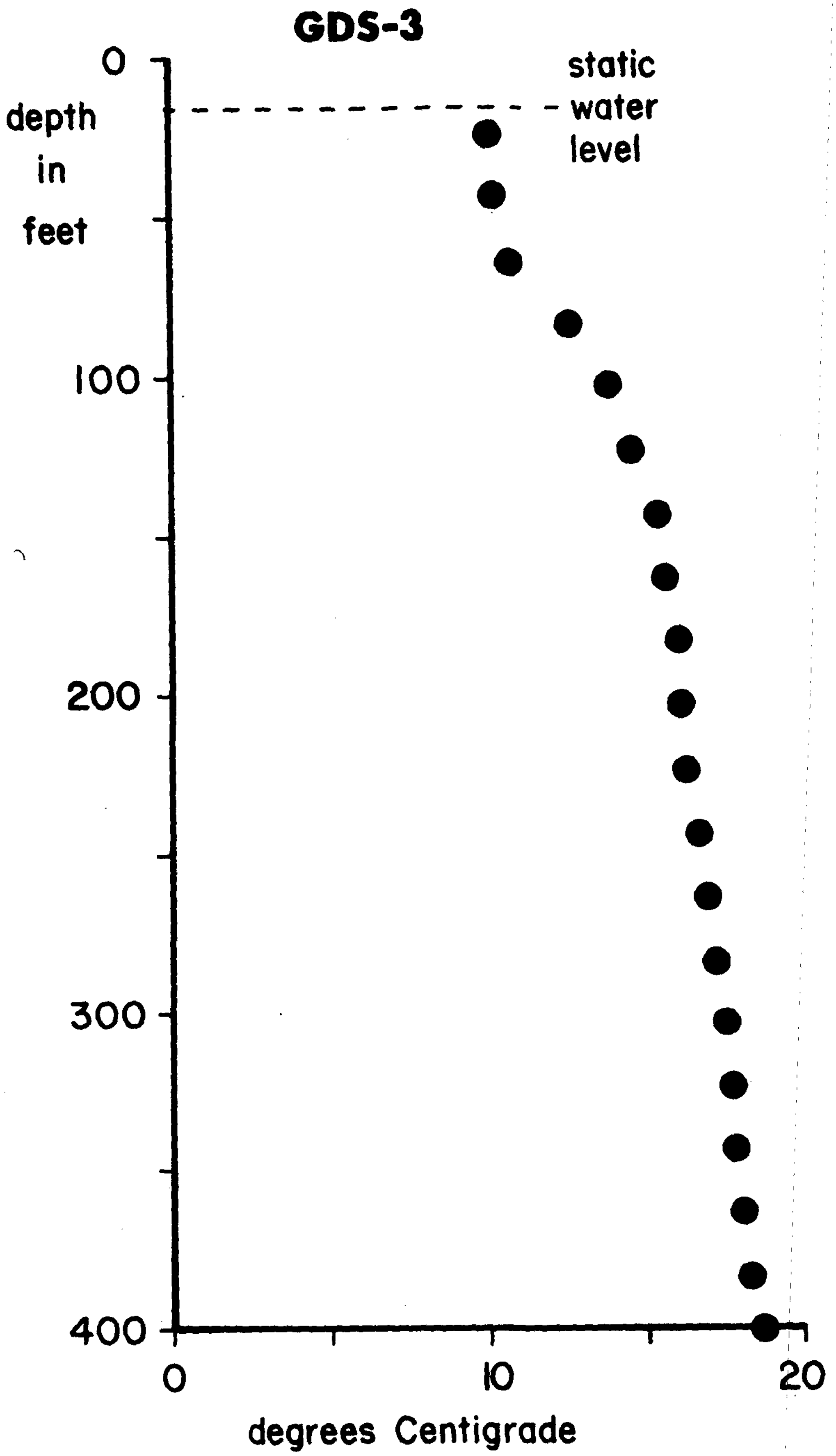


FIGURE 8. TEMPERATURE GRADIENT PROFILE OF GDS-3.

and when pump tested it flowed 94^{o}C water at 800 GPM for eight days.

A two-meter temperature probe survey was performed in the Hawthorne study area. Figure 9 shows the results of 96 two-meter probe locations. Normally, hole spacing was on the order of .25 miles in the area of this figure while a one mile spacing was used on a regional basis. In the southern portion of the figure the maximum temperatures recorded were 25^{o}C, the location of the El Capitan well. A north-trending isotherm closure of 24^{o}C can be noted west of the town of Hawthorne. This trend corresponds to trends noted in the low sun-angle photography as well as several gravity highs which were apparent from a micro-gravity survey performed in the area.

Figure 10 shows the equalibrium time for the two-meter probes at Golconda. Twenty-four hours after the hole is drilled the probes are in equalibrium and temperatures can be recorded that reflect the true soil temperature.

Chemical analyses of both thermal and non-thermal fluids indicate that thermal fluids tend to be richer in Na + K and SO_4 + Cl than non-thermal waters (fig. 11). However, several non-thermal fluids NAD 7, samples 3 and 5 have compositions within the range of thermal fluids. It should be pointed out that the first sample was not analyzed as part of this program and may represent laboratory error. This is also apparent for NAD 8 which had a temperature of 26.5^{o}C but was also not analyzed as part of this study.

Figure 12 represents data obtained from 15 analyses of hydrogen and oxygen stable light isotopes. The analysis with the

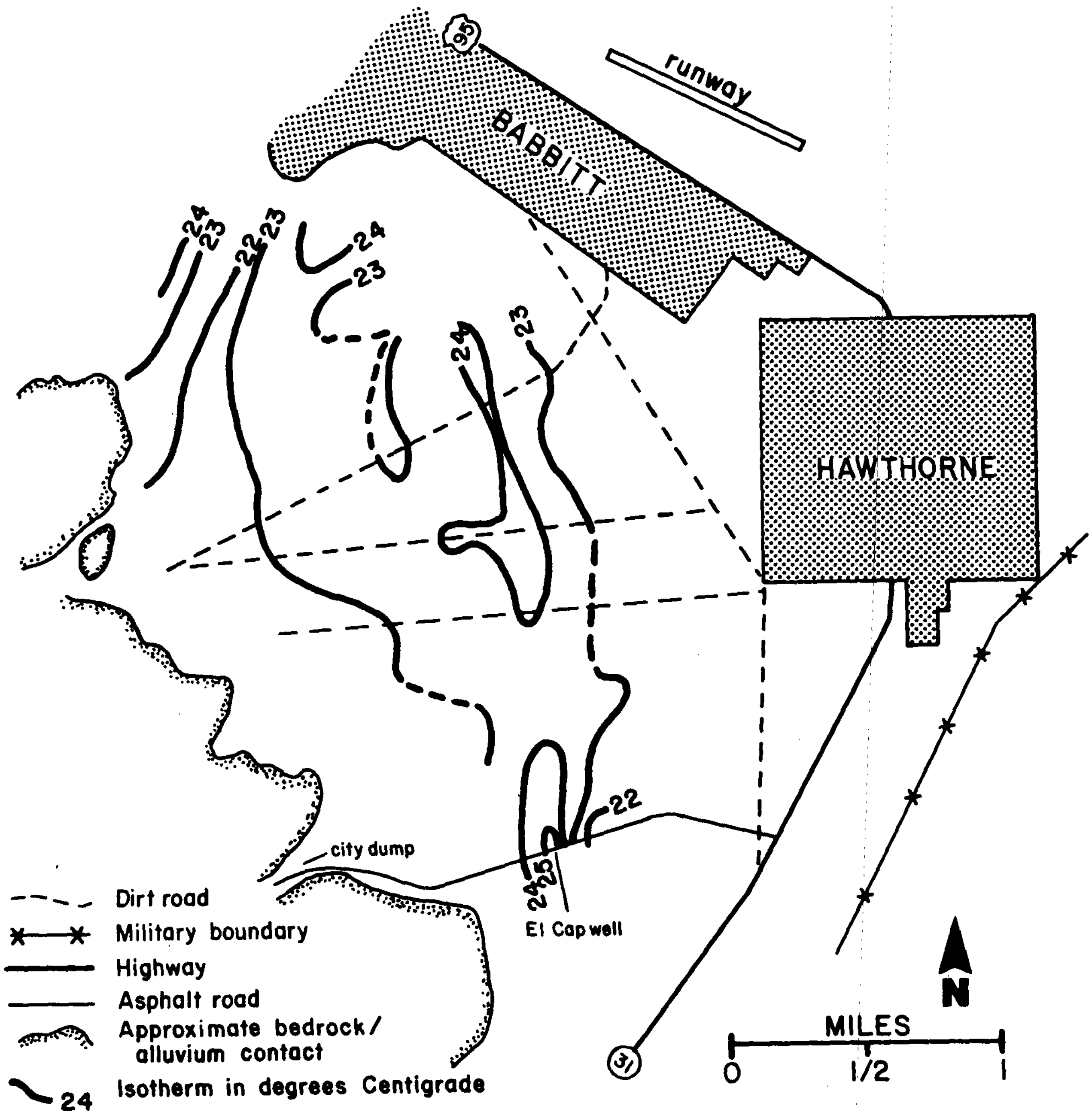
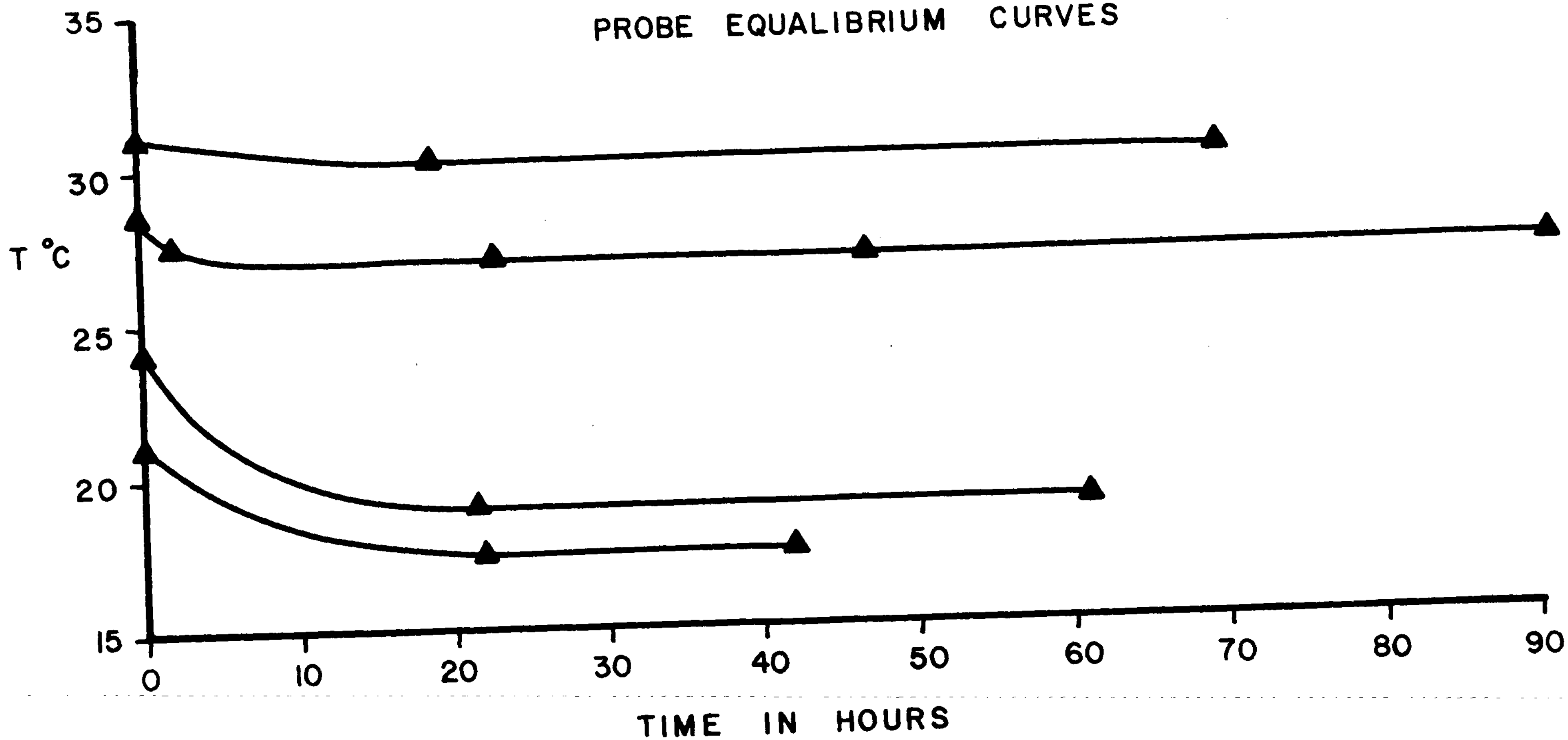
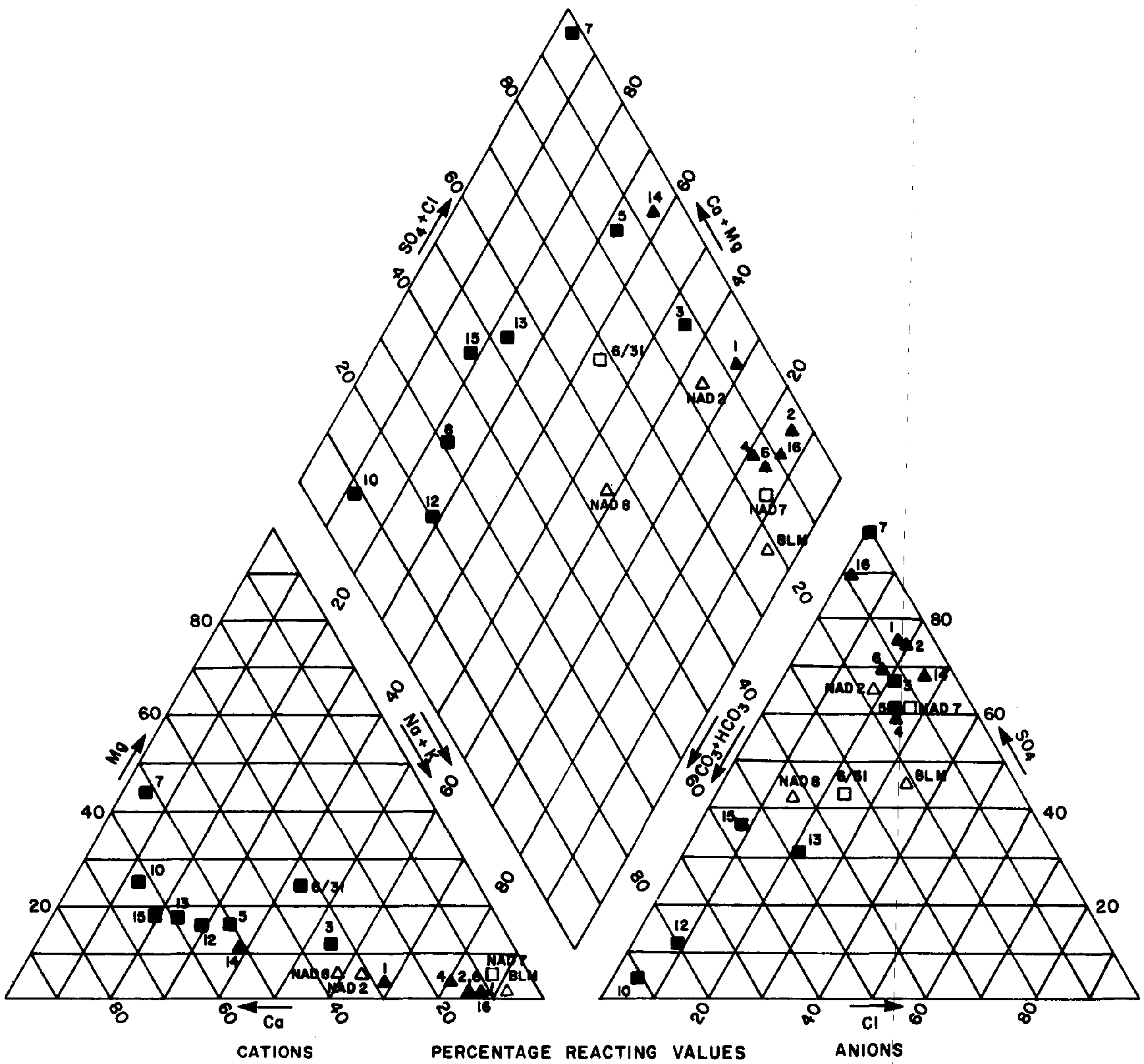


Figure 9. Possible isotherm configuration at a depth of 2 meters.

FIGURE 10.

TWO-METER DEPTH TEMPERATURE PROBE EQUALIBRIUM CURVES





1- HAAP well no. 1
2- HAAP well no. 5
3- HAAP well no. 6
4- HAAP well no. 3
5- HAAP well no. 4
6- El Cap. well
7- Alum Creek
8- Cottonwood Cr. 4800'
10- Cottonwood Cr. 9000'
12- Coreyville
13- Corey Canyon - Cottonwood Springs
14- HUD well no. 5
15- Corey Canyon well (HUD no. 4 ?)
16- Walker Warm Springs
BLM Whiskey Flat Well
6/31b irrigation well Whiskey Flat
NAD 7
NAD 8
NAD 2

■ **COLD (Less than 25 degrees Centigrade)**

▲ **HOT (Greater than 25 degrees Centigrade)**

Figure 11. Chemical characteristics of thermal and non-thermal fluids in the Hawthorne study area.

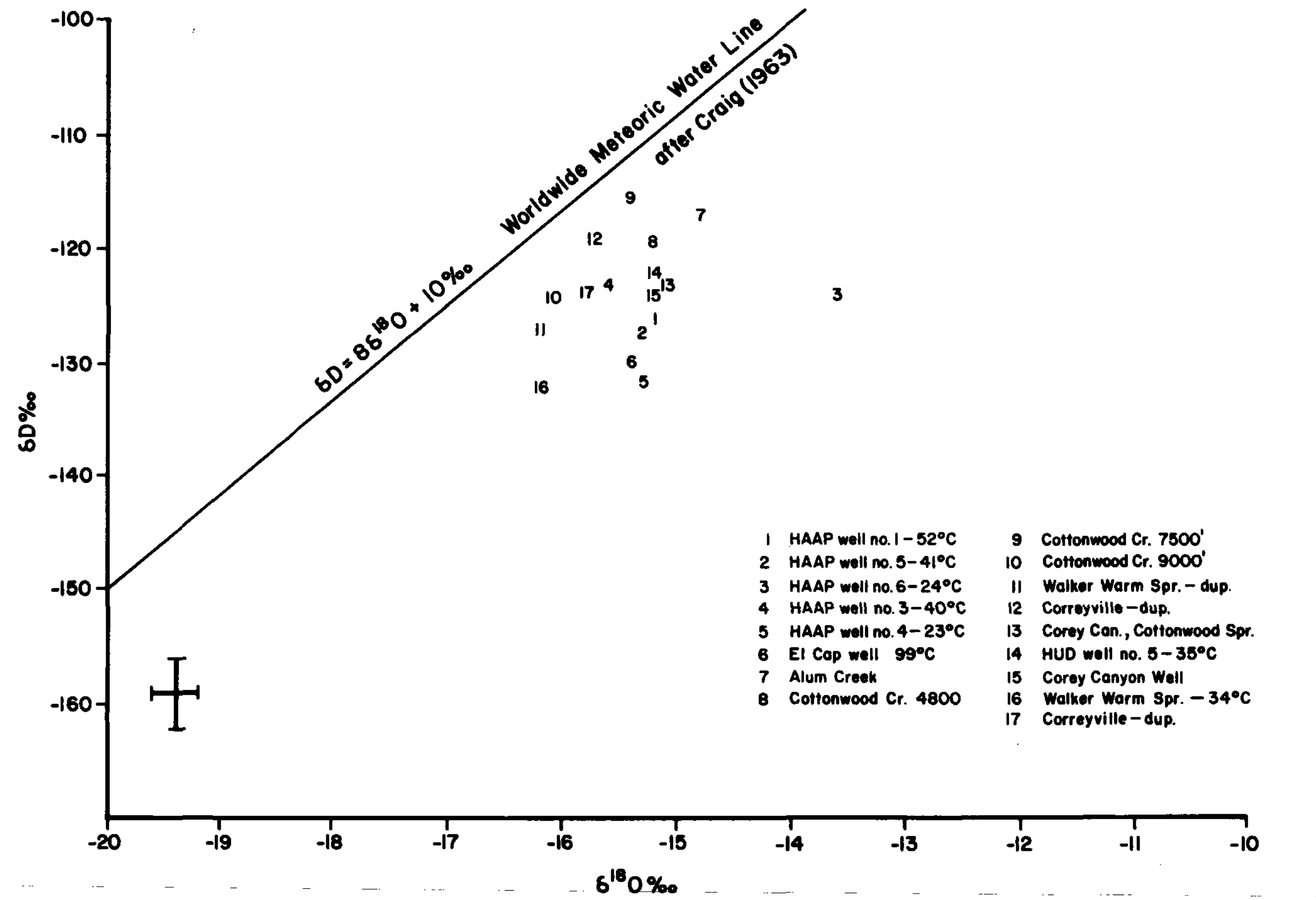


FIGURE 12. STABLE LIGHT ISOTOPIC COMPOSITION OF HAWTHORNE AREA WATERS.

large oxygen shift when compared to the meteoric water line (analysis No. 3) is not due to water-rock interaction but rather to a cracked sample bottle which allowed contamination by atmospheric oxygen. The overall variation between thermal and non-thermal fluids is not great and is even less when analytical error limits are considered. Surface water samples tend to be shifted toward the meteoric water line. Isotopic analyses did not prove useful in the Hawthorne study area.

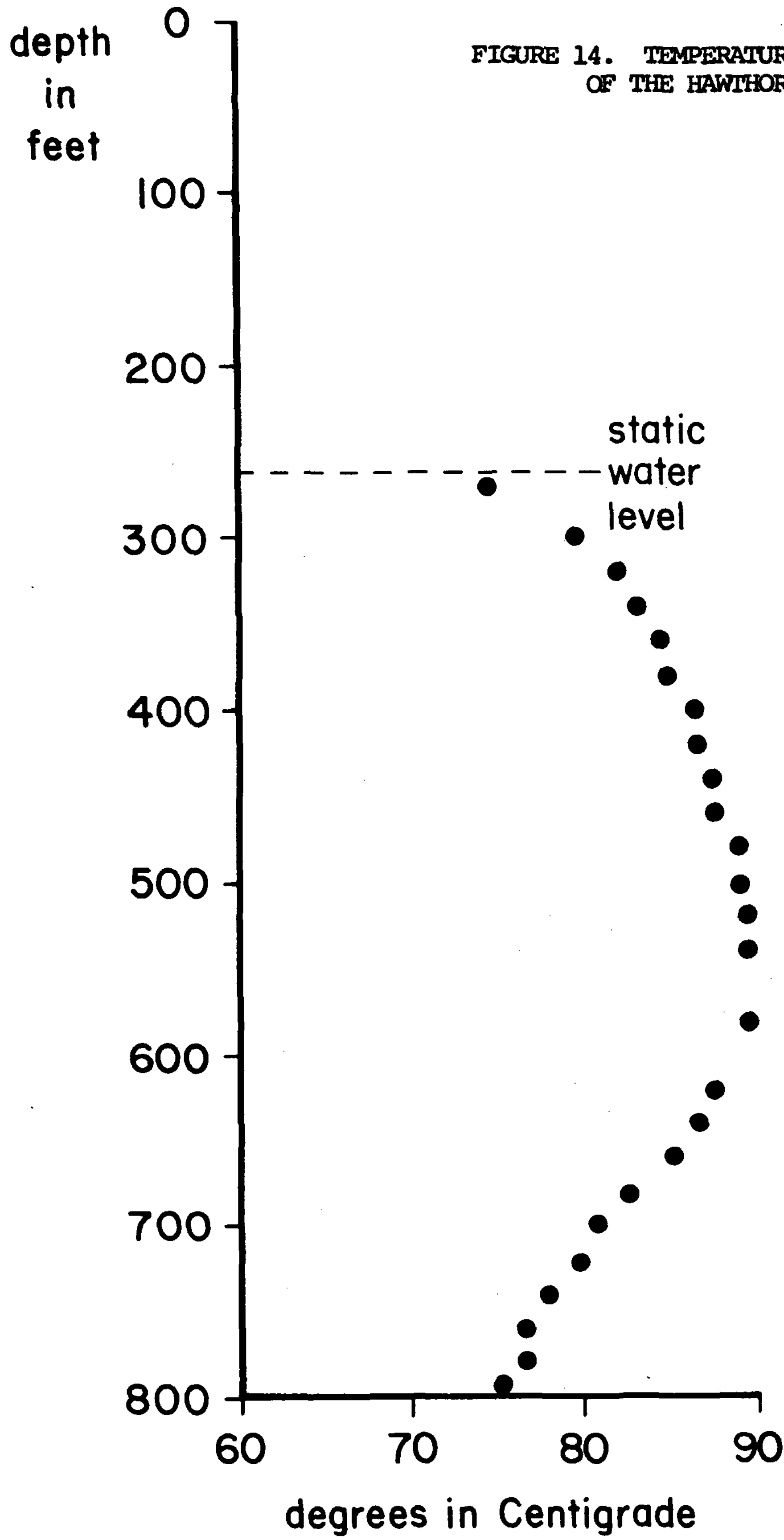
The photograph of the Hawthorne area (fig. 13) shows the location of a 265 meter hydrologic test hole. The siting of this hole was based mainly on the interpretation of gravity data, low sun-angle photography and two-meter depth temperature probe surveys. Soil mercury and chemical data provided more regional information but did not provide significant contribution to test hole siting. The location is approximately 1.5 miles north of the El Capitan well.

Temperatures recorded in the test hole exceeded 90°C. The hole was blind cased with 3" black pipe. It is anticipated that temperatures similar to those recorded in the El Capitan well will be maintained if the test well is produced. The temperature gradient profile shown in Figure 14 indicates the hottest zone to be between 480 and 580 feet which corresponds to the alluvium bed-rock contact. It appears that the thermal fluids rise by convection along the range bounding fault and migrate laterally down the hydrologic gradient to the east.

The Army Ammunition Plant on whose land the well was drilled is submitting proposals to Headquarters requesting funding to

FIGURE 13. LOW SUN-ANGLE PHOTOGRAPH OF THE HAWTHORNE AREA SHOWING THE LOCATION OF THE 235 METER TEST WELL.

FIGURE 14. TEMPERATURE GRADIENT PROFILE OF THE HAWTHORNE TEST WELL.



develop this resource for space heating of the base.

Fallon

The Resource Assessment Team was requested by the Department of Energy (DOE) to evaluate the geothermal resource for the Navy at the Fallon Naval Air Station. Several areas of potential high temperature resources surround the study area on the east (Salt Wells), northeast (Stillwater), northwest (Soda Lake), and to the south (Lee Hot Springs). No surface thermal manifestations exist in the immediate study area which includes 180 square miles.

Soil mercury analyses showed better correlation with major structures in the Fallon area than similar surveys in Golconda and Hawthorne (fig. 15). The mercury anomalies have trends with northwest and northeast orientations. Several soil samples had Hg concentrations in excess of 3000 ppb. The northwest trends probably are related to the Walker Lane structures while the northeast trends indicate an affiliation with the Carson Lineament and Midas Trench systems.

Similar orientations are apparent in the next figure which shows faults mapped from low sun-angle photography (fig. 16). Temperature gradient data from nine wells on the Fallon Naval Air Station range from $13^{\circ}C/100$ m in the north to $42^{\circ}C/100$ m at the southern boundary of the Naval Reservation. Well depths ranged from 380 to 505 feet. The higher temperature gradients appear to be associated with structures south of the Naval Air Station and it is our recommendation that the optimum site for a geothermal well for space heating should be located near the southern boundary of the Reservation.

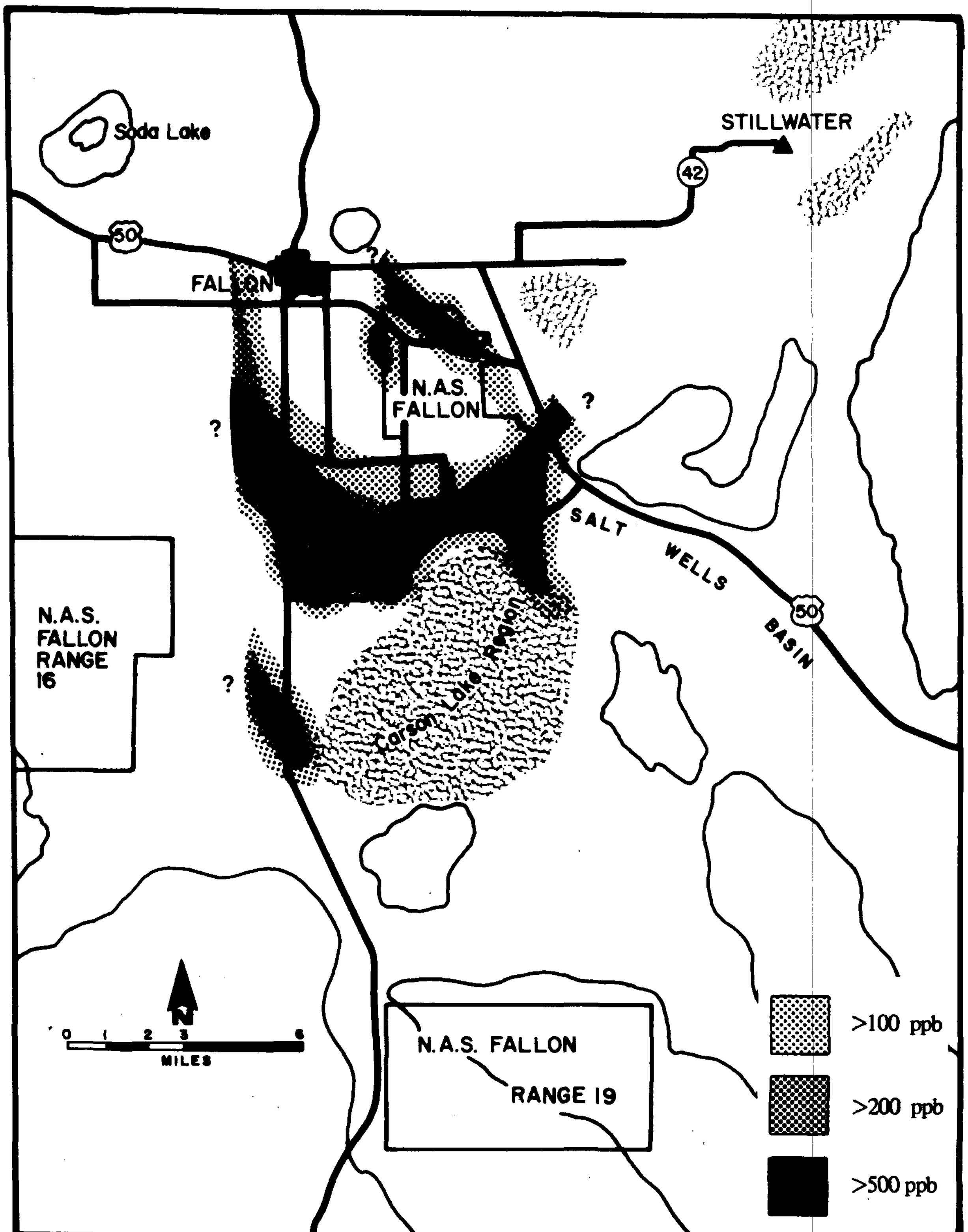


Figure 15. Soil mercury anomalies.

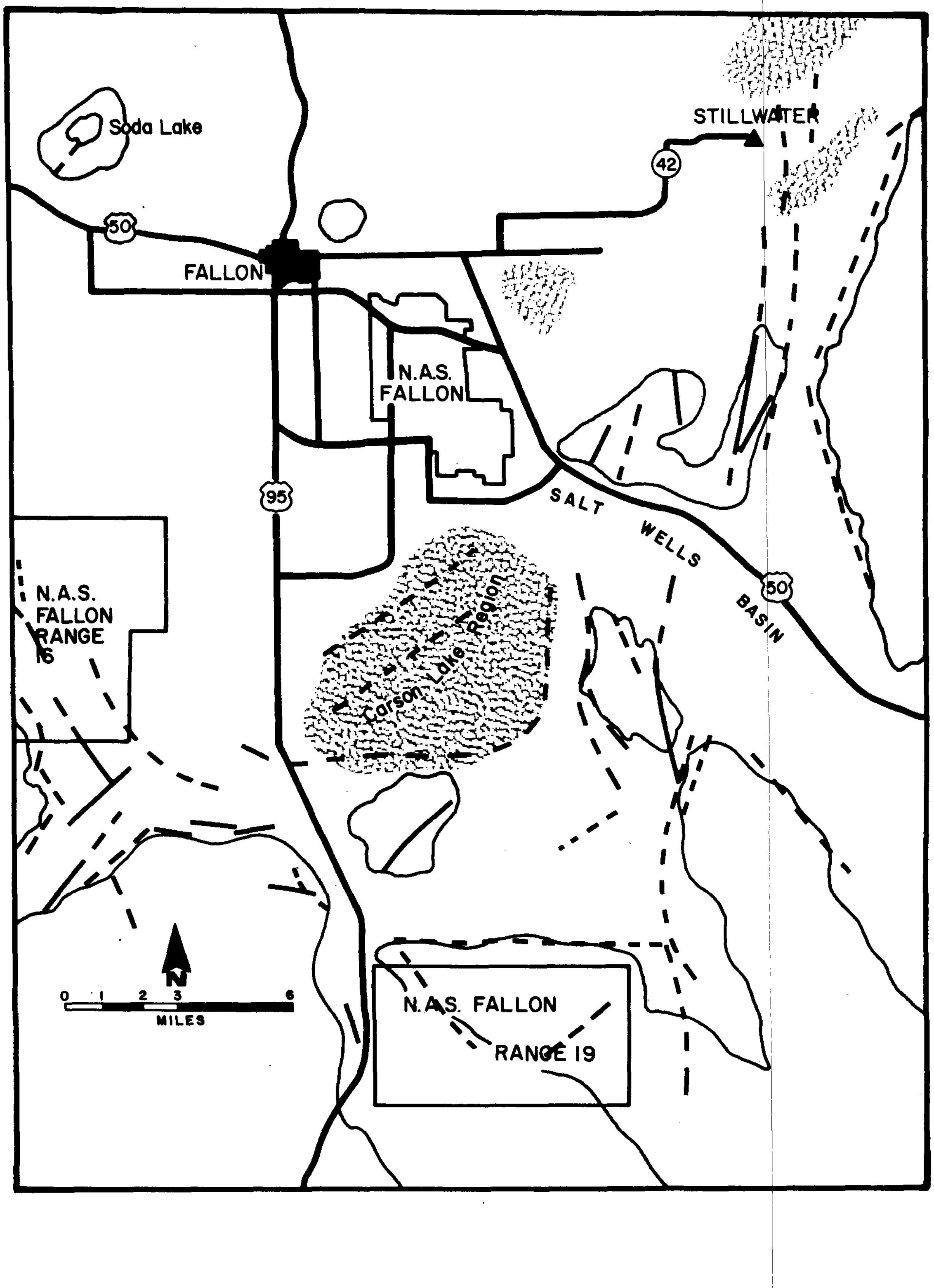


Figure 16. Faults (dashed where inferred).

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GROUNDWATER CONVECTION MODEL FOR RIO GRANDE RIFT GEOTHERMAL RESOURCES

Paul Morgan,[1] Vicki Harder,[2]
Chandler A. Swanberg[3] and Paul H. Daggett[2]

[1]Lunar and Planetary Institute, 3303 NASA Rd 1, Houston, TX 77058
[2]Dept. Geol. Sci., Univ. Texas El Paso, El Paso, TX 79968
[3]Depts. Physics/Earth Sciences, New Mexico State Univ., Las Cruces, NM 88003

ABSTRACT

It has been proposed that forced convection, driven by normal groundwater flow through the interconnected basins of the Rio Grande rift is the primary source mechanism for the numerous geothermal anomalies along the rift. A test of this concept using an analytical model indicates that significant forced convection must occur in the basins even if permeabilities are as low as 50-200 millidarcies at a depth of 2 km. Where groundwater flow is constricted at the discharge areas of the basins forced convection can locally increase the gradient to a level where free convection also occurs, generating surface heat flow anomalies 5-15 times background. A compilation of groundwater data for the rift basins shows a strong correlation between constrictions in groundwater flow and hot springs and geothermal anomalies, giving strong circumstantial support to the convection model.

INTRODUCTION

Harder et al. (1980) and Morgan and Daggett (1981) have presented evidence indicating that the primary source mechanism for the numerous geothermal anomalies along the Rio Grande rift, excluding the Jemez Mountains, is forced convection driven by inter and intrabasin groundwater flow. This deduction has important implications for geothermal development, as it implies a rapid decrease in geothermal gradient with depth in the anomalies, and no potential for electricity generation, if 3 km (10,000 ft) and 200°C (392°F) are taken as the economic cut-off values for maximum depth and minimum temperature respectively. The geothermal systems may have great potential, however, for direct heat utilization, and future electricity generation with binary, lower temperature systems. It is important, therefore to test as fully as possible the groundwater convection hypothesis to guide geothermal planning and exploration in the rift basins. In this paper we present both a theoretical and practical test of the groundwater convection model.

THEORETICAL TEST OF FORCED CONVECTION HYPOTHESIS

Whenever mass is transferred across a temperature difference, heat is also transferred by convection. In permeable strata of the earth, therefore, any component of groundwater flow parallel to the temperature gradient will convect heat. If the flow is caused by thermally induced buoyancy in the groundwater, the heat transfer is termed free convection. Where flow is driven by an externally derived hydraulic gradient, forced convection is said to occur. Analyses of free convection (e.g. Donaldson, 1962) indicate that a high thermal gradient must be present across the convecting region to drive free convection. It is therefore a secondary enhancement rather than a primary source of a geothermal anomaly. Harder et al. (1980) argue that variations in rock properties or magmatic intrusions are unlikely to cause the numerous geothermal anomalies and hence drive free convection in the Rio Grande rift. Forced convection must occur with the regional north to south flow of groundwater through the rift basins. The only question is whether or not forced convection is a sufficiently efficient heat transfer mechanism to explain the observed anomalies.

To model forced convection it is first necessary to determine the groundwater flow field. Unfortunately although it is technically feasible using numerical methods, there is insufficient permeability information from any of the rift basins, especially for the deeper portions of the basins, to produce a detailed flow model. We have therefore approached the problem in reverse, using an analytical solution to determine at what level of permeability forced convection becomes significant.

We have used the analytical solution for forced groundwater convection in a basin given by Domenico and Palciauskas (1973). These workers derived an approximation to the temperature perturbation caused by groundwater flow in a two dimensional rectangular basin, with groundwater flow along the length L of the basin driven by a hydraulic gradient along the basin derived from a sloping water table, which conforms to the equation $A - B \cos(\pi x/L)$, where A is the mean water table elevation, 2B is the total drop in water table elevation along the basin, and x is distance along the basin. Permeability is expressed as hydraulic conductivity K, and is assumed to be constant down to a depth D in the basin, below which the formations are impermeable. Steady state conditions are assumed to apply. If

Ts is the ground surface temperature, G the undisturbed temperature gradient, and α is a mixed thermal diffusivity (thermal conductivity of the saturated rock divided by the product of the groundwater density and specific heat), Domenico and Palciauskas (1973) give the temperature T (x,z) at a point (x,z) in the basin and surface temperature gradient Gs(x) at distance x along the basin as:

$$T(x,z) = Ts + G(z - D) - GKB/2\alpha \cdot [\cos(\pi x/L)/\cosh(\pi D/L)] \cdot \{(D - z)\cosh(z/L) + (L/\pi).[\sinh(\pi(z - D)/L)/\cosh(\pi D/L)]\},$$

and

$$Gs(x) = G/2[2 + (KB/\alpha)\cos(\pi x/L)\tanh^2(\pi D/L)].$$

It should be noted that the approximations used in the derivations of these equations become invalid as the magnitudes of the temperature or gradient perturbations approach the undisturbed values. The equations are accurate and sensitive, however, to the flow, and hence permeability threshold at which forced convection becomes significant.

For models of forced convection in the Rio Grande rift basins, a depth of 2 km to basement has been assumed, which data presented by Seager and Morgan (1979) indicate is a conservative estimate. A hydraulic gradient of .001 has been used, based on the slope of the Rio Grande between Socorro and Las Cruces. Two basin lengths have been tested, 20 and 100 km, both with ground surface temperatures of 20°C and undisturbed gradients of 40°C/km. The results of these models are shown in Figures 1 and 2. For the 100 km basin model significant convection occurs with hydraulic conductivities of the order of 1 to 2 X 10^{-4} cm/s (100 to 200 millidarcies). For the 20 km basin model convection is effective with hydraulic conductivities of 5 X 10^{-5} cm/s (50 millidarcies) or less. These hydraulic conductivities are down in the semipervious permeability domain of silt, stratified clay and oil rocks (Bear, 1972, p. 136) and are very realistic for a depth of 2 km in rift basins.

Are the results of the simple analytic models applicable to the complex groundwater flow systems in the Rio Grande rift basins? In detail, the answer is obviously no; local flows will be controlled by low permeability aquifers. Deep recharge and discharge will probably occur primarily in areas with relatively high vertical permeability and/or locally high hydraulic gradients. On a regional basis, however, the models are applicable, as by Darcy's Law, deep flow depends only on the hydraulic gradient transmitted to depth, and is independent of shallow flow conditions. We conclude, therefore, that forced groundwater convection is easily capable of causing major gradient anomalies along the rift, and is the most geologically and physically realistic primary source for the anomalies within the rift.

MODIFICATIONS OF REGIONAL FORCED CONVECTION

Although the general features of the forced

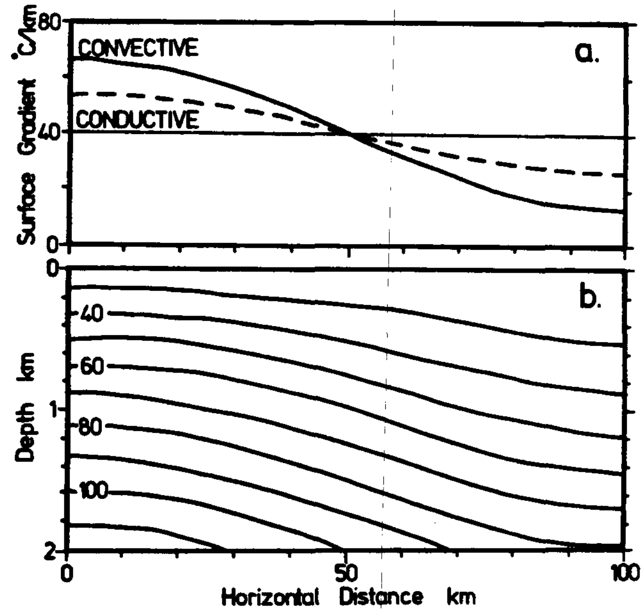

Figure 1. a) Surface geothermal gradient along, and b) isotherms in °C (contour interval 10°C) in a 100km long basin, based on the forced convection model with a hydraulic conductivity of .0002cm/s (200 millidarcies). The dashed curve in a) shows the predicted gradient for a hydraulic conductivity of .0001cm/s (100 millidarcies).

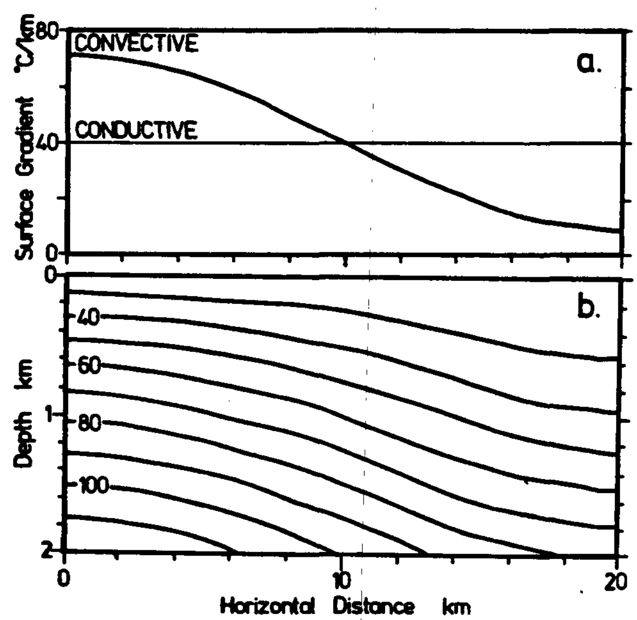

Figure 2. a) Surface geothermal gradient along, and b) isotherms in °C (contour interval 10°C) in a 20km long basin, based on the forced convection model with a hydraulic conductivity of .00005cm/s (50 millidarcies).

convection models presented above are believed to be applicable to the Rio Grande rift basins, the features of local anomalies will be strongly controlled by local permeability and hydraulic gradient conditions. In particular, a study of the basins' hydrologies (see below) indicates that discharge from one basin to the next generally occurs through relatively shallow and narrow groundwater constrictions. These constrictions focus the flow upward and laterally, and increase the surface gradients above the levels predicted by the two-dimensional flow models. In addition the discharge areas are commonly associated with faulting which increases permeability, especially in the deeper semipermeable rocks. Additional permeability may be created by solution by the flowing groundwater, a mechanism that would be particularly effective where pre-basin fill limestones are block faulted upward between basins. Without detailed structural and permeability information it is impossible to test rigorously these hypotheses. They indicate, however, that the numerous diverse geothermal anomalies along the rift could all be caused by the same basic forced groundwater convection mechanism.

The only published detailed subsurface temperature information in the rift is from the Las Alturas anomaly near Las Cruces, New Mexico (Morgan et al., 1979), and these data indicate a temperature inversion in part of the anomaly (Figure 3). This inversion is not predicted by the forced convection model. An inversion is predicted by free convection models when the Rayleigh number (or gradient) is greater than two and a half times the critical Rayleigh number (or gradient) (Donaldson, 1962), as shown in Figure 4. We believe, therefore, that free convection occurs in the Las Alturas system, driven by regional forced convection. This effect is predicted to occur where forced convection increases the gradient to values in excess of 150 to 200°C/km and where the permeability is low (cf. Kilty et al., 1979). Many of the rift anomalies may reflect mixed convection systems.

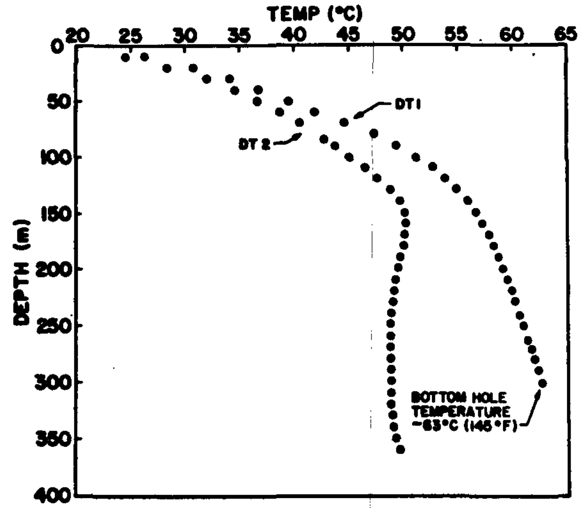

Figure 3. Temperature-depth data from two temperature test wells in the Las Alturas geothermal anomaly showing a temperature inversion in the DT2 well. (From Morgan et al., 1979).

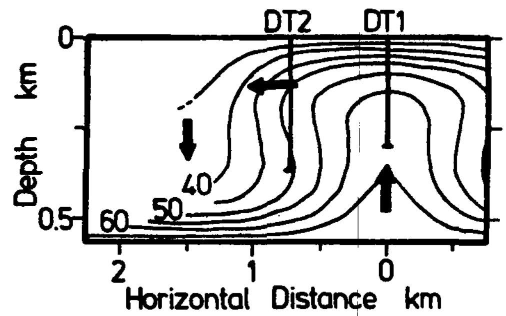

Figure 4. Isotherms in $^{\circ}$C (contour interval 5°C) in the Las Alturas geothermal anomaly, scaled from the free convection model of Donaldson (1962), assuming that forced convection increases the mean temperature gradient in the anomaly to a level equivalent to a Rayleigh number of 96. Possible positions of test wells DT1 and DT2 within the system are shown. An upper temperature of 30°C accounts for overlying low conductivity dry sediments. Cell aspect ratio is approximately 3 to 1 (width to depth).

PRACTICAL TEST OF FORCED CONVECTION HYPOTHESIS

The acid test for any model is whether or not it explains observed data. To test the convection hypothesis we have compiled hydrology data from the rift basins to locate constrictions in groundwater flow where water discharges from one basin to the next. The convection model predicts that these are the areas where geothermal anomalies should occur. The groundwater constrictions are shown in Figure 5, and show a very strong correlation with geothermal anomalies and hot springs, as listed in Table 1. The exact location of each geothermal system depends on the local plumbing of the system, but the strong general correlation between groundwater flow constrictions and geothermal anomalies indicates that the primary source of the anomalies is forced convection.

DISCUSSIONS AND CONCLUSIONS

There can be little doubt that many, if not all of the geothermal anomalies along the Rio Grande rift, excluding the Jemez Mountains, are primarily the result of forced convection by inter and intrabasin groundwater flow. In some anomalies forced convection creates high enough gradients for free convection to occur also, creating mixed convection systems. The temperatures within these systems are limited by the maximum depth of water circulation, and the regional geothermal gradient. These conditions are unlikely to produce high enough temperatures at shallow enough depths for electricity generation under current economic conditions. The geothermal systems have great potential for direct heat applications, however.

Although the forced convection models have been developed for Rio Grande rift basins, where

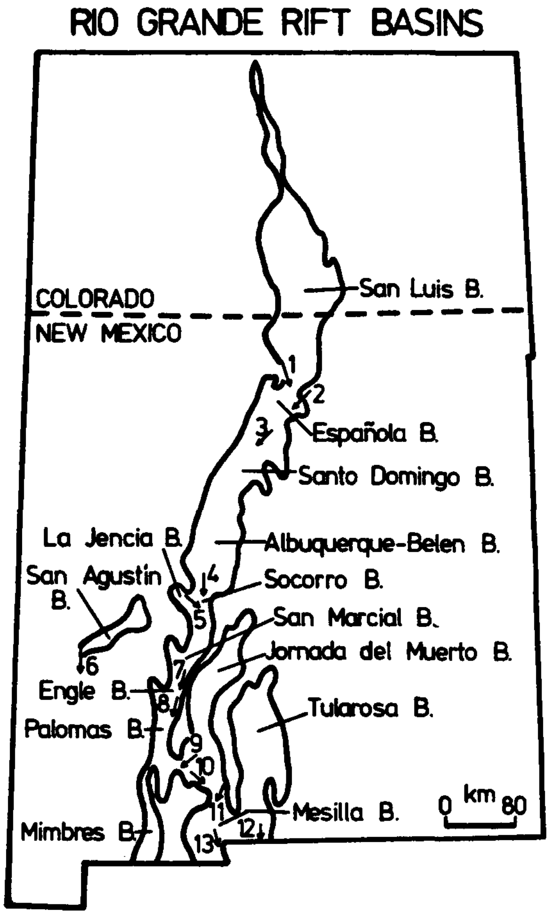

Figure 5. Rio Grande rift basins with interbasin groundwater flow constrictions shown by arrows, which also indicate direction of flow. Numbers refer to constriction numbers in Table 1.

similar permeability and hydraulic gradient conditions exist, either in or out of basins, in other areas, the models should be generally applicable. In particular, the models may explain many of geothermal anomalies throughout the Basin and Range province of the western U.S. A preliminary test on the geothermal anomalies in the Basin and Range of southwestern New Mexico indicates that the models are applicable to this area. We conclude, therefore, that many, if not the majority, of geothermal anomalies in the western U.S. are caused by forced groundwater convection.

ACKNOWLEDGEMENTS

Our understanding of the limitations of the analytic forced convection model was increased by discussions with David D. Blackwell. Part of this study was carried out at the Lunar and Planetary Institute, operated by the Universities Space Research Association under contract number NASW - 3389 from the National Aeronautics and Space Administration. The study was also partially funded by subcontract L - X60 - 2133K - 1 from Los Alamos Scientific Laboratory to Purdue University.

Constriction Number	Source Basin	Geothermal Feature and Remarks
1	San Luis	Ojo Caliente, Hot Baths.
2	San Luis	Rio Grande Gorge hot springs, Embudo constriction.
3	Espanola	None reported, White Rock Canyon, La Bajada constriction.
4	Albq.-Belen	Socorro area, KGRA.
5	La Jencia	Socorro Peak, KGRA.
6	San Agustin	Possibly feeds hot springs on south and southwest of Gila area, with additional forced convection before discharge.
7	San Marcial	None reported, Pankey channel, submerged by Elephant Butte reservoir.
8	Engle	Truth or Consequences, Hot baths.
9	Jornada	San Diego Mt., KGRA.
10	Palomas	Radium Springs, KGRA.
11	Jornada	Las Alturas, Direct heat use on New Mexico State U. campus.
12	Tularosa	Hueco Tanks, West Texas geothermal project.
13	Mesilla	Anthony area, Anomalous low electrical conductivity zones currently under investigation.

Table 1. Geothermal features and their source basins along the Rio Grande rift. Constriction numbers refer to Figure 5.

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GEOTHERMAL EXPLORATION METHODS USED
IN THE CAPITAL DISTRICT OF NEW YORK

Margaret R. Sneeringer
Dunn Geoscience Corporation

1.0 INTRODUCTION

1.1 Saline and carbonated saline waters occur in the Capital District of New York, most notably in the Saratoga Springs vicinity. The springs at Saratoga are not thermal, their temperatures being generally the same or slightly cooler than normal ground water; but they do have an unusual chemistry with up to 20,000 ppm total solids and large volumes of carbon dioxide. There are fresh water thermal springs occurring in nearby Lebanon Springs, New York, and in Williamstown, Massachusetts. The presence of these thermal springs and the unusual chemistry of the Saratoga waters have led to exploration of a possible geothermal system in the Capital District area. The program is funded by the New York State Energy Research and Development Authority (ERDA) and the U.S. Department of Energy.

1.2 The presence of a convective geothermal system in the Capital District area is suggested indirectly from geochemical data and more directly from thermal gradient measurements. Exploration techniques have included: detailed water chemistry; a silica-water temperature survey; free and dissolved gas analysis; a passive seismic survey; a small scale gravity survey coupled with recalculation of existing data; and, a thermal gradient measurement program. The thermal gradient measurement program has produced the most direct evidence of subsurface heat. The temperature data are supported by the geochemical and gravity data, showing a coincidence of the apparent thermal area, the locations of saline and carbonated wells, and locations of possibly related gravity structures. A brief account of the methods used, and the results and value of the techniques will herein be presented.

One of the greatest difficulties in attempting to evaluate this area is that it is culturally highly developed. Large areas are unavailable for geochemical sampling or gradient measurements because city water supply systems have replaced the need for private wells, and many of the existing wells have long since been buried, filled in, or simply forgotten. Geophysical work also suffers from the high cultural activity of the area. In particular, it was extremely difficult to locate the microseismic network stations where they would not be affected by pumping stations, heavy truck traffic, quarrying operations, and microwave or other frequency transmission towers. The cultural factor will also affect upcoming work, as the area of primary interest for drilling is highly populated.

In addition to the cultural overprint, a geological barrier to water sampling and gradient measurement also exists. A large part of the area has a thick glacial cover, and abundant water supplies are available near the surface, precluding the need for deep wells penetrating rock.

In spite of these hurdles, however, significant headway has been made in evaluating the geothermal potential in the area.

2.0 GEOCHEMICAL METHODS

Three geochemical methods have been applied to this area in order to characterize the water chemistry, determine the extent of the area over which the potential geothermal fluids are issuing, and to try and identify geothermal components (if any) being added to the system. These methods have provided indirect evidence of a geothermal system.

2.1 Water Chemistry

Water samples were collected from abandoned and operating wells producing saline and carbonated waters in a nine-county area surrounding the Capital District. The samples were analyzed for a complex suite of elements, including: Na; K; Ca; Mg; Cl; SO_4; Br; nitrogen species; total Fe; SiO_2; Al; F; I; phosphate; Ba; Sr; Li; Hg; Zn; TOC; and, total solids. Water temperature, pH, and titrated alkalinity were determined at the time of sampling.

Results of these analyses were difficult to interpret because of a complex mixing problem of the saline and carbon dioxide-rich components with the surface waters, and an apparent ionic filtration of the waters away from assumed zones of issuance. The most concentrated solutions, however, and those producing the more unusual trace element compositions did tend to be associated with major fault zones, particularly in the Saratoga Springs vicinity. Perhaps the most notable trace indicator present in high concentrations was silica, occurring in concentrations up to 70 ppm in Saratoga Springs itself. Other trace elements of interest that occur in these waters include boron (up to 4 ppm), lithium (up to 25 ppm), fluorine (up to 4.6 ppm), and strontium (up to 100 ppm). Oxygen and carbon isotopes were previously measured on carbon dioxide gas exsolving from the Saratoga waters, and the pertinent ratios indicated a thermal origin for the gas. Another series of isotope analyses are currently underway.

2.2 Silica-Water Temperature Analysis

Consequent to finding high silica contents in some waters in Saratoga Springs, an effort was made to determine the extent of the silica anomaly. Water samples were collected from wells penetrating bedrock in a broad area around Saratoga Springs and analyzed for silica content and water temperature. A correlation of higher silica contents and warmer temperatures was noted, and a general trend of slightly higher silica contents around Saratoga Springs and along the fault zones, decreasing away from these areas. No values of the magnitude noted in Saratoga Springs were found elsewhere, but the pattern observed appears to be useful in exploration and the data base will be increased and further analyzed. Silica geothermometry applied to the highest measured SiO_2 concentration (70 ppm) yields a last equilibration temperature of approximately 89°C for a chalcedony silica species, and of 118°C for a quartz silica species.

2.3 Gas Analyses

Water samples for dissolved gas analyses and free gas samples were collected at the same time water chemistry samples were taken, and analyzed for carbon dioxide, oxygen and argon, nitrogen, helium, and light hydrocarbons. Some difficulty was experienced in avoiding air contamination of the samples, either during sampling, or through the plumbing system, limiting the amount of information obtained from the program.

Carbon dioxide has long been known to exsolve from the waters at Saratoga Springs, but has not been recognized elsewhere in the area. This work indicated that the CO_2 is exsolving from a much larger area, and that it is associated with the major faults in the area. Methane was also determined to be a major component of the gases, with apparently increasing concentration to the south of the area of interest. Helium was present in some gas samples in quantities exceeding 4000 ppm, but it does not have a clear association with either the CO_2 or methane.

3.0 GEOPHYSICAL TECHNIQUES

Three geophysical techniques have been employed in the Capital District area, including a microseismic monitoring network, a small-scale gravity survey, and a temperature gradient measurement program.

3.1 Seismic Monitoring

Members of the New York State Geological Survey have set up a five-station seismic network in an effort to determine whether there is unusual fault activity or seismic noise which might be geothermally related. The network operated for a period of four months during which time eight minor earthquakes were recorded. Two of these were just west of Lebanon Springs, and one on the northern extension of the McGregor Fault. All reported earthquakes were compatible with the historical record for the area.

The vicinity around Saratoga Springs is an area of non-recorded tremors. Tremors were reported during the operational period of the seismic network, but the instruments did not record the events.

An alteration of the station distribution is planned to more effectively cover the area of interest.

3.2 Gravity Survey

The gravity data base for the Saratoga Springs to Schenectady area was expanded, and existing data was recalculated and contoured to show more clearly the configurations of gravity anomalies in the area. Negative bouguer gravity anomalies were emphasized in the areas west and south of Saratoga Springs and also just to the south of Schenectady. These gravity features cannot be directly related to the thermal system we are dealing with at this time, but there is a general correlation of high thermal gradients and carbonated waters with them. Further work may be merited in this area.

3.3 Temperature Gradient Measurements

Approximately 80 thermal gradients have been measured in abandoned water wells ranging in depth from 80 to 605 meters using a thermistor probe and a Yellow Springs Instruments thermometer. This has been the most successful indicator of a possible geothermal system to date, with the highest reproducible gradient thus far measured of 44.3°C/km. The highest gradients observed are in the area between Saratoga Springs and Schenectady and appear to be strongly related to the Saratoga and McGregor Faults. The regional background appears to be on the order of 8°C/km to 10°C/km based on gradients measured around the area. Bearing this in mind, we have a system apparently producing gradients from two to four times background and even up to almost two times the worldwide average gradient. It is extremely important to determine if these gradients can be extended to depth, and future efforts will be in that direction.

4.0 CONCLUSIONS AND PROJECTIONS

4.1 Conclusions

Direct evidence of anomalous geothermal heat has been demonstrated through the measurement of temperature gradients in abandoned water wells throughout the Capital District. New and previous geochemical data support these results and indicate that the Saratoga and McGregor Faults are acting as major conduits for mineralized waters and thermally derived carbon dioxide. Issuant points for these waters and higher geothermal gradients correspond with gravity anomalies in the area which are also suggestive of conduits from depth.

4.2 Projections

Further exploration will involve an expanded silica sampling program, continued seismic monitoring of the area, continued thermal gradient measurements in abandoned wells, and a drilling program designed to confirm some of the higher gradients and to determine the value of the geothermal resource in the Capital District.

AN EVALUATION OF THE HYDROTHERMAL RESOURCES OF NORTH DAKOTA

Kenneth L. Harris, Francis L. Howell, Laramie M. Winczewski,
Brad L. Wartman, Howard R. Umphrey, and Sidney B. Anderson

For the past two years, the North Dakota Geological Survey has been working under a cooperative agreement with the Department of Energy, Division of Geothermal Energy (DOE-FCØ7-79ID12Ø3Ø) to evaluate North Dakota's hydrothermal resources.

Our first year of work utilized the North Dakota Geological Survey's (NDGS) oil and gas well data to evaluate deep, principally Paleozoic, aquifers. Information from the NDGS oil and gas well files was encoded and compiled into a computer library system (WELLFILE). Data stored in WELLFILE was used to summarize the depth, thickness, expected temperature, and water quality of potential hydrothermal aquifers.

A geothermal gradient map (fig. 1) of North Dakota was produced by interrogating WELLFILE for the recorded bottom-hole-temperature and total depth of wells drilled in the state. Gradients were calculated (fig. 2) and the average gradient for each township was contoured. The geothermal gradient map is displayed in degrees Celsius per kilometre so that it may be easily compared with geothermal gradient maps produced by other workers in the region.

Using the information contained in WELLFILE, we can summarize the hydrothermal prospects of the Mississippian Madison Formation. Four important factors in evaluating a potential hydrothermal aquifer are depth, thickness, water temperature, and water quality. WELLFILE was interrogated for the depth of the top and base of the Mississippian Madison Formation in order to construct maps showing the depth to and the thickness of the Madison (fig. 3 and 4). An expected water temperature map was constructed by mapping the bottom-hole-temperatures of all wells, on record, that bottomed in the Madison (fig. 5). The Madison Formation water quality map (fig. 6) was produced by mapping the total dissolved solids reported in analyses of water recovered from Madison drill stem tests. The Madison Formation is a medium- to low-temperature hydrothermal reservoir. The water contained in the reservoir is typically very high in NaCl, with concentrations of total dissolved solids as high as 300,000 mg/l. Although the Madison contains water in a useful temperature range, the generally poor quality of the water and great depth will probably prevent its development as a significant hydrothermal aquifer in North Dakota.

Since oil and gas exploration is confined mainly to the western two-thirds of the state, any study based soley on these data does not provide information on the entire state. Consequently our second year of work utilized "shallow" well data to fill in those gaps left by our previous study and to evaluate the hydrothermal potential of shallower aquifer systems in areas of interest indicated by our study of the oil and gas data. Our efforts have been directed at three main tasks: geothermal gradient and heat flow studies, stratigraphic studies, and water quality studies.

Geothermal gradient and heat flow studies have involved the temperature logging of available groundwater observation wells (fig. 7); and locating, obtaining access to, and casing "holes-of-opportunity" to be used as heat flow determination sites (fig. 8). Information obtained from the temperature logs of groundwater observation wells will be used to construct a geothermal gradient map, and expected temperature maps for the upper 100 metres (328 feet) of sediment in the state. Although this work is not yet completed, figure 9 shows a preliminary geothermal gradient map based on groundwater observation well temperature logs run this past year.

Stratigraphic studies and water quality studies involve the continued development and expansion of our WELLFILE system. Information on "shallow" observation wells, water wells, and other test holes, drilled throughout the state, has been collected for the past twenty years. This county by county groundwater resource study has been conducted jointly by the North Dakota Geological Survey, North Dakota State Water Commission (NDSWC), and United States Geological Survey Water Resources Division (USGSWRD). We have obtained digitized magnetic tape summaries of the data collected through these studies from the USGSWRD. The information contained on these tapes has been reduced and assembled in a computer library system (WATERCAT). Summaries of the depth, thickness, expected water temperature, and water quality can now be constructed for "shallow" aquifer systems in North Dakota.

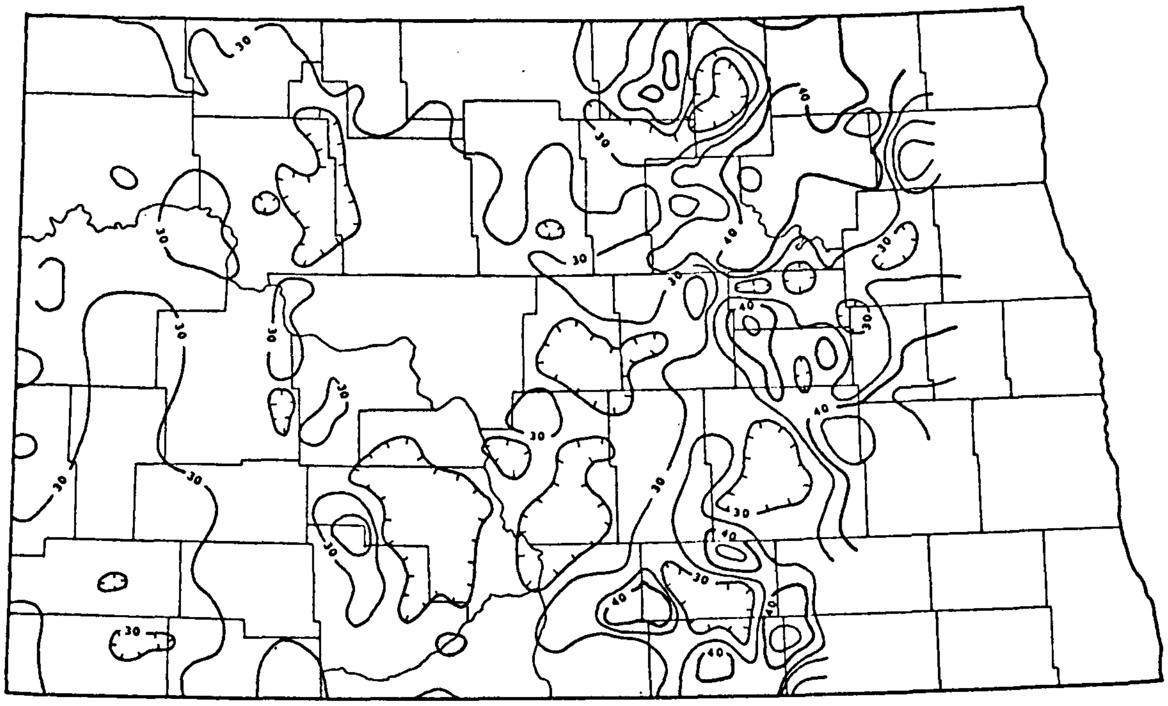


FIGURE 1 - GEOTHERMAL GRADIENT. THE CONTOUR INTERVAL IS 5°C/Km.

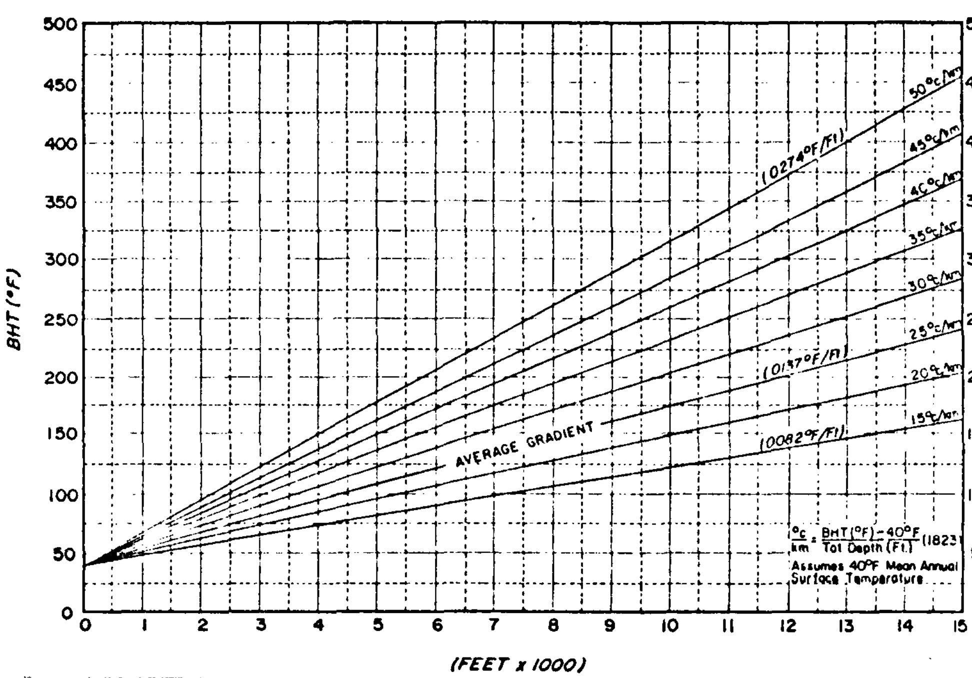


FIGURE 2 - METRIC GEOTHERMAL GRADIENT vs BOTTOM-HOLE-TEMPERATURE (°F) AND TOTAL DEPTH (feet).

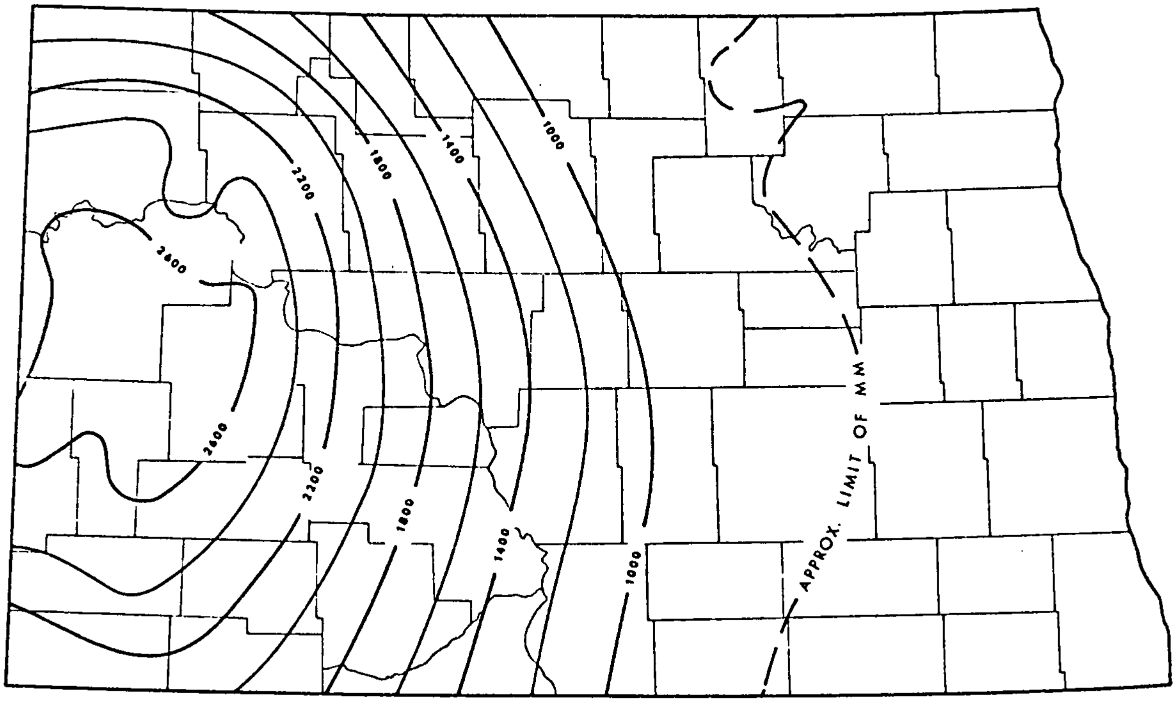


FIGURE 3 - DEPTH TO MADISON, CONTOUR INTERVAL IS 200 METERS.

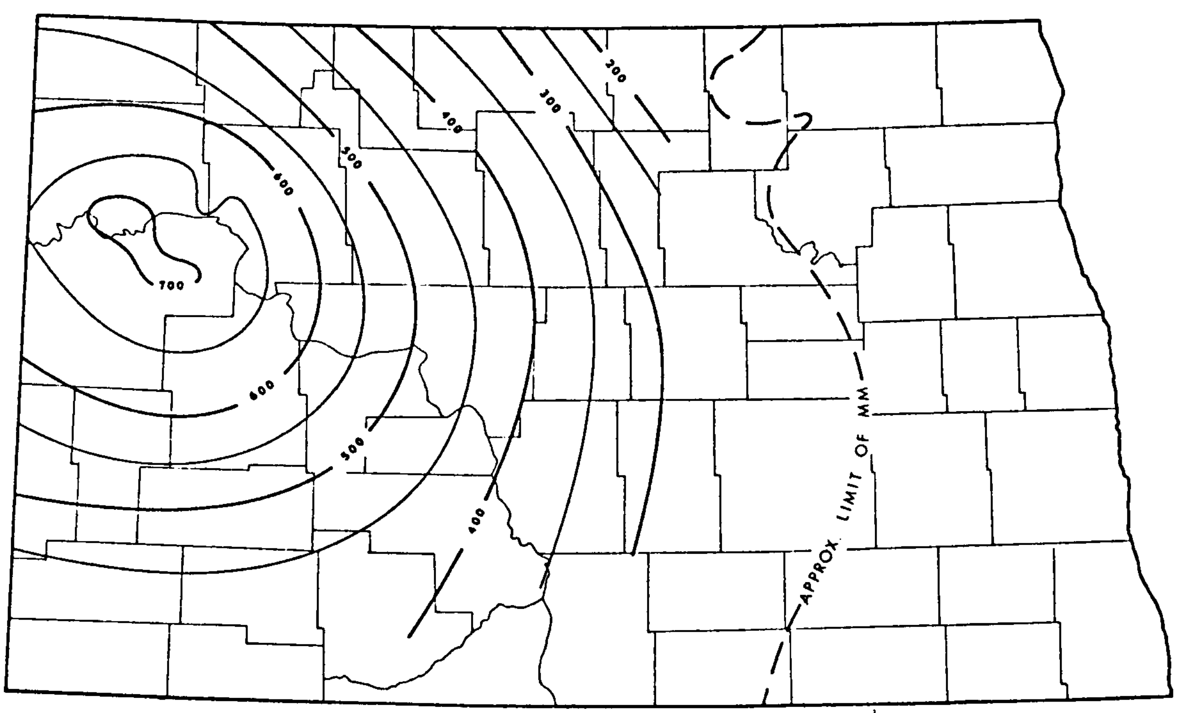


FIGURE 4 - ISOPACH MAP OF THE MADISON, CONTOUR INVERVAL IS 100 METERS.

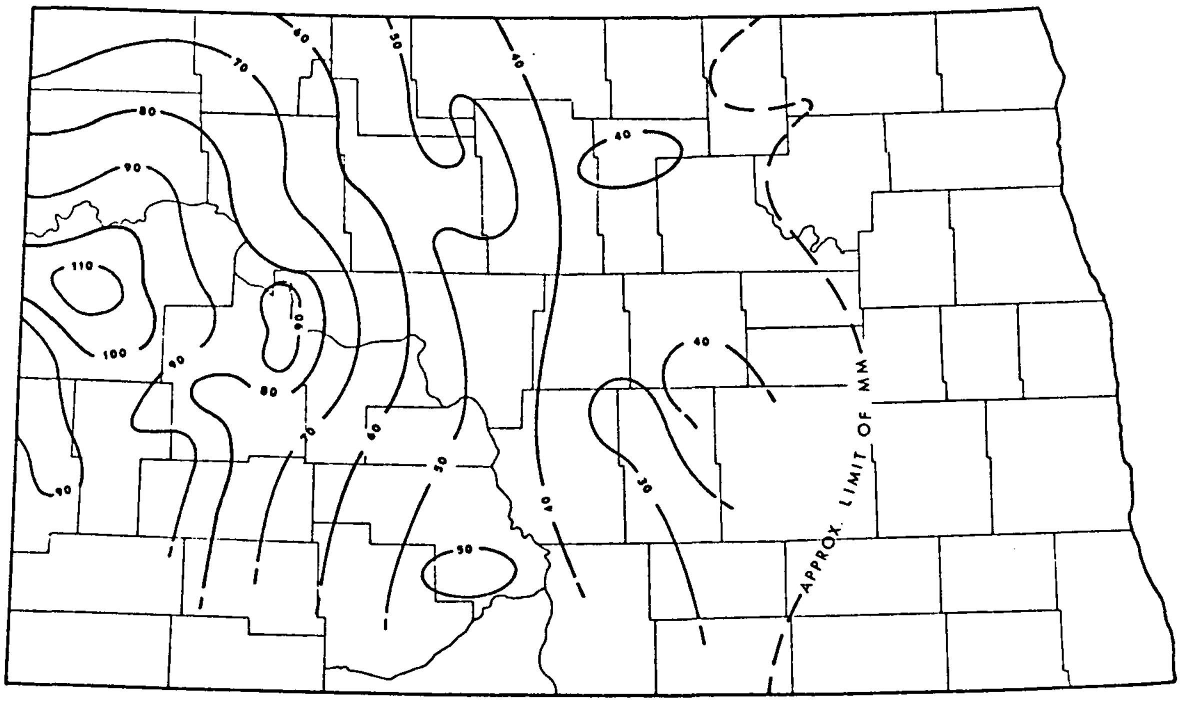


FIGURE 5 - ISOTHERM MAP OF THE EXPECTED TEMPERATURE OF MADISON WATER, CONTOUR INTERVAL IS 10°C.

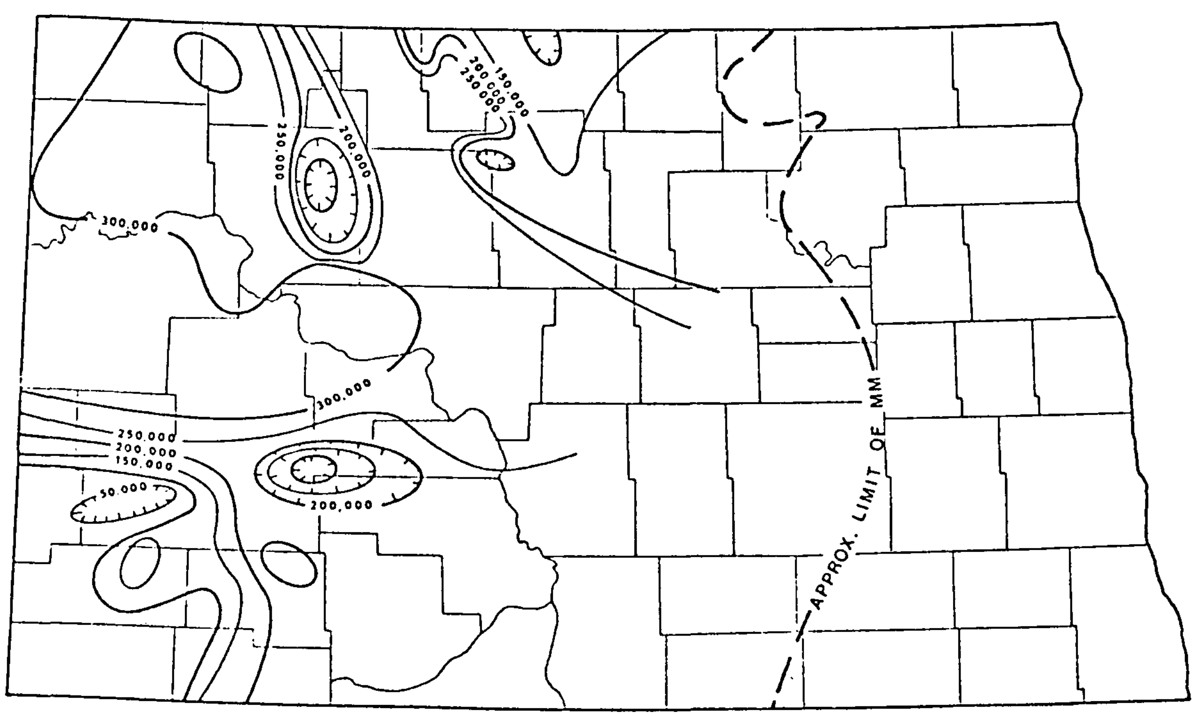
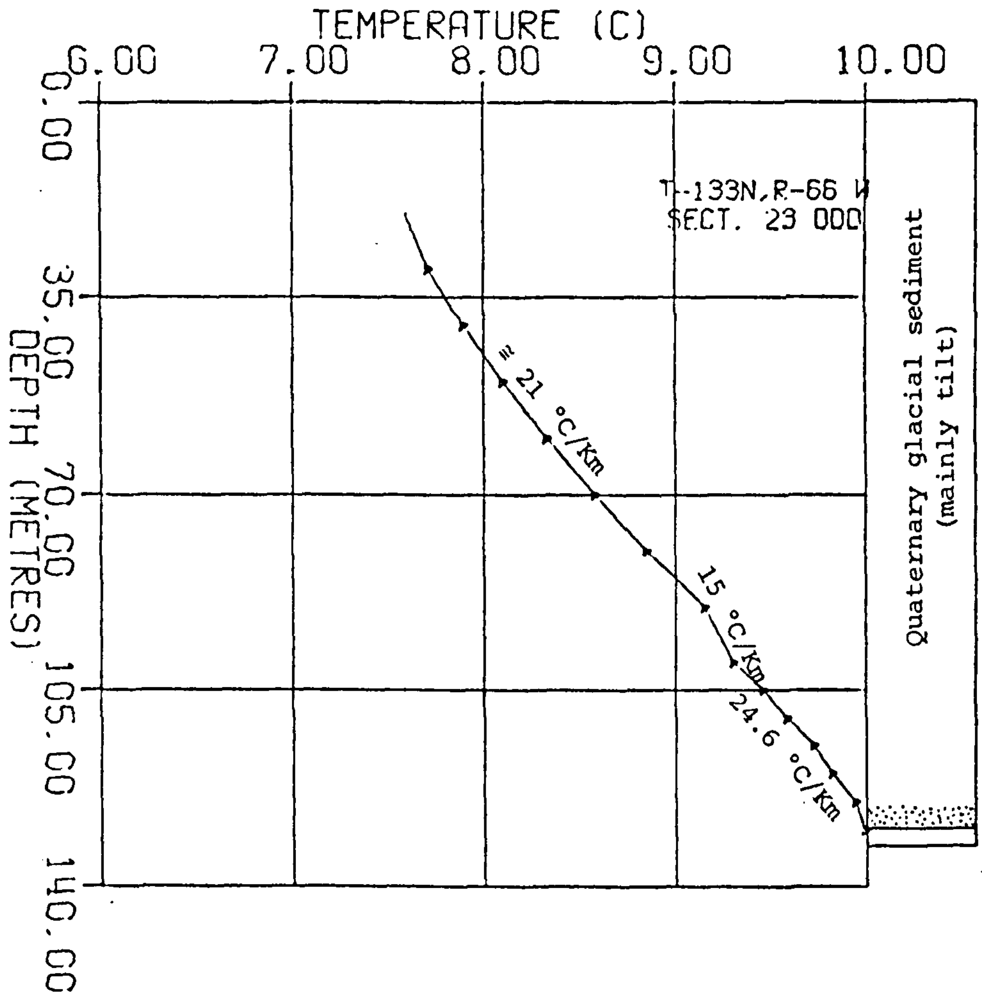
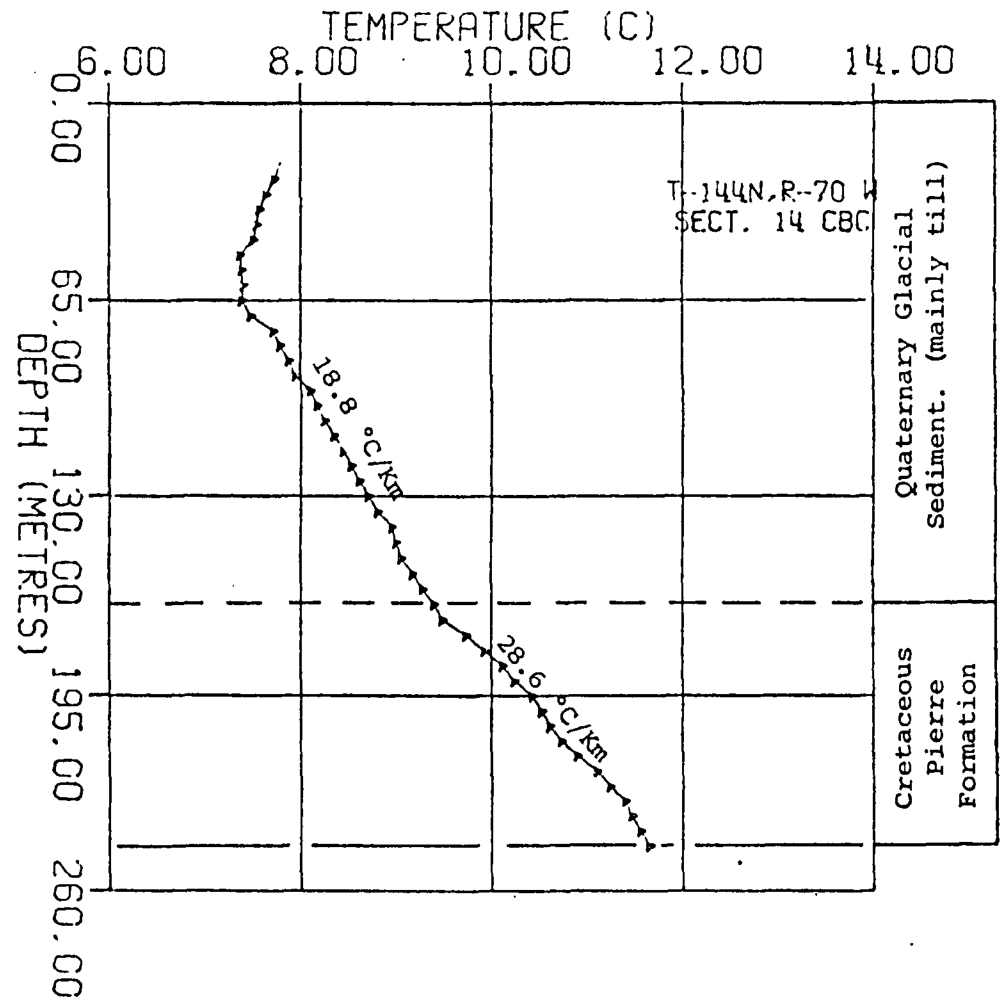


FIGURE 6 - CONCENTRATION OF TOTAL DISSOLVED SOLIDS (TDS) IN THE MISSISSIPPIAN MADISON FORMATION. CONTOUR INTERVAL IS 50,000 Mg/l.



Location - LaMoure Co., T-133N, R-66W, Sect. 23, DDD
Date drilled - 10/15/74
Total depth - 134 m (440 ft.)
Surface elevation - 591 m (1940 ft.)
Temp. log run - 08/30/80

FIGURE 7 - TEMPERATURE LOG RUN ON A GROUNDWATER OBSERVATION WELL. GENERAL WELL DESCRIPTION IS GIVEN AT RIGHT SIDE OF THE TEMPERATURE-DEPTH PLOT.



Location - Kidder Co., T-144N, R-70W, Sec. 14, CBC

Date drilled - 07/24/80
Total depth - 241 m (793 ft.)
Surface elevation - 570 m (1870 ft.)
Temp. log run - 08/30/80

FIGURE 8 - TEMPERATURE LOG RUN ON A "HOLE-OF-CONVENIENCE" CASED AS A HEAT FLOW DETERMINATION SITE. GENERAL WELL DESCRIPTION IS GIVEN AT THE RIGHT SIDE OF THE TEMPERATURE DEPTH PLOT.

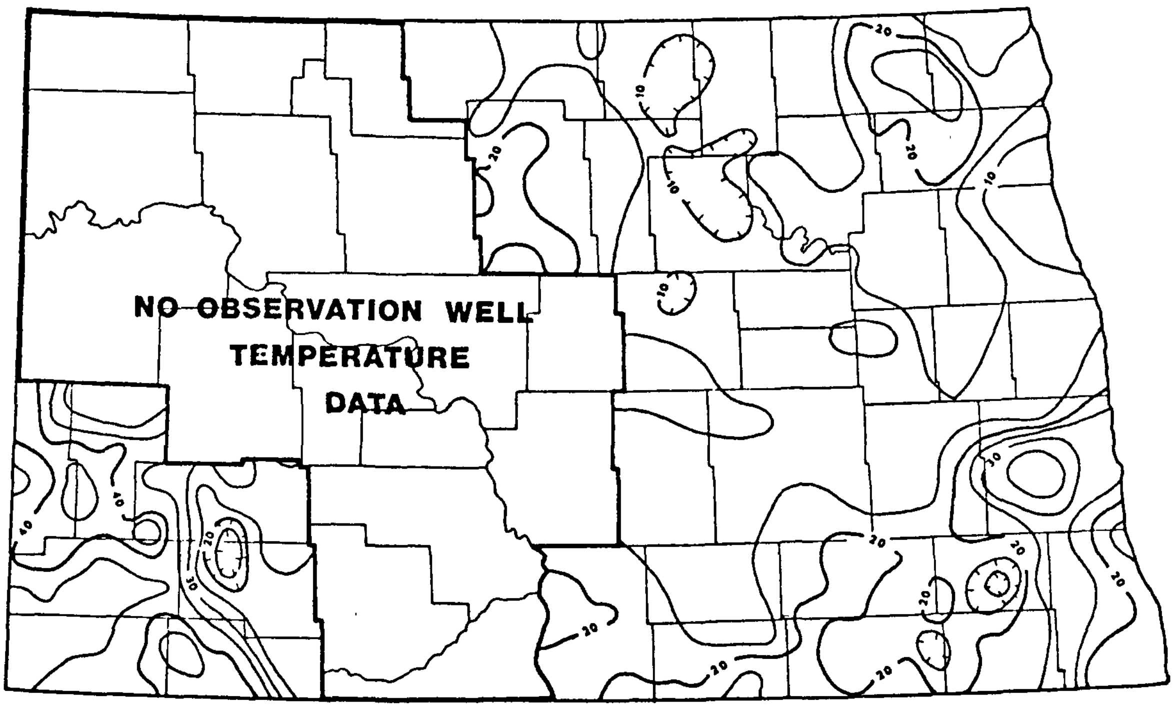


FIGURE 9 - PRELIMINARY "SHALLOW" GEOTHERMAL GRADIENT MAP OF NORTH DAKOTA. BASED ON MEASURED TEMPERATURES BETWEEN 30 AND 100 METRES IN 241 GROUNDWATER OBSERVATION WELLS. CONTOUR INTERVAL IS 5°C/Km.

195-198
297-300

Geothermal Exploration Philosophy for Mount St. Helens (and other Cascade Volcanoes?)

by

J. Eric Schuster

Washington State Division of Geology and Earth Resources
Olympia, WA 98504

April, 1981

Introduction

The most promising geothermal targets in the Cascade Range are the young stratovolcanoes. The five stratovolcanoes in Washington are reported to support fumaroles near their summits, and there are hot springs on or near the flanks of three of the volcanoes. Geologic studies have demonstrated the extreme youth of volcanic deposits produced by each of these volcanic systems, and there are records of historic eruptions on three of the peaks.

However, the most impressive demonstration of the energy potential of these volcanic systems is the 1980 series of eruptions of Mount St. Helens. The total energy released by this series of eruptions has been estimated at 2.0×10^{25} ergs, 90 percent of which was released in the cataclysmic eruption of May 18. The equivalent electrical energy is about 63,000 megawatt years, or roughly 100 years of power generation at The Geysers.

Tremendous energy sources probably exist beneath Cascade stratovolcanoes, even during periods between eruptions, but exploration has been hampered and slowed by a number of factors:

1. Several of the stratovolcanoes are included within Wilderness Areas or National Parks. This precludes or severely limits possibilities for exploration and development of geothermal resources on and near these volcanoes.
2. Most of the stratovolcanoes are highly regarded for their scenic and recreational values. These values conflict or seem to conflict with the preception of geothermal development as an industrial activity.
3. Geothermal exploration and development on a stratovolcano presents a number of difficult logistical problems.
4. Most Cascade stratovolcanoes lack robust surface manifestations of geothermal systems. Some investigators believe this is due to cooling and dilution by a "cold water blanket" which results from the heavy precipitation received by many parts of the Cascade Range. As a result, specific targets are lacking.

5. Secondary minerals resulting from hydrothermal alteration apparently seal many of the older fault and fracture zones. Even when moderately deep drill holes penetrate these zones and temperatures are encouraging, there may be no significant water flow from the hole.
6. There has been very little geothermal leasing of federal lands in the Cascades.

Progress toward exploration and development of geothermal resources near Cascade stratovolcanoes can be made if some of these problems can be avoided. Exploration for geothermal resources near stratovolcanoes must obviously take place in close enough proximity to the volcano to take advantage of the heat source that the volcano represents. At the same time, in order to proceed and be cost-effective, exploration must take place outside of National Parks and Wilderness Areas, outside of the areas where scenic and recreational values are so high, and away from the logistical problems presented by the peaks themselves. Furthermore, exploration probably should not depend heavily on surface manifestations as a means for locating targets. Active faults should be sought where hydrothermal alteration has not had time to precipitate secondary minerals to seal the system.

Recent experience at Mount St. Helens suggests that it may be possible to develop an exploration philosophy which incorporates many of these attributes. First, it is necessary to review what was known about geothermal energy in the Mount St. Helens area prior to 1980, and what has been learned as a result of the 1980 eruptions.

Mount St. Helens Prior to 1980

Prior to 1980 the geologic record of Mount St. Helens indicated a history of eruptions stretching back nearly 36,000 years. Historic records confirmed eruptions between 1832 and 1857. Mount St. Helens was believed to be the Cascade volcano most likely to erupt explosively.

A seismic network had been operated in western Washington by the University of Washington for nearly a decade. Plots of about 20 earthquake epicenters recorded in the Mount St. Helens area prior to 1980 suggested the presence of a structure (fault or fault zone?) trending to the north-northwest from the mountain. Little was known about the nature of this structure or its relationship to volcanism at Mount St. Helens.

One mineral spring was known to exist along the trend of this structure; one shallow heat-flow hole drilled along this trend at a distance of 8 km from the volcano showed no anomalous temperatures; small fumaroles near the summit fell on or near the NNW trending structure. Young volcanic features were known to occur to the SSE of Mount St. Helens, including Marble Mountain, Soda Peaks, West Crater, and Trout Creek Hill volcano, but only Marble Mountain is located 15 km or less from the stratovolcano.

Mount St. Helens, 1980

During the 1980 eruptions, the relative seismic quiescence changed dramatically. Seismic evidence for a major active fault trending NNW-SSE through Mount St. Helens, and a lesser fault trending NE-SW and intersecting the major trend beneath the volcano, has become very convincing. Thousands of earthquakes, up to a Richter magnitude of 5.5, have occurred. A significant number of these earthquakes have occurred along the NNW-SSE trending fault at distances up to 30 km from the volcano and at depths as shallow as 5 km or less.

The 1980 volcanism has not generated new hot springs or other surface manifestations except for those related to the central vent and the hot pyroclastic deposits filling the valley to the north. Neither have changes been observed in the cold springs surrounding the base of the mountain, nor in the temperatures of two remaining heat-flow holes near the mountain (a third hole, mentioned above, was destroyed by the May 18 Toutle River debris flow). The "cold water blanket" over the Mount St. Helens area appears, then, to have thus far remained unaffected by the eruptions, except for the area near the central vent.

Exploration Philosophy

The 1980 seismicity and the interpretations which are beginning to grow out of the recent studies of Mount St. Helens may be the key to developing a philosophy for geothermal exploration around Mount St. Helens, and perhaps other Cascade stratovolcanoes as well.

The seismic activity can be interpreted as follows:

1. An active fault zone which is intimately related to volcanic activity extends through Mount St. Helens.
2. The fact that the fault zone is active means that permeabilities along it are good, allowing for fluid movement.

3. Hypocenter depths are, at least in part, within reach of a deep drill hole.
4. Fluids migrating along the fault zone may be hot water or even magma.
5. The fault zone extends beyond Mount St. Helens far enough so that exploration may be able to avoid the serious logistical and environmental problems associated with the mountain.

In the case of Mount St. Helens, geothermal exploration should first concentrate on the development of a better seismic velocity model. This will allow more precise calculation of earthquake hypocenter locations and, therefore, more precise definition of the width, trend, dip and mechanics of the NNW-SSE trending fault zone. This will require shot-hole refraction seismic work. An additional objective of shot-hole work would be to test for the presence, shape, size, and depth of a magma chamber beneath Mount St. Helens.

Seismic work might be supplemented by additional geologic, geochemical, or geophysical work focused toward determining the nature of the fault zone and whether heated fluids are, indeed, present. Further study will be required to determine which additional studies would be most appropriate.

Once the fault zone has been defined as well as practical by surface methods, the zone should be explored by deep drilling.

Depending heavily on a single method of investigation (in this case seismic studies) to locate geothermal targets for deep drilling is certainly not standard practice. However, because of the 1980-81 eruptions Mount St. Helens cannot be seriously questioned as a source of heat, the NNW-SSE trending fault zone is almost certainly real and is quite probably a zone of permeability, and other exploration methods around Cascade stratovolcanoes have either been difficult to interpret or have failed to penetrate the "cold water blanket". Deep drilling based on the best available seismic data and interpretations is logical at Mount St. Helens.

At other Cascade stratovolcanoes existing seismic data should be analyzed for structural trends and these trends investigated. If seismic data are poor or lacking, long-term seismic monitoring networks should be set up.

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CONDUCTIVE THERMAL MODELING OF WYOMING GEOTHERMAL SYSTEMS

by

Henry P. Heasler
Department of Geology and Geophysics
University of Wyoming
Laramie, Wyoming

Introduction

The purpose of this paper is to present a summary of techniques used by the Wyoming Geothermal Resource Assessment Group in defining low-temperature hydrothermal resource areas. Emphasis will be placed on thermal modeling techniques appropriate to Wyoming's geologic setting. Thermal parameters discussed include oil-well bottom hole temperatures, heat flow, thermal conductivity, and measured temperature-depth profiles. Examples of the use of these techniques will be from the regional study of the Bighorn Basin and two site specific studies within the Basin. Funding for this work has been primarily from the U.S. Department of Energy Cooperative Agreement DE-FC07-791D12026.

General Geologic Setting

Wyoming is in the Rocky Mountain and Great Plains physiographic provinces (Fenneman, 1946). This region, often referred to as the Rocky Mountain Foreland, is primarily situated between geosynclines to the west and the stable craton to the east. Much of Wyoming is essentially a group of large intermontane basins separated by major mountain ranges or arches (Figure 1). For most of the region the present distribution of mountains and basins resulted from the Laramide orogeny beginning in the Late Cretaceous (approximately 70 million years ago) and terminating in the middle to late Eocene (approximately 40 million years ago). The tectonic style of Wyoming is characterized by compression, uplift and thrust faulting (Blackstone, 1971; Houston, 1969).

The major mountain uplifts expose rocks of Precambrian age while the basins contain sediments of Paleozoic, Mesozoic and Cenozoic age. Some of the Paleozoic and Mesozoic sediments are porous and permeable forming aquifers that exist over entire basins. The Laramide deformation of these has resulted in a structural relief of many kilometers in the basins. For example, portions of the Bighorn Basin have a structural relief of over 9 kilometers (Prucha et al., 1965). Also, due to the physical properties and deformation of the sediments, a great amount of hydrocarbon exploration has taken place in Wyoming.

There has been volcanic activity in Wyoming in Yellowstone National Park as recent as the Pliestocene (Love et al., 1972), in the Absaroka Mountains in the middle Eocene (Smedes and Prostka, 1972), in the Rattlesnake Hills in central Wyoming in the middle Eocene (Pekarek, 1974) and in the Black Hills in the middle Eocene (Houston, 1963). Preliminary geothermal studies by out group have not identified any geothermal resources near the Eocene age volcanics.

Thermal Techniques

A major portion of the geothermal assessment of Wyoming has been the compilation and depiction of existing bottom hole temperatures from oil and gas wells. The largest problem with using such data is assessing its reliability. The problem of the thermal equilibration of a well after drilling has been discussed by Lachenbruch and Brewer (1959), and the problem of thermal instability of a large diameter drillhole is adressed by Diment (1967). However, with oil and gas well data many of the correction factors for thermal equilibrium and stability are unknown and consequently the bottom hole temperatures cannot be absolutely corrected for various thermal perturbations.

Our attempt at solving this problem has been to define anomalous points within the oil well bottom hole temperature data set. Many parameters must be considered in the definition of anomalous points. First, the data are analyzed only for areas of similar geology. Parameters considered here are the character of aquifers present, lateral extent and continuity of formations, oil and gas producing units, tectonic style, and actual rock types present. By analyzing the data within a similar geologic area we are attempting to eliminate the variability in the data that would be due to differences in crustal heat flow and thermal properties of vastly different rock units.

After the data is gathered for an area, a series of computer plots are made showing temperature versus depth, thermal gradient versus depth, and temperature versus thermal gradient. The effect of average and maximum mud temperatures on the thermal gradient for the region is then plotted with the thermal gradient versus depth data (see Figure 2). From these plots, data with anomalously high thermal gradients can be identified and located on a map. If these data points cluster in an area, we feel confident of the anomaly and study the anomalous areas in greater detail. Also, from the bottom hole temperature data set maps are compiled of the thermal gradient and the major aquifer temperatures.

To further assess the reliability of the oil well bottom hole temperatures considerable effort is expended measuring temperatures in drill holes. Basically, temperatures are measured with a thermistor probe and wheatstone bridge combination at 5 to 10 meter intervals down the wells. Decker (1973) gives a complete discription of equipment used and assesses the errors involved. Least squares gradient computations for linear segments of the temperature versus depth plots and absolute measured temperatures are then compared with the gradients and temperatures from the oil well bottom hole temperature data set. To date what has been found is that the oil well bottom hole temperatures and gradients for holes generally deeper than 750 meters (2500 feet) have always been less than or equal to the measured temperatures and gradients. Above this depth the effect of warm drilling fluids raising the equilibrium temperature must be considered. However, since most of the oil data is below 750 meters (2500 feet), this method of bottom hole temperature analysis generally results in gradient and temperature values less than equilibrium values.

As previously mentioned, a great amount of effort is expended on thermally measuring drill holes. This serves three main purposes. First it is important in the determination of the reliability of the oil and gas temperature data. Second, the resulting temperature-depth profiles show the change of gradient within differing rock units and with depth. Finally, thermally measured holes when combined with rock thermal conductivity can be used to estimate the heat flow for an area. Both of these last two uses of thermally measured holes are critical to the modeling of hydrothermal systems found in the basins of Wyoming.

Most of the hydrothermal systems in Wyoming's basins function in a similar way. Water enters aquifers in the surrounding mountains, flows down the continuous dip slopes to where it becomes heated in a syncline, and is then forced back to the surface or near surface in anticlines. Critical parameters that need to be defined for potential developers of such systems are the maximum temperature of the system, depth to the hydrothermal reservoir, and extent of the reservoir. Conductive thermal modeling is one approach to answering these questions.

The purpose of a conductive thermal model is to calculate the temperature of aquifers in the synclinal portion of these hydrothermal systems. By modeling the temperatures in the aquifers a judgement can be made as to whether the observed thermal anomaly can be explained by the regional heat flow, thermal conductivity of the rocks, depth of the syncline, and water flow direction within the aquifers. If the model fits the thermal anomaly then the critical parameters of maximum temperature, depth to the reservoir, and extent may be predicted with some certainty.

The conductive thermal modeling of an area begins with an understanding of stratigraphy, structural geology, and hydrology. These are parameters which set limits on the thermal conductivity, thermal gradient, and depth to aquifers. Next, a regional heat flow value is determined using published values and new values calculated as a result of our thermal investigations. Since a necessary part of the heat flow determinations is the measurement of rock thermal conductivities, this conductivity data will already exist for use in the thermal model. To model the temperature at a given depth in the syncline one uses the equation:

$$T_a = T_s + [(Q/K_1)\ dx_1 + (Q/K_2 dx_2) + \ldots]$$

where T_a is the sought after temperature in the aquifer, T_s is the mean surface temperature, Q is the regional heat flow, K_1 and dx_1 are the thermal conductivity and thickness of lithologic unit closest the ground surface, K_2 and dx_2 are the thermal conductivity and thickness of the lithologic unit below unit 1, and so on until the aquifer is reached.

Other thermal parameters may also be usefully modeled. For example, the flow of a hot artesian well or spring may be modeled in an attempt to assess how much the temperature is decreased in flowing from the hydrothermal reservoir to the surface (see Truesdell et al., 1977). This is useful in helping to define the maximum temperature of the hydrothermal reservoir. Another useful parameter to model in the syncline-anticline hydrothermal system is the total conductive heat gain and the heat loss of the system. By using the regional heat flow, and the flow and temperature of hot wells and springs a minimum area can be calculated over which the water must flow to attain the needed heat. This calculation will not prove the conductive syncline-anticline thermal model is correct but can help illustrate inconsistencies in the model if the area needed for heat gain is much larger than that available.

Application and Results

The methods discussed have been applied with success to the Bighorn Basin in northwestern Wyoming. Over 1,900 oil well bottom hole temperature points were used in the analysis of anomalously high thermal gradient areas. Gradient-depth data were plotted along with curves representing a gradient resulting from the effect of isothermal drilling mud (Figure 2). This aided in defining areas of anomalously warm fluids (40-70°C (104-158°F)) at shallow depths (150-750 meters (500-2,500 feet)). Based on the data a thermal gradient contour map was compiled and anomalous areas identified near the towns of Cody, Thermopolis, and Greybull (Figure 3). Gradients in these areas were in excess of 90°C/km (50°F/1000 ft).

The Cody and Thermopolis areas were thermally modeled using the described techniques. Heat flow and thermal conductivities were primarily from Decker et al. (1980) and new values determined during the course of this study. Geologic constraints for the modeling resulted from an analysis of existing geologic literature, limited field mapping and an analysis of existing hydrologic data. An important contribution to the thermal model was the actual measurement of temperatures down drill holes in the thermal areas. Twenty four wells were thermally logged near the resource areas. These wells not only resulted in accurate temperature and gradient data but helped define the intermixing of fluids within some aquifers by their isothermal character (Figure 4).

An example of a map presenting geologic and oil well thermal data is shown for the Cody area in Figure 5. The results of thermal studies including a DOE sponsored drilling program in the Cody area defines the area of greatest use to be in T.52N., R.102W., sections 2, 3, 11, and 16. In this area warm waters (34°C (93°F)) can be reached at shallow depths (51 to 300 meters (168 to 1,000 feet)). The maximum temperature of this system may approach 55 to 65°C (131 to 149°F) at depths of 260 to 500 meters (853 to 1640 feet). Warm waters will be found at the shallower depths in the more western portions of this potential use area (see Heasler and Decker, 1980).

Thermal modeling of the Thermopolis low temperature resource area predicts maximum temperatures in the Madison aquifer of 77°C (170°F) northwest of the Thermopolis townsite and 60°C (140°F) in the vicinity of the townsite. Observed temperatures in this area agree well with the model as can be observed from the temperature-depth plot in Figure 4 which has a measured maximum temperature of 71°C (161°F). Depths to the hydrothermal fluid along the Thermopolis anticline vary between 150 to 300 meters (500 to 1000 feet) (see Hinckley et al., 1981).

Conclusion

The use of oil and gas well temperature data and conductive thermal modeling have been shown to be useful techniques in defining low temperature hydrothermal systems in Wyoming. The oil and gas well data are used in locating areas of high thermal gradients by considering the effects of drilling mud temperature, rock thermal conductivities, surface temperature, and drilling duration; and by comparing oil and gas well thermal gradients to gradients computed from measured temperature-depth data. Conductive thermal modeling is accomplished by using regional heat flow data, rock thermal conductivity information, measuring thermal profiles of wells, and applying geologic constraints to the model. Parameters that have been successfully addressed by this method of thermal modeling are the maximum temperature of the hydrothermal systems, extend, and depth to hydrothermal reservoirs.

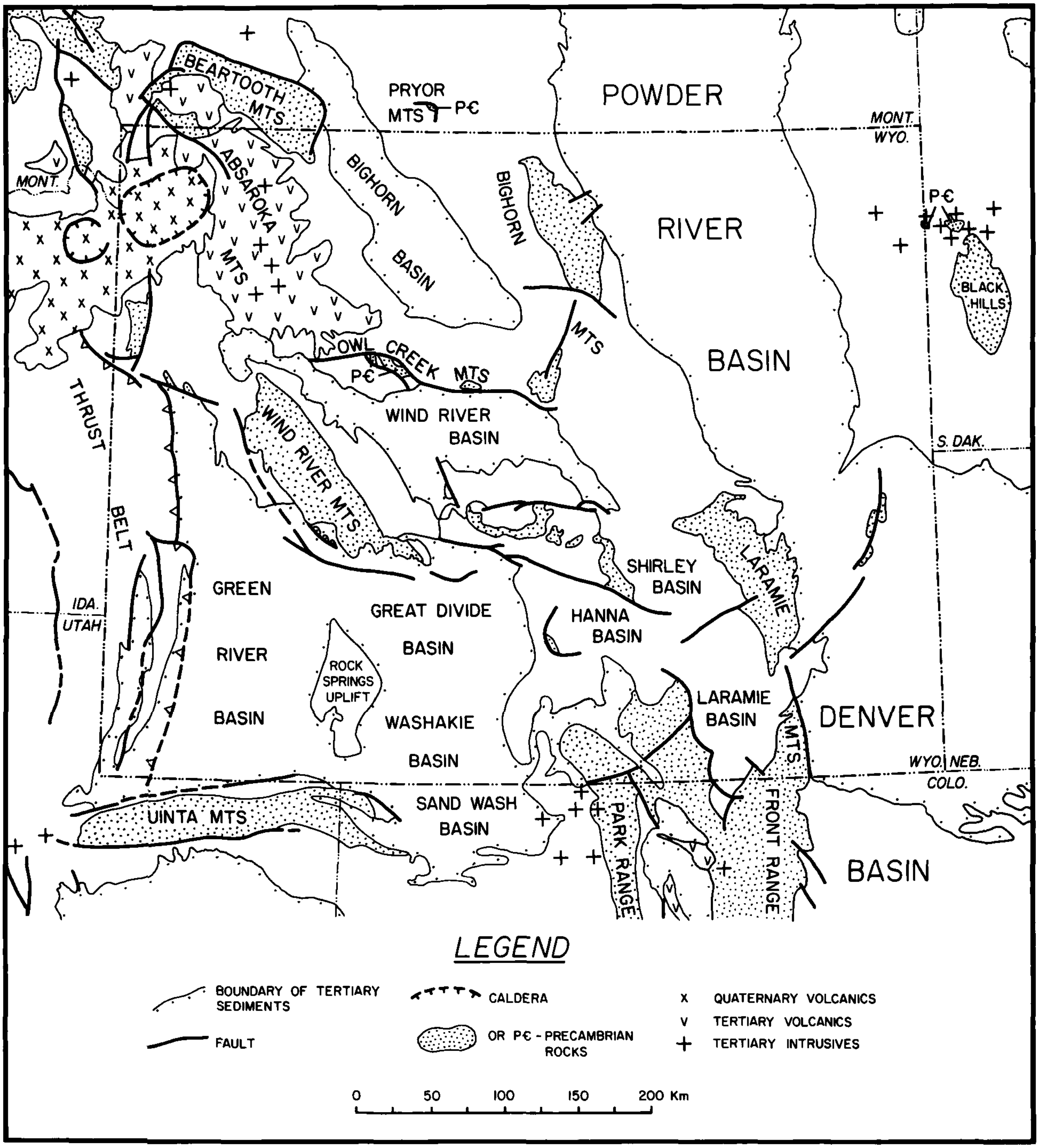
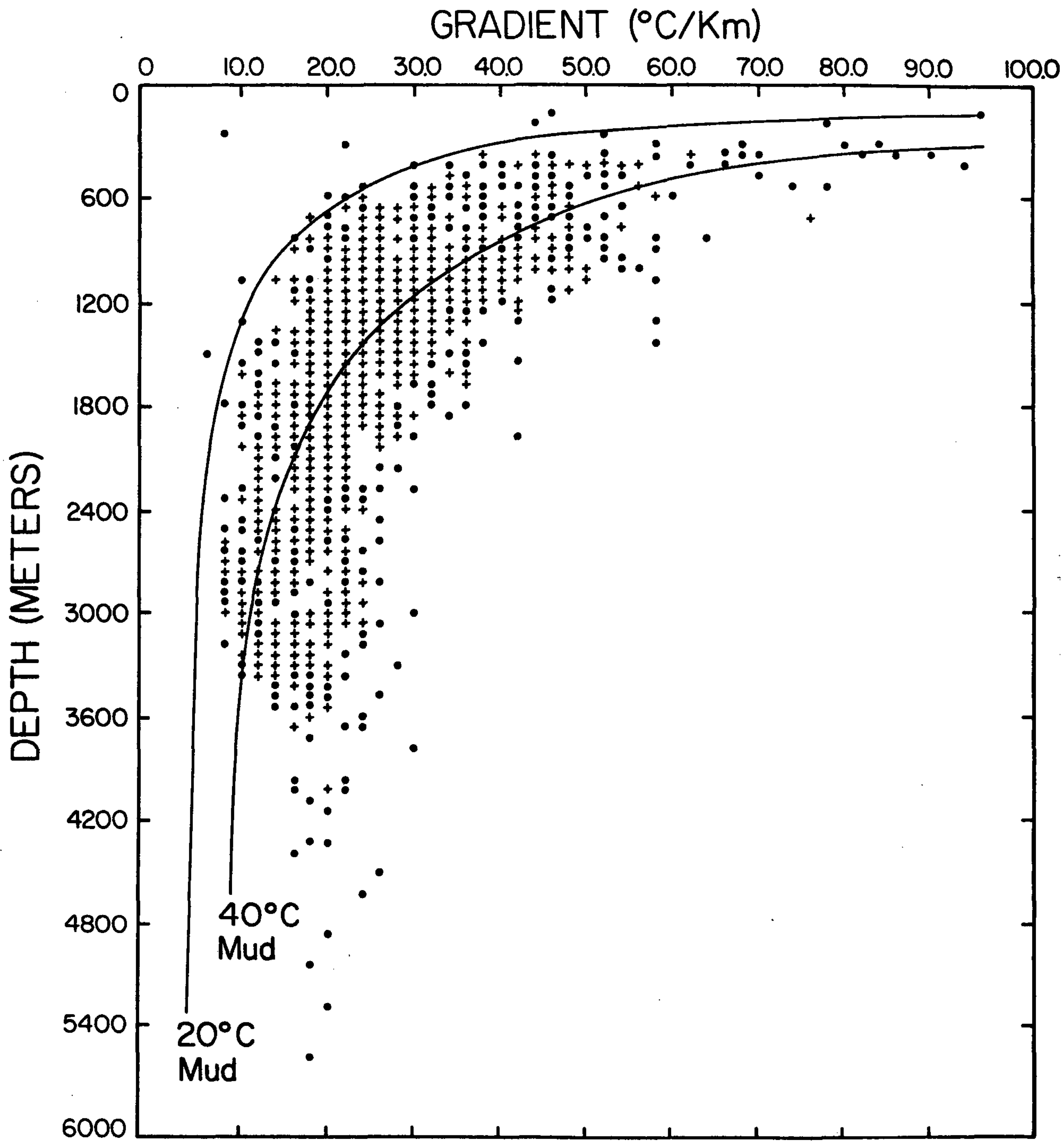


Figure 1. Generalized geology of Wyoming showing major structural basins.



GRADIENT-DEPTH PROFILE FOR THE BIGHORN BASIN

Figure 2. Oil well temperature data from the Bighorn Basin. The solid line represents a gradient assuming an isothermal drilling mud temperature. Pluses (+) represent more than one data point at that depth and gradient.

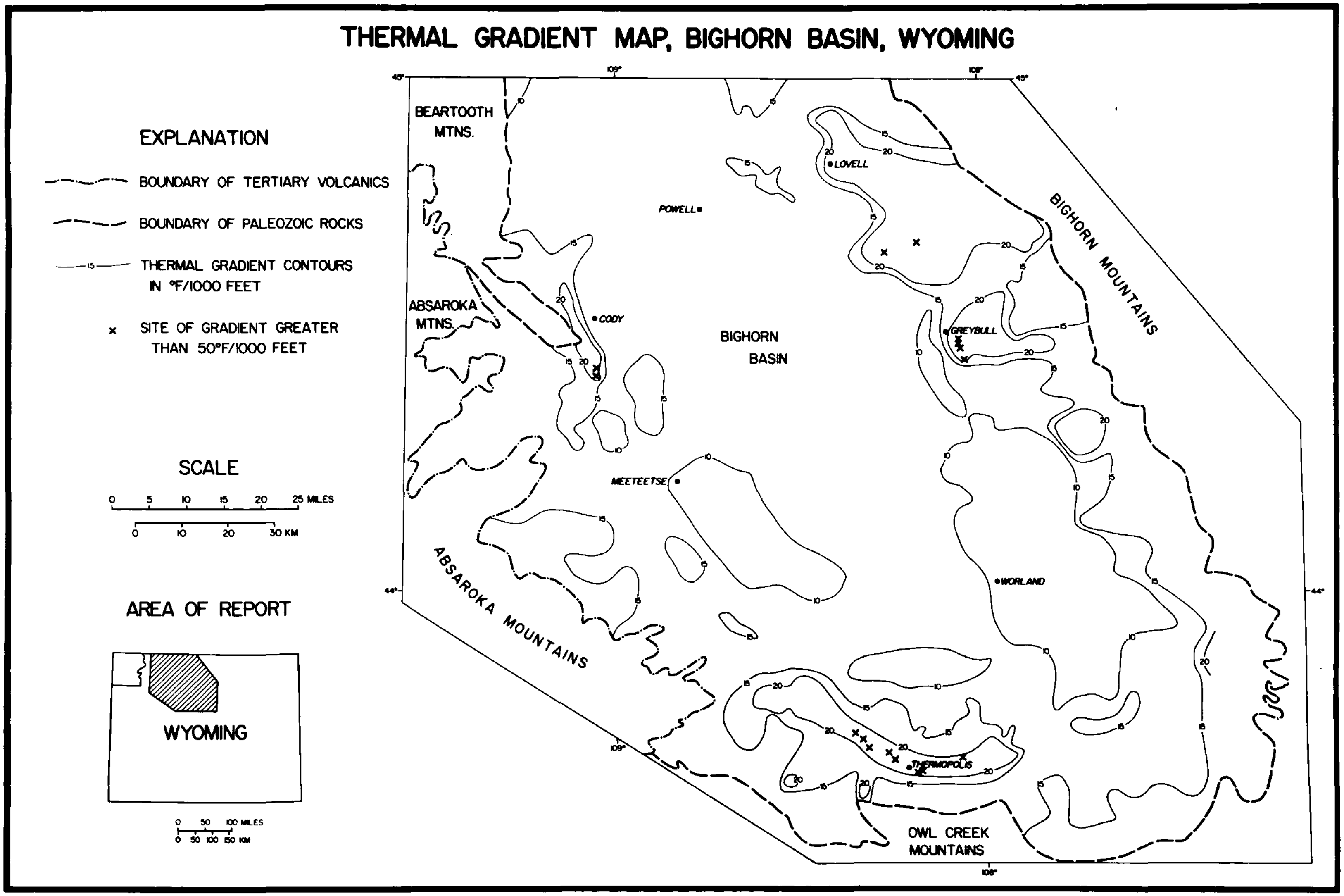


Figure 3. Thermal gradient contour map of the Bighorn Basin. Contours are in °F/1000 feet.

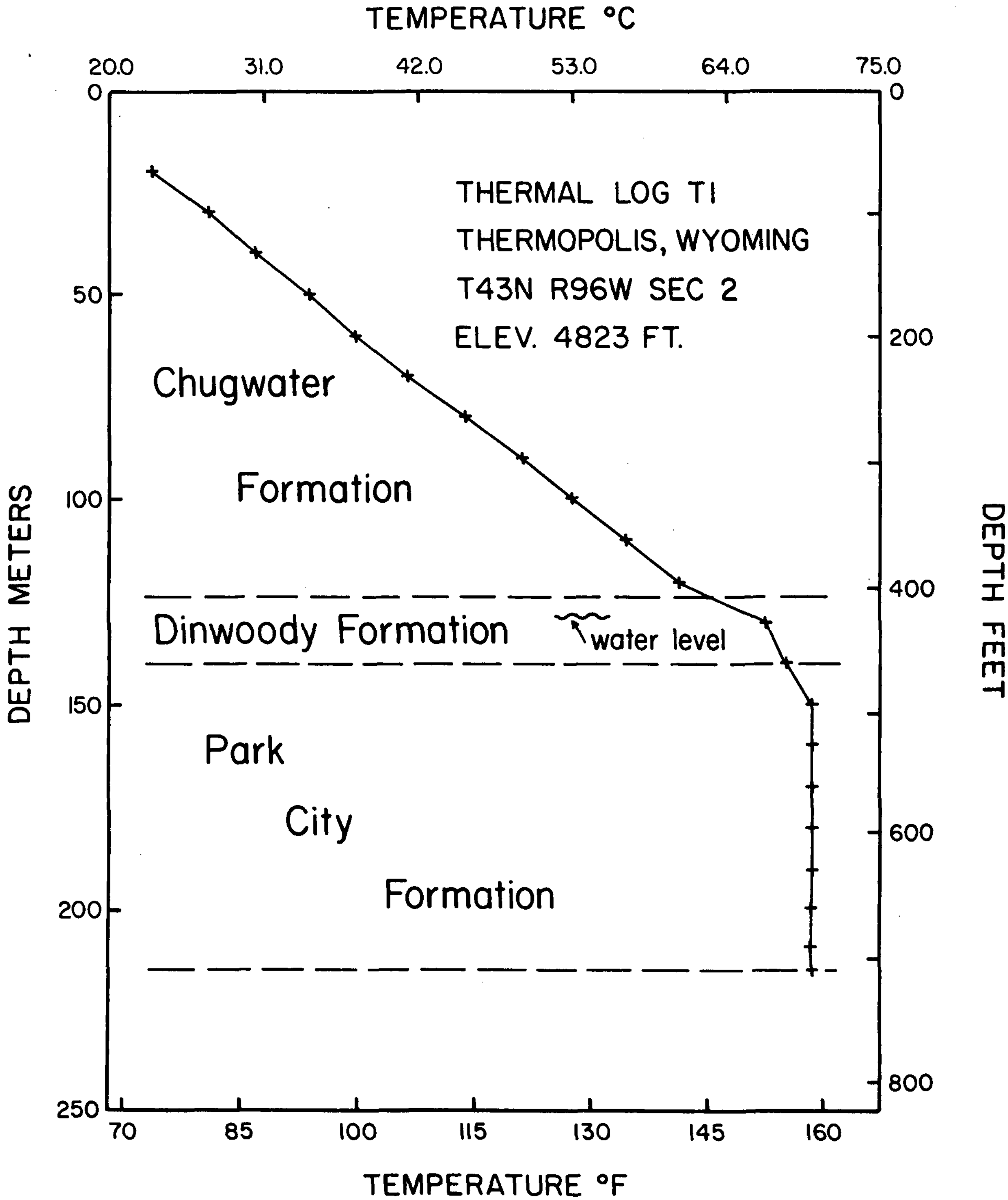


Figure 4. Temperature-depth plot for a well on the Thermopolis anticline showing an isothermal character and implied mixing zone within the Park City Formation.

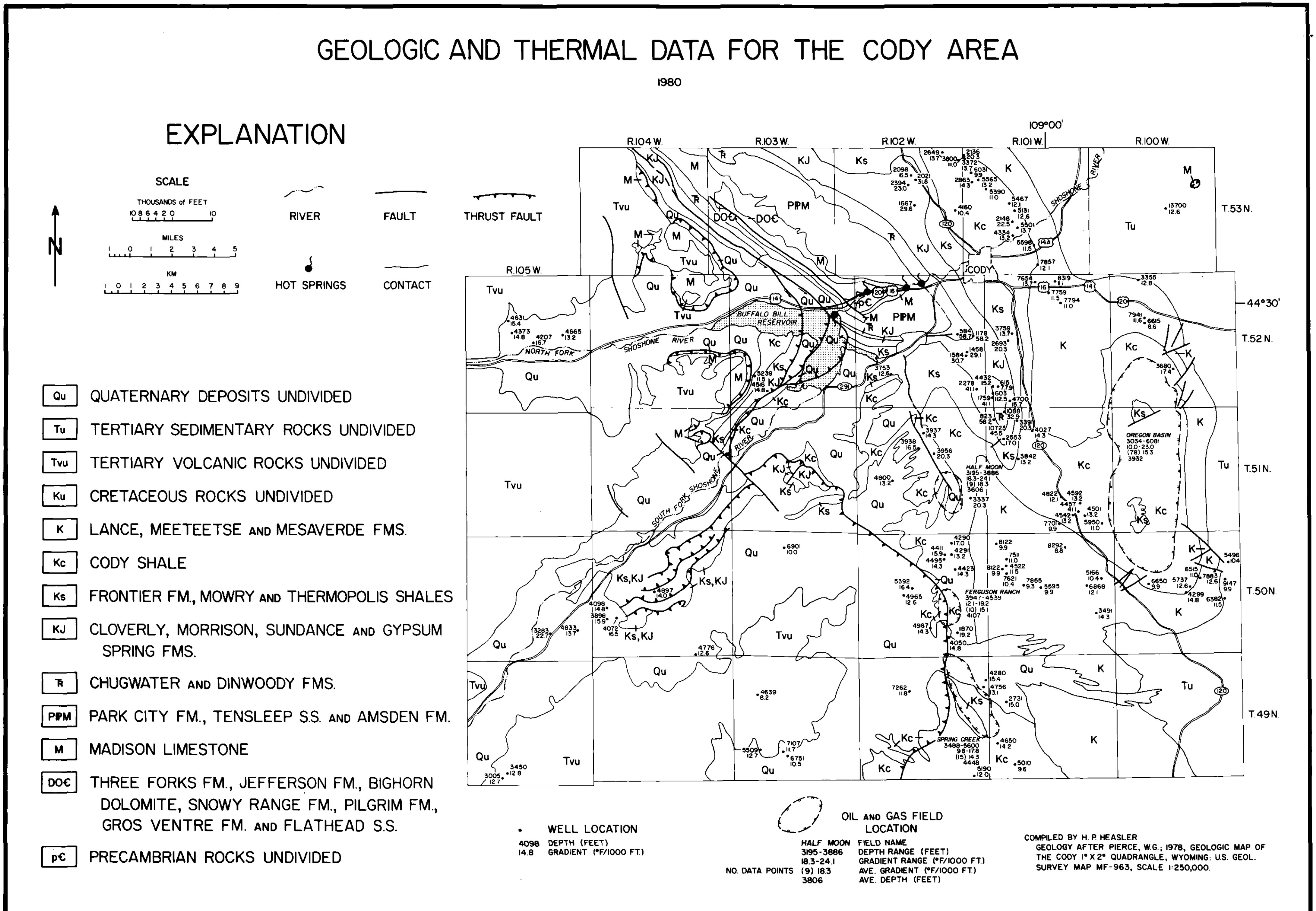


Figure 5. Geologic and thermal data for the Cody area.

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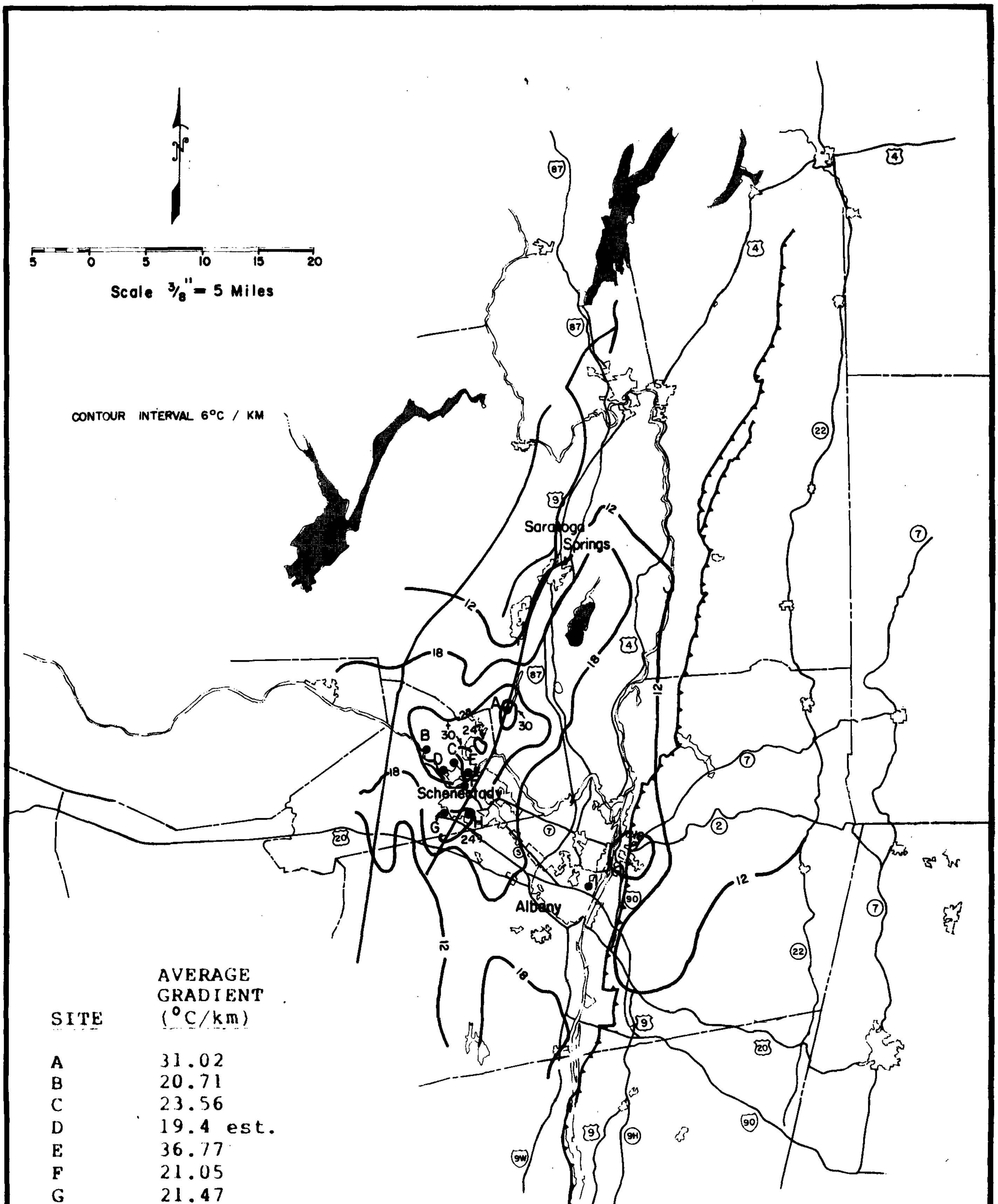


SITE	AVERAGE GRADIENT (°C/km)
A	31.02
B	20.71
C	23.56
D	19.4 est.
E	36.77
F	21.05
G	21.47
H	20.24
I	17.25

Drill Site Locations &
Generalized Gradient Contour Map

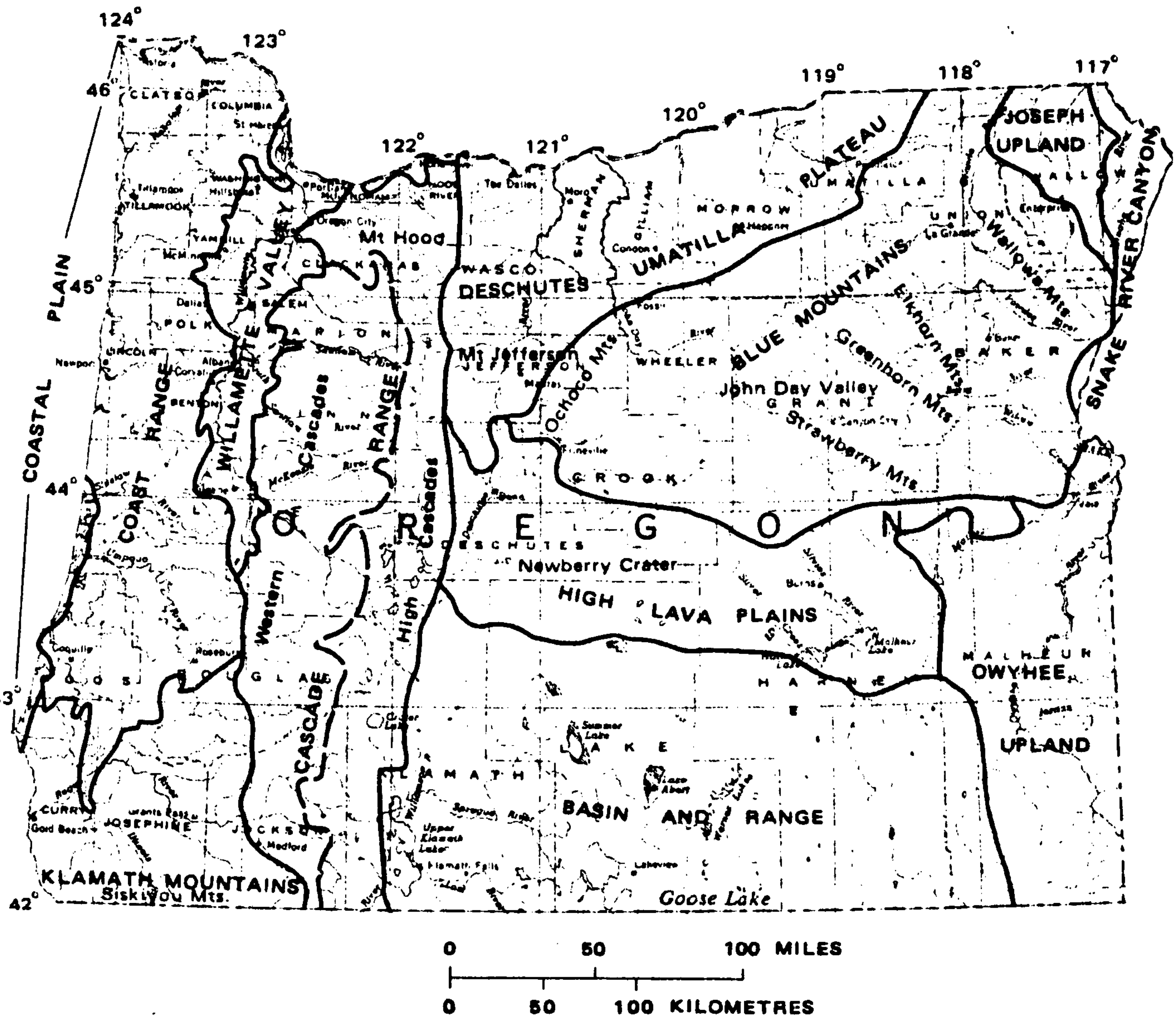


Figure 1. Physiographic provinces of Oregon.

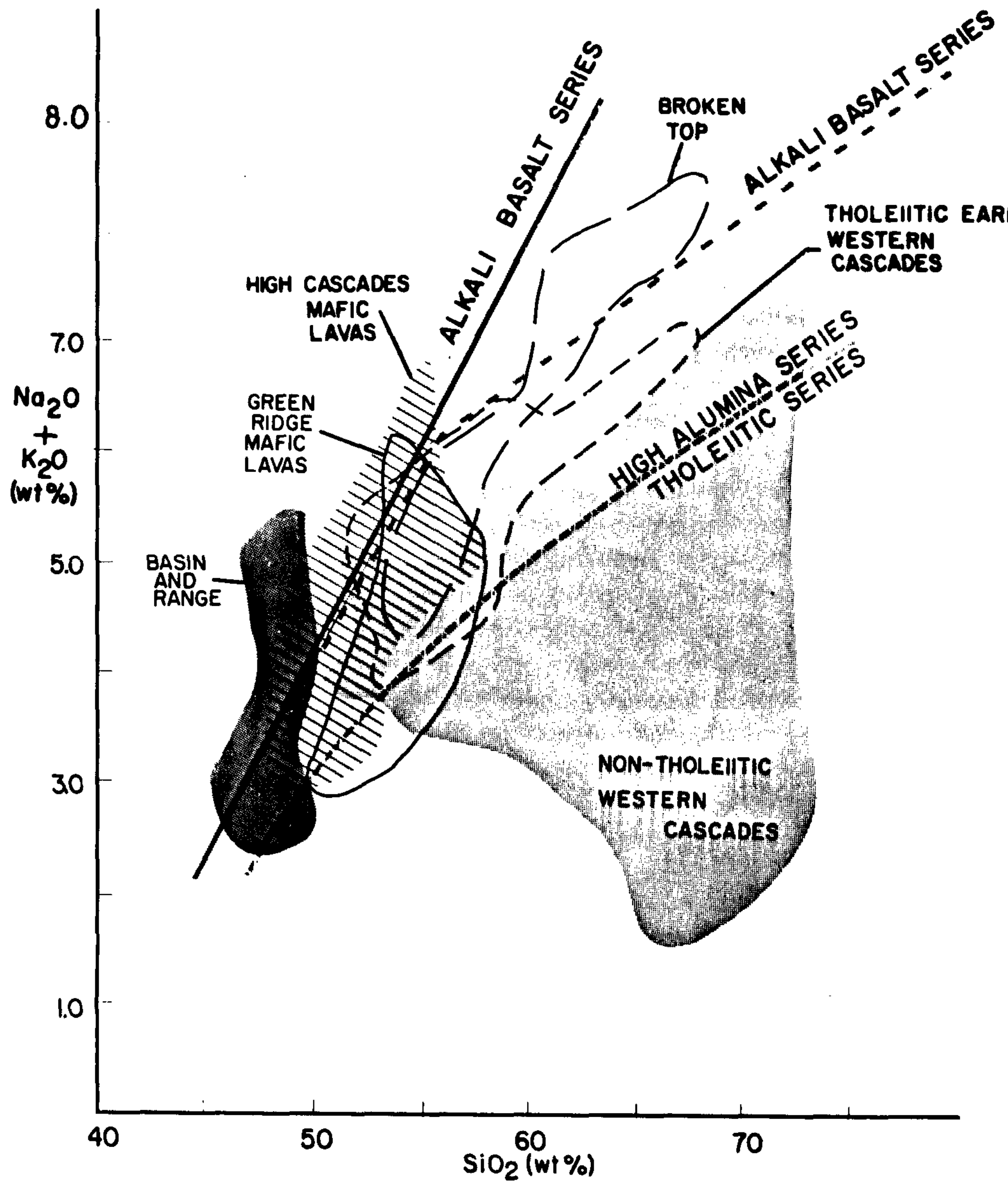


Figure 4. Compositional fields of Cascade and Basin and Range rocks shown relative to the fields of composition of Kuno's (1966) alkali basalt series (straight dashed lines) and the line separating Macdonald and Katsura's (1964) alkali basalt and tholeiitic basalt fields (straight solid line). The Basin and Range field is from Hart (1981). The High Cascade field is from data of this study with selected samples from Barnes (1978), Jan (1967), Maynard (1974), and White (1980b). The non-tholeiitic Western Cascade field is from this study. The tholeiitic early Western Cascade field is from data of this study and samples of the Scorpion Mountain lavas of White (1980a; 1980b). The Broken Top field is from Taylor (1978) and the Green Ridge field is from Hales (1975).

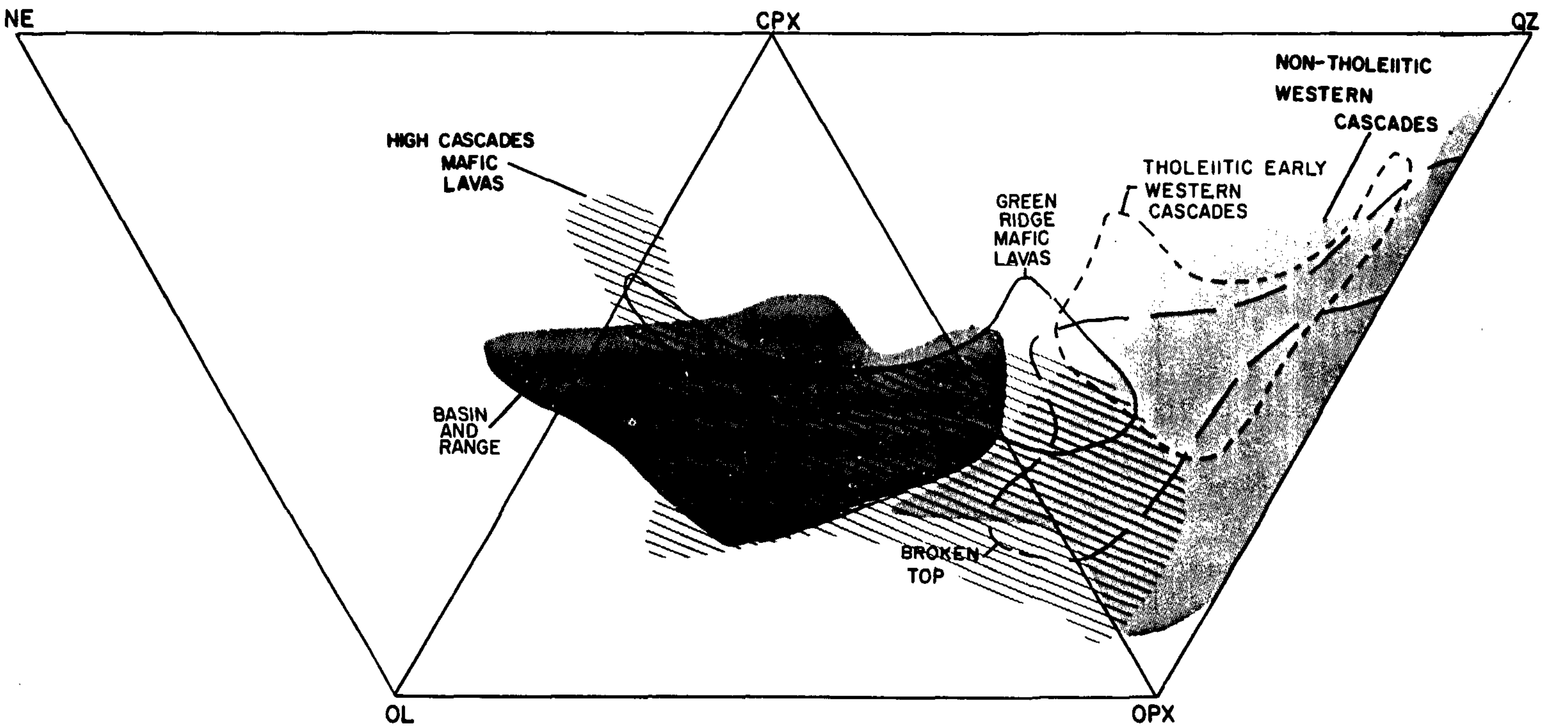


Figure 5. Nepheline (NE) - clinopyroxene (CPX) - olivine (OL) - orthopyroxene (OPX) - quartz (QZ) quadrilateral showing the same fields of composition as Figure 4. Normative minerals calculated with an Fe_2O_3/FeO molecular ratio of 0.28.

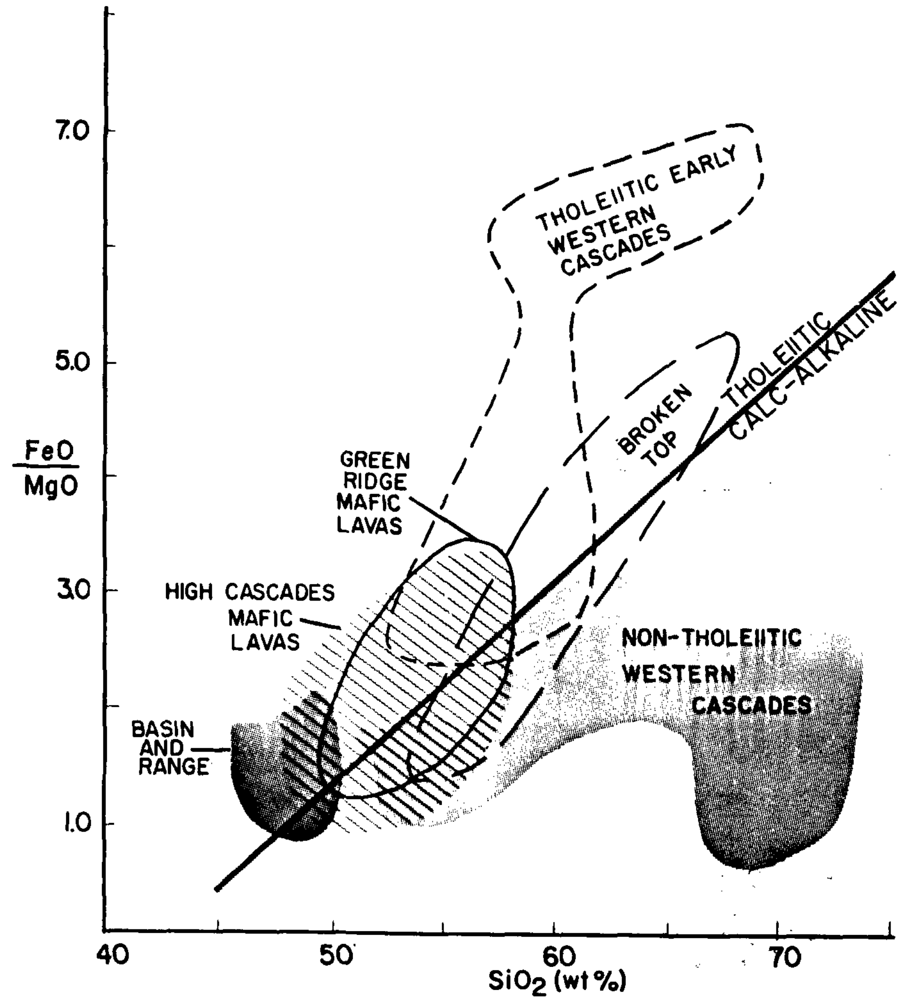


Figure 6. Iron to magnesium ratio versus silica showing the same compositional fields as Figure 4 relative to fields of tholeiitic differentiation series and calc-alkaline differentiation series of Miyashiro (1974). FeO = total iron recalculated to Fe^{+2}.

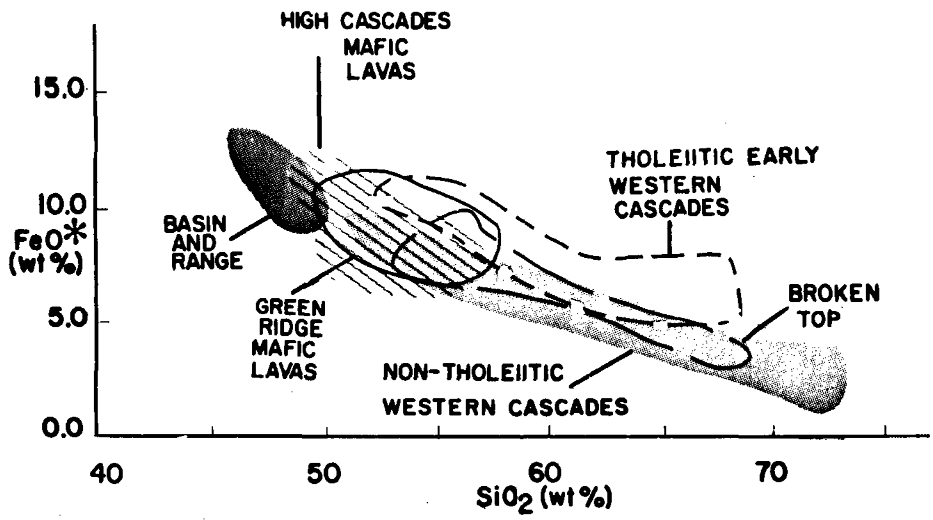
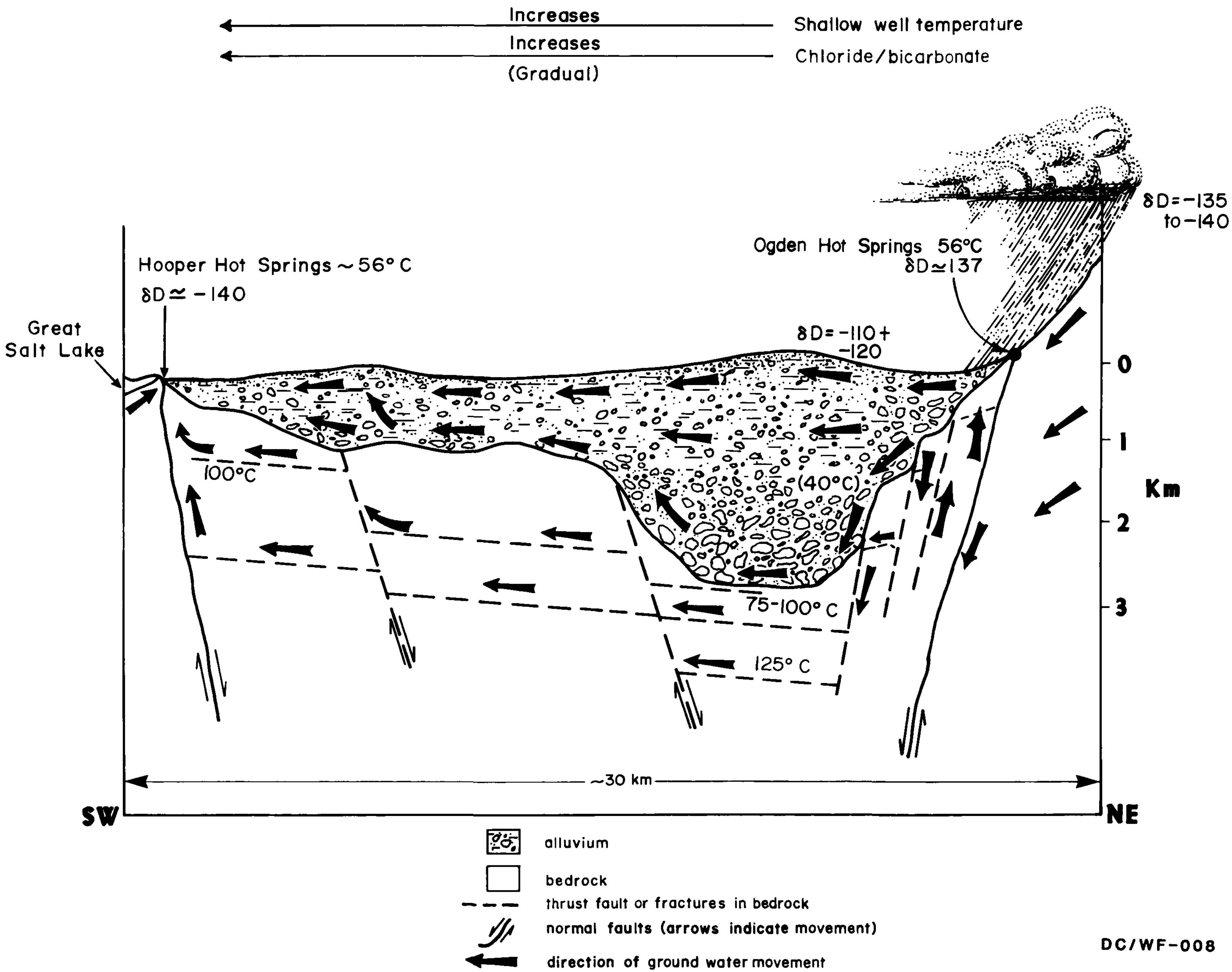


Figure 7. FeO* (total iron recalculated to Fe^{+2}) versus silica for the same fields as Figure 4.

GROUND WATER MODEL



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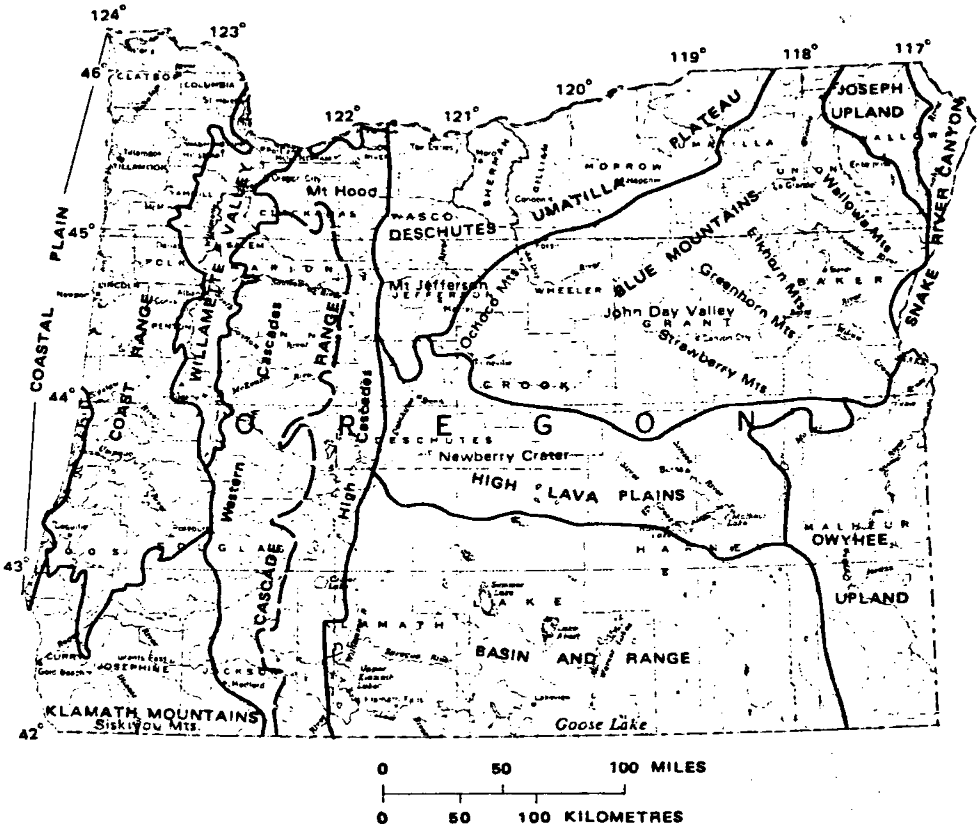

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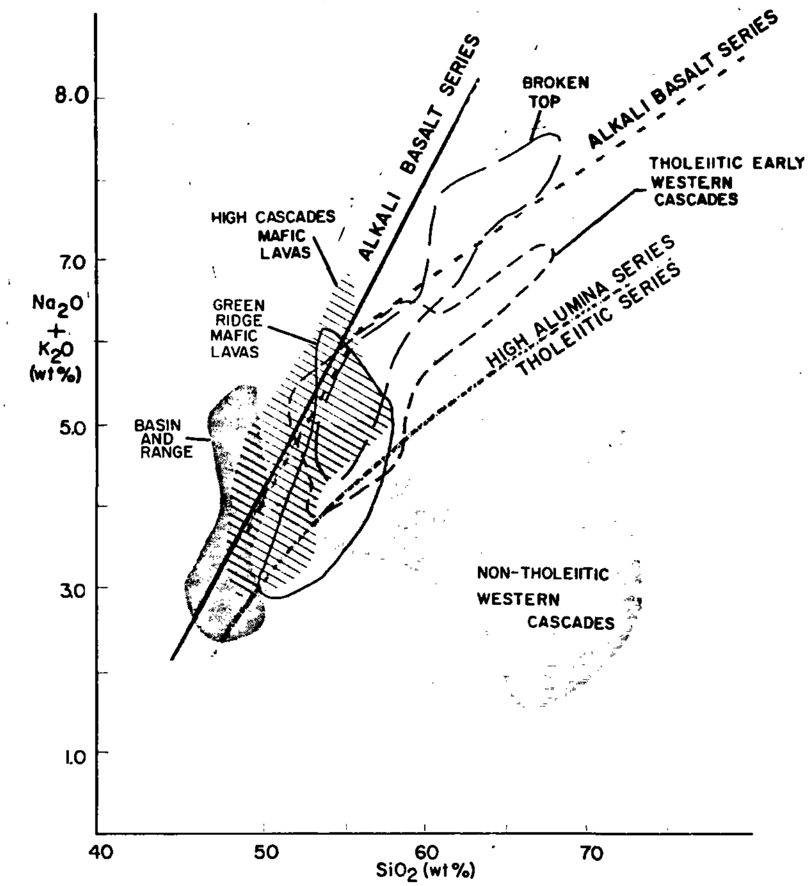

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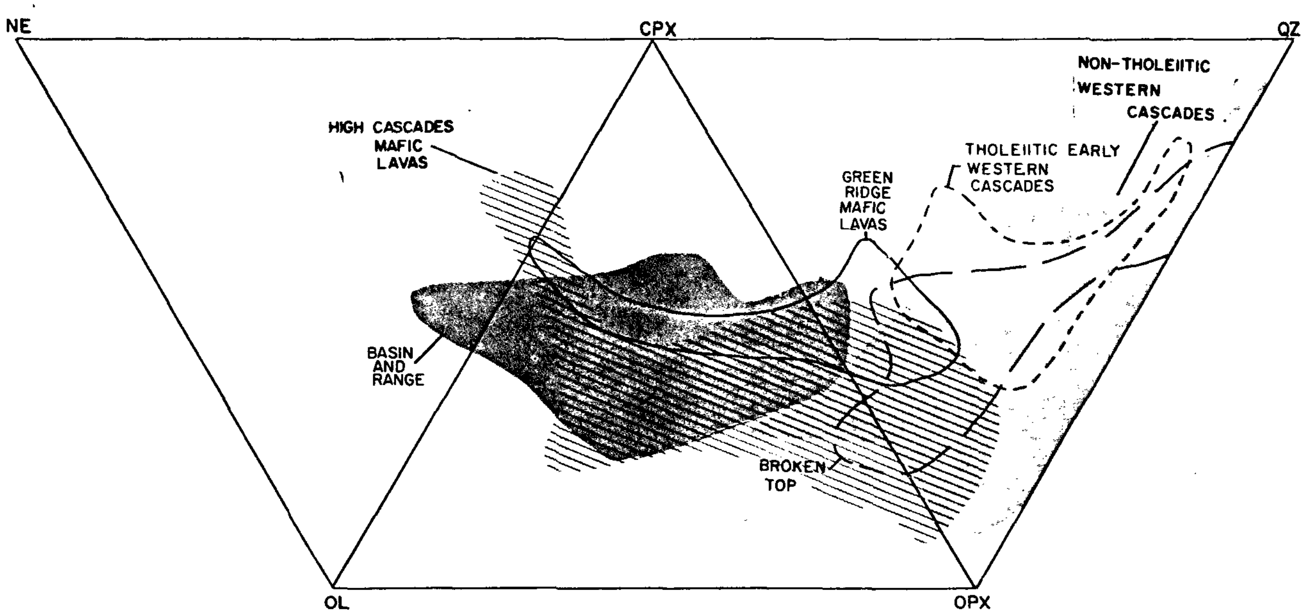

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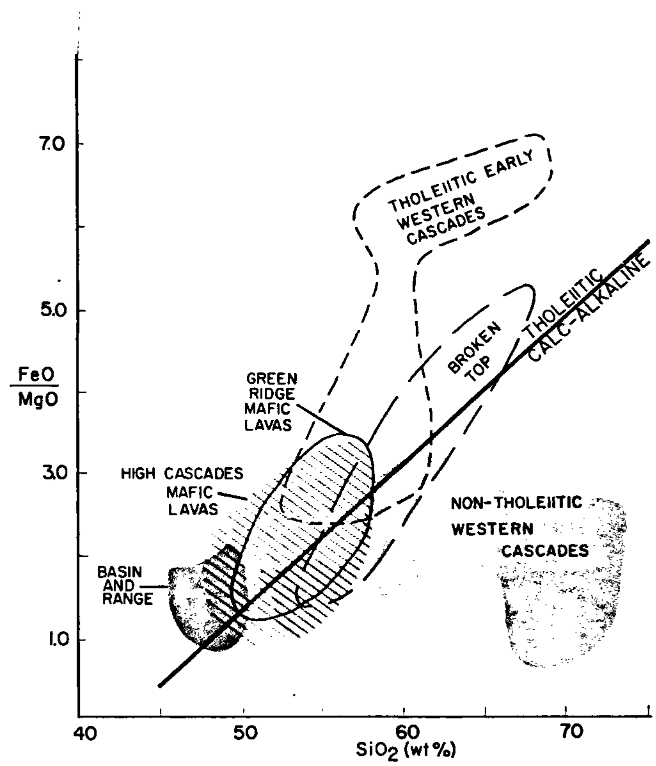

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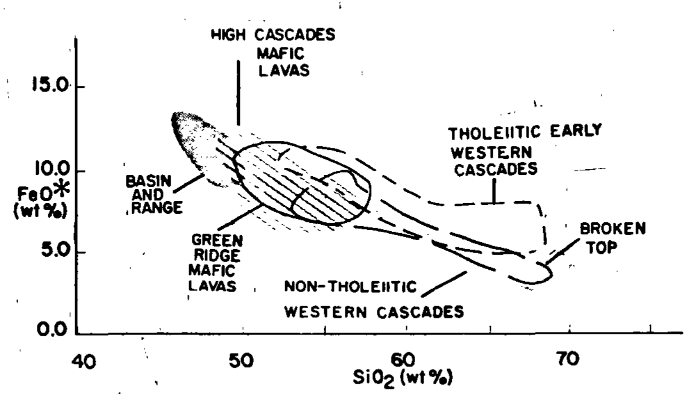

Figure 7. FeO* (total iron recalculated to Fe^{+2}) versus silica for the same fields as Figure 4.

Hawaii Institute of Geophysics

GEOTHERMAL RESOURCE ASSESSMENT PROGRAM

1981-82 PUBLICATIONS

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Cox, Malcolm E., Cuff, Kevin E., Lienert, Barry R., Sinton, John B. and Thomas, Donald M., in press, A preliminary geothermal evaluation of the Mokapu Peninsula on the island of Oahu, Hawaii: Hawaii Institute of Geophysics, Univ. of Hawaii, Tech. Rept. 82-2.

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Los Alamos

Los Alamos National Laboratory
Los Alamos, New Mexico 87545

DATE: June 28, 1982
IN REPLY REFER TO:
MAIL STOP: J978
TELEPHONE: 667/8477

Dr. Dave Blackwell
% Carl Rusetta
University of Utah Research Institute
Earth Science Laboratory
420 Chipeta Way, Ste. 120
Salt Lake City, UT 84108

Dear Dr. Blackwell:

Enclosed is my typed original of your paper and your rough draft. I hope it will be satisfactory.

Grant Heiken, Tom Shankland and Mark Anders will mull over the information gathered and when they write a report they will send you a copy.

Sincerely,

K. Florin

K. Florin

kf
Enc. a/s

XC: CRMO MS A150
G. Heiken File

HEAR FLOW AND GEOTHERMAL POTENTIAL OF THE CASCADE RANGE

David D. Blackwell

Southern Methodist University
Department of Geological Sciences
Dallas, Texas 75275

ABSTRACT

In the past three years extensive regional assessment of the geothermal potential in the Cascades has been a major part of the DOE-state-coupled program of Oregon and Washington. Part of this program has included regional studies of heat flow and geothermal gradients in holes drilled for the purposes of heat flow and in holes of opportunity. The results of these studies allow a regional evaluation of the geothermal potential of the Cascade Range that was not previously possible and at the same time the understanding of many of the factors involved in volcanic and intrusive activity in an island arc setting which effect heat flow and geothermal system development. The early part of the study indicated that very high thermal gradient and heat flow values are found in the Oregon part of the Cascade Range. There measurements over an east-west width of 20-30 km and a north-south length of 200 km indicate background gradients of 60-65^{o}C/km and heat flow values of 100 mWm^{-2}. The gradients decrease abruptly to the west in the western Cascade Range and Willamette Valley where heat flow values of 40 mWm^{-2} and geothermal gradients of 25-35^{o}C/km are typical. Just south of Mt. Hood, a north-south regional change in gradient and heat flow occurs. North of this region average gradients are approximately 50^{o}C/km and heat flow values are approximately 80 mWm^{-2}. This pattern has been documented at least as far north as Mt.Rainier. North of Mt.Rainier in the vicinity of Snoqualmie Pass there seems to be a gap in the high heat flow band. Further north in the Stevens Pass region the gradients and heat flow values increase again to values typical of those observed in the southern Washington Cascades.

In addition to the regional data, detailed exploration has taken place around Mt. Hood and a deep hole was drilled by the USGS at Newberry volcano. The studies around Mt. Hood indicate that the heating associated with some andesitic volcanoes is quite local and no major geothermal system has been

demonstrated to exist associated with the Mt. Hood volcano. On the other hand, demonstrated high temperature geothermal systems have been located associated with Lessen volcano (temperatures in excess of 180°C) and Newberry volcano (temperatures in excess of 250°C). In British Columbia BC Hydro has located a geothermal system with temperatures in excess of 200°C associated with Meager Mountain volcano.

The very high regional gradients, the large extent of the area and its association with young volcanics indicates the geothermal potential of the Cascade Range is very large. Deeper drilling in potential geothermal systems remains for a complete evaluation of the geothermal potential.

HEAT FLOW ~~IN~~ AND GEOTHERMAL POTENTIAL~~S~~ OF THE CASCADE~~S~~ RANGE

By David D. Blackwell
Southern Methodist University
Department of Geological Sciences
Dallas, Texas 75275

ABSTRACT

In the past three years ~~the~~ extensive regional assessment of the geothermal Cascades has been a major part of the DOE state-coupled program of Oregon and Washington. ~~A major~~ part of this program has included regional studies of heat flow ~~in~~ and geothermal gradient in holes drilled for the purposes of ~~geothermal exploration~~ heat flow and in holes of opportunity. The results of these studies allow a regional evaluation of the geothermal potential of the Cascade Range that was not previously possible and at the same time the understanding of many of the ~~physical~~ factors ~~which are~~ involved in volcanic ~~activity~~ and intrusive activity in an island arc setting which effect heat flow and geothermal system development. ~~Previous~~ Early parts of this studies ~~have~~ indicated that very high thermal gradient ~~or~~ and heat flow values are found in the Oregon part of the Cascade Range. ~~In this area,~~ There measurements over ~~a~~ an east-west width of 20-30 km and a north-south length of 200 km indicate background gradients of 60-65°C ~~per km~~ /km and heat flow values of 100 mW ~~per square meter~~ m^{-2}. The gradients decrease abruptly to the west in the Western Cascade~~s~~ Range and ~~in~~ Willamette Valley where heat flow values of 40 mW ~~per square meter~~ m^{-2} and geothermal gradients of 25-35°C ~~per km are typical~~ /km are typical. Just south of Mt. Hood ~~the~~ a N-S regional change in gradient and heat flow occurs. ~~to the north of Mt. Hood~~ North of this region ~~Regional~~ average gradients are approximately 50°C ~~per km~~ /km and heat flow values are approximately 80 ~~mW per square meter~~ mWm^{-2}. This pattern has been documented at least as far north as Mt. Ranier. North of Mt. Ranier in the vicinity of Snoqualmie Pass there seems to be a gap ~~with low heat flow.~~ in the high heat flow band. ~~There may be a gap in the high heat flow in the vicinity of ___ pass where 2 low heat flow value and gradient holes have been obtained. To~~ Further the north in the Stevens Pass region the gradients and heat flow values increase again to values ~~of those~~ typical of those observed in southern Washington Cascades. ¶ In addition to the ~~gradients on a~~ regional data ~~basis~~, detailed exploration has taken place ~~beyond~~ around Mt. Hood and a deep hole~~s~~

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GEOTHERMAL RESOURCE ASSESSMENT IN OKLAHOMA

William E. Harrison
Kenneth V. Luza
Oklahoma Geological Survey
Norman, Oklahoma 73019

M. Lynn Prater
E-Systems
Greenville, Texas 75401

Paul K. Cheung
Tangram Resources LTD.
Calgary, Alberta T2P 1J2

SUMMARY

In September 1980, the Oklahoma Geological Survey began a program to assess the geothermal potential of the State. The program, thus far, consists of (1) the preparation of a detailed geothermal-gradient map of Oklahoma at a scale of 1:500,000 and (2) site-specific investigations of gradient and subsurface conditions in areas that appear to have geothermal potential.

Prior to this investigation, the best available geothermal-gradient map for Oklahoma was prepared by Cheung (1978) as part of his thesis investigation at Oklahoma State University. The American Association of Petroleum Geologists' (AAPG) North American Geothermal Project (1976) provided the data base for the Oklahoma State University work, although initially the Oklahoma Panhandle and northeastern and southeastern Oklahoma were excluded from Cheung's study. Several well-correction factors (e.g., maximum time since circulation, air-drilled versus mud-drilled, and geologic province) were applied to the raw data and to electric-log data in order to determine temperature gradient.

In 1981, Cheung expanded his geothermal-gradient program to include the unmapped areas of the State. The mapping for the Panhandle was completed in August 1981. Unfortunately, temperature data for the northeastern and southeastern parts of the State were not detailed enough to complete a temperature-gradient contour map.

Two areas where recent mapping has shown the high gradients (2.1°F/100 feet) were selected for detailed study. These areas are in (1) Haskell and

(2) Pittsburg Counties and are subsequently referred to as the Haskell and Pittsburg anomalies. We estimated volume and deliverability of formation water potentially available from several sandstone units for geothermal applications. The Spiro and Cromwell sands were chosen for the Haskell anomaly and the Hartshorne sandstone was chosen for the Pittsburg anomaly. We hope similar investigations of subsurface formations can be expanded into other areas which have relatively "high" geothermal gradients.

GEOTHERMAL-RESOURCE APPRAISAL

Introduction

A number of attempts have been made to map geothermal gradients in Oklahoma. McCutchin (1930) recognized a correlation between oil-bearing anticlinal structures and high geothermal gradients. Schoeppel and Gilarranz (1966) prepared a geothermal-gradient map of Oklahoma from corrected bottom-hole temperatures (fig. 1). Their study indicated that actual formation temperature could be determined from temperature measurements and the time since circulation in the well bore, provided that certain factors are known, such as (1) temperature of the drilling mud at the surface, (2) pipe size, (3) circulation rate, (4) size of annulus, and (5) heat capacities and thermal conductivities of the drilling fluid, drill pipe, and country rock. A comprehensive study sponsored by the AAPG was initiated in 1968 to map geothermal gradients in North America. That portion of the resulting map covering Oklahoma is shown in figure 2.

The best available geothermal-gradient map for Oklahoma was prepared by Cheung (1978, 1979). This map, which includes the 1981 mapping in the Oklahoma Panhandle, is shown in figure 3. Several correction factors were applied to the raw temperature data to reflect actual formation temperature more accurately. The procedures and methods used to develop this temperature map (figure 3) are discussed below.

(fig. 5). These areas are in (1) Haskell and (2) Pittsburg Counties, in the Arkoma Basin.

The Arkoma Basin is composed of a series of anticlines and synclines in Pennsylvanian clastic rocks, broken by thrust and (or) growth faults near the center of the basin (Fay, 1970; McQuillan, 1977). The resistant sandstones cap high ridges near the centers of the synclines, whereas shales occupy stream valleys along anticlinal axes.

Three sandstone units, the Spiro, Cromwell, and Hartshorne, were selected as potential sources of water for low-temperature geothermal applications (fig. 6). Completion cards were used to assist in determining the tops of these sandstone units on well logs. Each sand exhibits a characteristic pattern on the spontaneous-potential and (or) gamma-ray track as well as the resistivity track of the logs. Thickness and water saturation were then calculated from the log information.

These sandstone units were chosen because of their continuity over the anomalous areas. The Hartshorne was chosen for the Pittsburg anomaly, and the Spiro and Cromwell were chosen for the Haskell anomaly. All three sands are gas productive, so it is common for them to be penetrated and logged by drilling companies. The deeper Spiro and Cromwell sands have not yet been penetrated in the area of the Pittsburg anomaly, however.

Water-Volume Estimate

An isovolume map was prepared for each formation. In this process, the thickness of water at each well location can be calculated. Because of the size of the area, one well per section was used in this study. The amount of water that can be produced from one well is estimated by the following equation:

from each 20 feet of the formation.

Summary

The minimum amount of water in a locality can be estimated by using the isovolume map. Furthermore, an estimate can be made over an area by using the isopach map along with average values for porosity and thickness. A summary of formation characteristics and minimum water-in-place estimates for the Hartshorne, Spiro, and Cromwell sandstones are listed in table 1.

The initial flow of water into the bore hole can be determined by using Darcy's Law. Values for the variables in the equation can be taken from appropriate maps (except for permeability). It is best to have a laboratory analysis for good permeability data. For a rough estimate, a chart of empirical petrophysical relationships can be used.

Because of the uncertainty involved in such calculations, site-specific areas which may be considered for geothermal applications should be subjected to detailed studies. Such studies would be in order so that an operator could estimate the productive lifetime of a given low-temperature geothermal resource. Estimates of in-place water volumes and deliverability calculated for the present study are regional in scope and are intended as order-of-magnitude assessments.

FUTURE INVESTIGATIONS

Temperature data from geophysical logs usually indicate geothermal gradients that are somewhat lower than measurements obtained after thermal equilibration. Therefore, a temperature-confirmation program will be initiated to determine the relationship between temperature data from geophysical logs and equilibration temperatures. Then we can ascertain (1) the magnitude of the variation and (2) whether the variation is systematic.

If the difference between geophysical-log temperature and true temperature is systematic, it may be possible to make a standard correction to the geothermal-gradient map in order to obtain an approximation of equilibration temperatures. Should the difference, however, vary with other characteristics, such as petrophysics or geologic province, correction factors will be somewhat more complicated.

The temperature-confirmation program will consist of two systems, non-retrievable and retrievable. The first system involves the continuous recording of temperatures in recently abandoned oil and (or) gas test holes for a period of several months. A temperature sensor will be attached to a 7-conductor cable and lowered into an abandoned borehole to depths to a depth of 500 feet. Concrete will be added to the top 30 feet of the borehole, thus making this installation somewhat permanent. A 10-channel analog data logger will continuously record the temperature data on a cassette tape. The tape will be retrieved once a month and processed at the Oklahoma Geophysical Observatory. Time-temperature data will be compared with temperature data obtained from geophysical logs. The difference between the two temperatures will enable us to assess equilibration variations. The Pittsburg and Haskell anomalies are the principal target areas. We plan to install several sensors in recently drilled abandoned oil and (or) gas wells drilled in the target areas. Thus far, almost every oil and gas well recently drilled in this region has been productive.

The second system will use a retrievable temperature probe to perform the temperature-assessment work. A temperature sensor will be attached to a 3,000,-foot 4-conductor logging cable. A D.C.-powered hoist will be used to raise and (or) lower the temperature probe. A ditigal voltmeter can be used to measure voltage variations, which can be converted to temperature measurements.

This system will be used principally in abandoned boreholes drilled during the last 10 years. The surface plug, generally 30 to 50 feet thick, will be drilled out. Then the temperature probe will be lowered into the borehole to depths between 700 and 1,200 feet. Generally the depth will depend upon the length of surface casing left in the hole.

We feel that both of these systems will provide temperature data that will give us even greater reliability in evaluating bore-hole geophysical-log temperatures in Oklahoma.

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PROGRESS REPORT ON THE

GEOTHERMAL ASSESSMENT OF THE

JORDAN VALLEY, SALT LAKE COUNTY,

UTAH

by

Robert H. Klauk, Riki Darling, Deborah A. Davis,
J. Wallace Gwynn and Peter J. Murphy

May, 1981

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LIST OF ILLUSTRATIONS

ABSTRACT

Two known geothermal areas have been investigated previously by Murphy and Gwynn (1979) in the Jordan Valley, Salt Lake County, Utah. These two reports indicate meteoric water is being circulated to depth and heated by the ambient temperature derived from normal heat flow. This warm water subsequently migrates upward along permiable fault zones.

The gravity survey conducted in the valley indicates a number of fault blocks are present beneath the unconsolidated valley sediments (Plate II). The faults bounding these blocks could provide conduits for the upward migration of warm water.

Four areas of warm water wells, in addition to the two known geothermal areas, have been delineated in the valley (Plate III). However, the chemistry of the Jordan Valley is quite complex and at this time is not fully understood in regard to geothermal potential (Plates IV and V).

Thick sequences of unconsolidated valley fill could conceal geothermal areas due to lateral dispersion or dilution within the principal aquifer, as well as retardation of warm water flow allowing time for cooling prior to discharge in wells or springs. Other areas are possibly diluted and cooled by high quality, ground water recharge from snow melt in the Wasatch Range.

INTRODUCTION

The Utah Geological and Mineral Survey (UGMS) has been conducting research to advance the utilization of low temperature geothermal resources in the State of Utah as per U. S. Department of Energy (DOE) contract DE-AS07-77ET28393. Prior to this study, UGMS was concentrating its investigations on known geothermal areas along the Wasatch Front from Utah Valley north to the Idaho/Utah state line. The concentration of the study in this region was due primarily to the number of known geothermal resources near major population centers of the state, hopefully resulting in timely resource development.

In February, 1980 it was determined that efforts should begin in the evaluation of the area wide geothermal resource potential of the following Wasatch Front Valleys: (1) Utah Valley, (2) Jordan Valley; (3) Ogden Valley; (4) Bear River Valley; (5) Malad Valley, and (6) Cache Valley. These areas were decided upon because of their inherent low temperature geothermal potential and because they encompass the three major population centers of the state. The initial major effort in this assessment study was concentrated in the Jordan Valley because of the inclusion of Salt Lake City, the major population center.

PHYSIOGRAPHIC SETTING

The Jordan Valley encompasses an area of approximately 1024 square kilometers (400 square miles) in north-central Utah in central Salt Lake County (figure 1). This valley has substantial relief, ranging in elevation from 1280 meters (4200 feet) at the Great Salt Lake to approximately 1585 meters (5200 feet) where adjoining the mountains. The east side of the valley is a boundary between two major physiographic provinces: the Rocky Mountain province to the east and the Basin and Range province to the west.

The Jordan Valley is bounded to the east, south, and west by the Wasatch, Traverse, and Oquirrh mountain ranges, respectively, while the north end is open to the Ogden Valley with an arbitrary boundary being an extension of the Salt Lake salient which is an intermediate fault block that extends west from the Wasatch Mountains into the valley for approximately 6.4 km (4 miles), (figure 1). The Wasatch Mountains, including the Salt Lake salient, are part of the Rocky Mountain physiographic province while the Traverse and Oquirrh Mountains are part of the Basin and Range province.

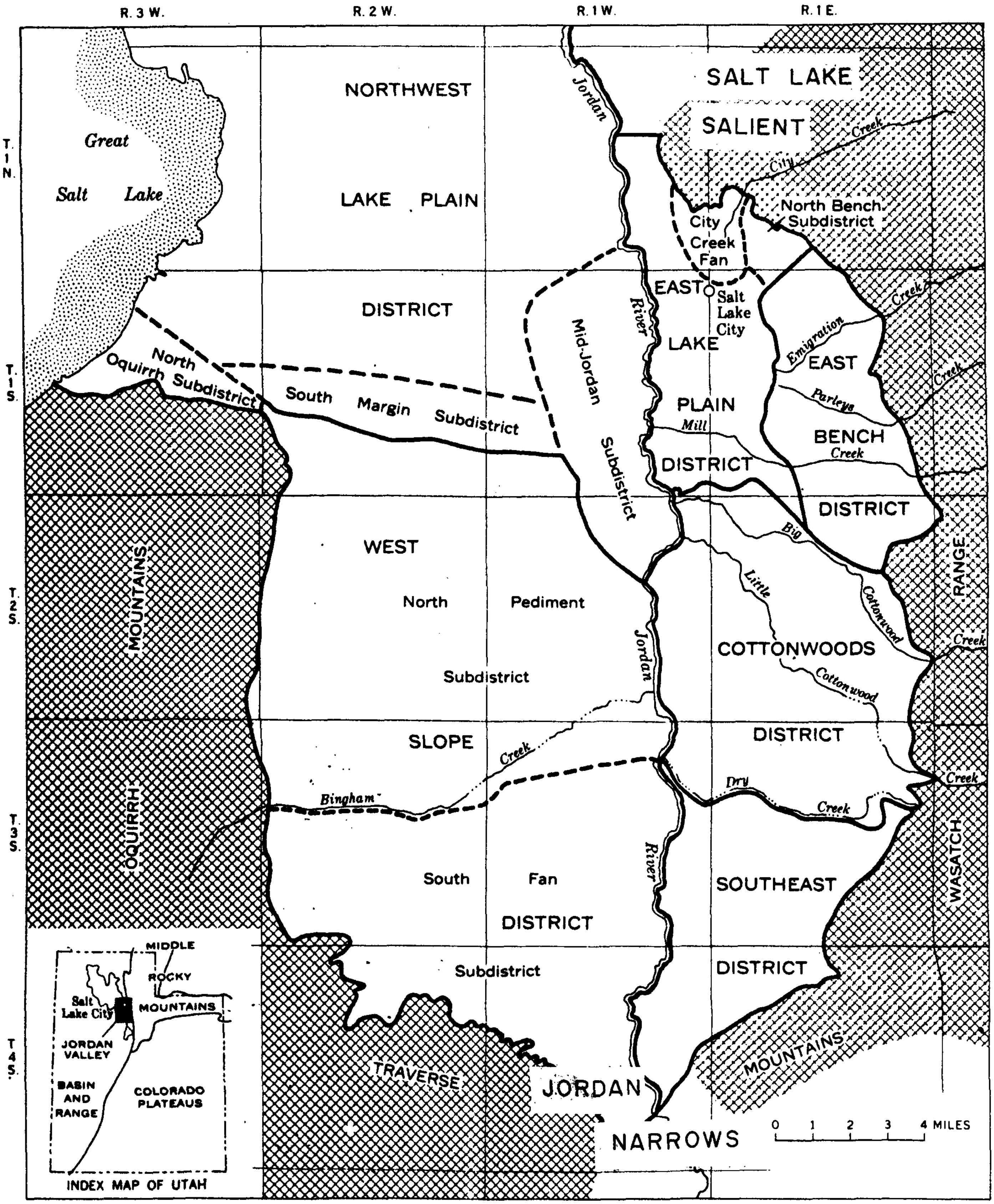

Figure 1. Index map of the Jordan Valley, Utah, showing ground-water districts and subdistricts (modified from Marine and Price, 1964)

A principal water source to the Jordan Valley is the Jordan River which flows north from Utah Valley through a gap in the Traverse Mountains referred to as the Jordan Narrows and continues through the entire length of the valley, eventually entering into the Great Salt Lake. In addition to the Jordan River, the other principal water sources are the creeks draining the Wasatch Mountains (figure 1).

REGIONAL GEOLOGIC SETTING

Rocks of Pre-Cambrian through Pliocene Age are exposed in the mountains bordering the Jordan Valley. In the Wasatch Mountains, sedimentary and metamorphic rocks consist of Pre-Cambrian, Paleozoic, Mesozoic, and Cenozoic sandstone, limestone, shale, conglomerate, siltstone, tuffaceous clay, tillite, quartzite, shist and gneiss. Intrusive rocks consist of early Tertiary monzonite, diorite and granodiorite, and in places are covered by Pleistocene glacial deposits as well as alluvium.

The core of the Traverse Mountains is primarily composed of the Pennsylvanian Oquirrh formation consisting of quartzite with some calcareous sandstone and limestone. In some areas the quartzite has been broken and cemented in place – Marsell (1932) referred to this as "autoclastic breccia". Also, some Mississippian-Pennsylvanian age Manning Canyon shale and Mississippian Great Blue Limestone are present. Tertiary rocks consist of the Salt Lake group of Slentz (1955) which generally is composed of marlstone, mudstone, siltstone, travertine, and fanglomerate. Tertiary volcanics, primarily of andesite and augite-andesite porphyry composition, are also present.

Rocks of the Paleozoic Oquirrh Mountain facies are the primary units exposed in the Oquirrh Mountains. The following formations, according to Crittenden (1964), comprise the Oquirrh Mountain facies: The Great Blue Limestone, the Manning Canyon shale, the Oquirrh Formation, the Kirkham Limestone, and the Diamond Creek Sandstone. These formations are somewhat different in the central and northern parts of the range where units in the central area are referred to as the Bingham sequence while units to the north are the Rogers Canyon sequence. Tertiary rocks, also, are exposed in the Oquirrhs. These include: (1) Harker's fanglomerate of Slentz's (1955) Salt Lake group, (2) andesite and latite-andesite flows, and (3) intrusive stocks, sills and dikes of granite, monzonite, granite porphyry and rhyolite - quartz latite.

Generally, the sediments exposed in the Jordan Valley consist of unconsolidated deposits of boulders, gravel, sand, silt, and clay deposited by streams, lakes, glaciers, wind, and mass wasting during Quaternary and Recent time. Isolated outcrops of pre-Quaternary rocks are found in areas where pediments extend from the bordering mountains. Subsurface sediments differ greatly from surface sediments in the valley and will therefore be described later in detail, since they comprise the aquifers tapped by the wells investigated for the geothermal assessment study.

REGIONAL STRUCTURAL SETTING

The Jordan Valley is at the intersection of three major tectonic elements: (1) the north-trending thrusts and folds known as the Sevier orogenic belt which extends from southern Nevada to the northwest corner of Alaska (Crittenden, 1976); (2) the east-trending Uinta Arch; and (3), the north-south trending Basin and Range faulting.

Thrusting

Three episodes of thrusting have been discovered in the Wasatch Mountains east of the Jordan Valley. The first two are known as the Alta and Mount Raymond thrusts dated at 125 and 85-90 million years old respectively (Crittenden, 1964). The third, and most extensive, is the Charleston-Nebo Thrust, reported to have a displacement of 64 kilometers (40 miles) or more from the west and dated at 75-80 million years (Crittenden, 1964). The inferred trace of this thrust fault has been extended westward between the east Traverse Mountains and the Little Cottonwood Stock disappearing beneath the Jordan Valley sediments where it is believed to continue to the northwest. It then passes between Antelope and Fremont Islands, eventually connecting with its northern counterpart, the Willard-Paris thrust, east of Ogden (Crittenden, 1964). The identification of this fault explains why the Pennsylvania rocks of the Oquirrh and Traverse Mountains differ from the rocks of the Wasatch east of the Jordan Valley and also explains why there is no visible continuation of the Uinta anticline in the Oquirrhs.

Uinta Arch

The Uinta arch is the largest structural feature within the Wasatch Range (figure 2). It consists of a broad anticline oriented in an east-west direction forming the axis of the Uinta Mountains. East of the Jordan Valley, the anticline is exposed at the mouth of Little Cottonwood Canyon where the axis plunges approximately 30 degrees east.

Basin and Range Faulting

Another significant structural event, block faulting of the Basin and Range, occurred in the late-Tertiary with the Jordan Valley being part of the eastern border. The "Wasatch Fault Zone", which extends along the east side of the valley, separates the Basin and Range from the Wasatch Mountains. This fault zone is part of the Intermountain Seismic Belt, a 100 kilometer (62.5 mile) wide zone of high seismic activity extending from northern Arizona to northwestern Montana (figure 3). Seismic studies indicate the zone is an active rift system with the tensional axis oriented in an east-west direction (Murphy, 1979).

GEOLOGY OF THE JORDAN VALLEY

Marine and Price (1964) divided the Jordan Valley into six ground water districts, three of which are divided into a total of nine subdistricts based on geologic characteristics of driller's logs and well cuttings (figure 1). The following is a brief description of their account of the geologic materials characteristic of each of these areas:

East Bench District

The East Bench district is bounded to the north, south and west by the East Bench fault and to the east by the Wasatch Mountains (figure 1). South of Emigration Creek a pediment extends approximately 1.6 Kilometers (1 mile) west of the range front and is primarily composed of sandstone, limestone and shale of Jurassic and Triassic age. In most areas, this pediment is only a few feet below the surface and is covered by channel sands and gravels. In other areas of the district the sediments consist predominantly of boulders, gravel, sand, silt and clay. The sources of this material are primarily mud rock flows as well as channel, colluvial and flood-plain deposits. Thicknesses of these materials range from less than 1 meter (3.28 feet) in the area of the pediment to as much as 213 meters (700 feet) in the alluvial fans at the mouths of Parleys and Mill Creek Canyons.

East Lake Plain District

The East Lake Plain district is bounded to the east by the East Bench fault, to the west by the Jordan River, to the north by the Salt Lake salient and to the south by an abandoned channel of Big Cottonwood Creek (figure 1). This district is divided into three subdistricts which are as follows:

East Lake Plain Subdistrict

The East Lake Plain subdistrict is composed principally of lake bottom clays with intercalated, discontinuous lenses of gravel. In places, these sediments are modified by recent flood plain deposits of the Jordan River as well as by the broad alluvial fans of City Creek, Emigration Creek, Parleys Creek and Mill Creek. These deposits are underlain at depth by sediments of the Lake Bonneville Group which in turn are underlain by pre-Lake Bonneville deposits. Underlying these unconsolidated sediments are Tertiary limestone or shale. Shale was encountered in a well in Section 12, T. 1 S., R. 1 W at 356 meters (1,168 feet).

City Creek Fan Subdistrict

The City Creek Fan subdistrict sediments are pre-Lake Bonneville alluvial fan material consisting primarily of well-sorted boulders and gravel. The Wasatch formation underlies this subdistrict at a depth of approximately 152 meters (500 feet).

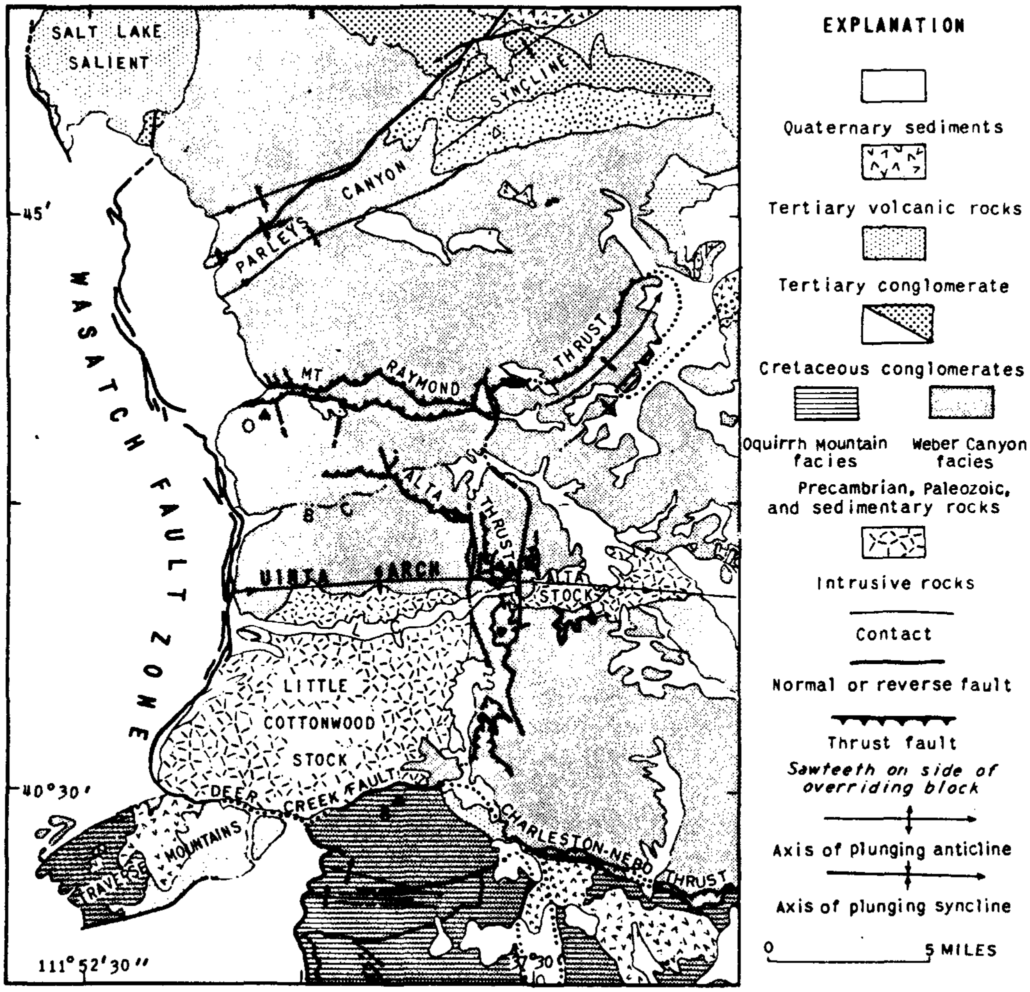

Figure 2. Generalized structure of central Wasatch Range east of Salt Lake City (from Crittenden, 1976, p. 365)

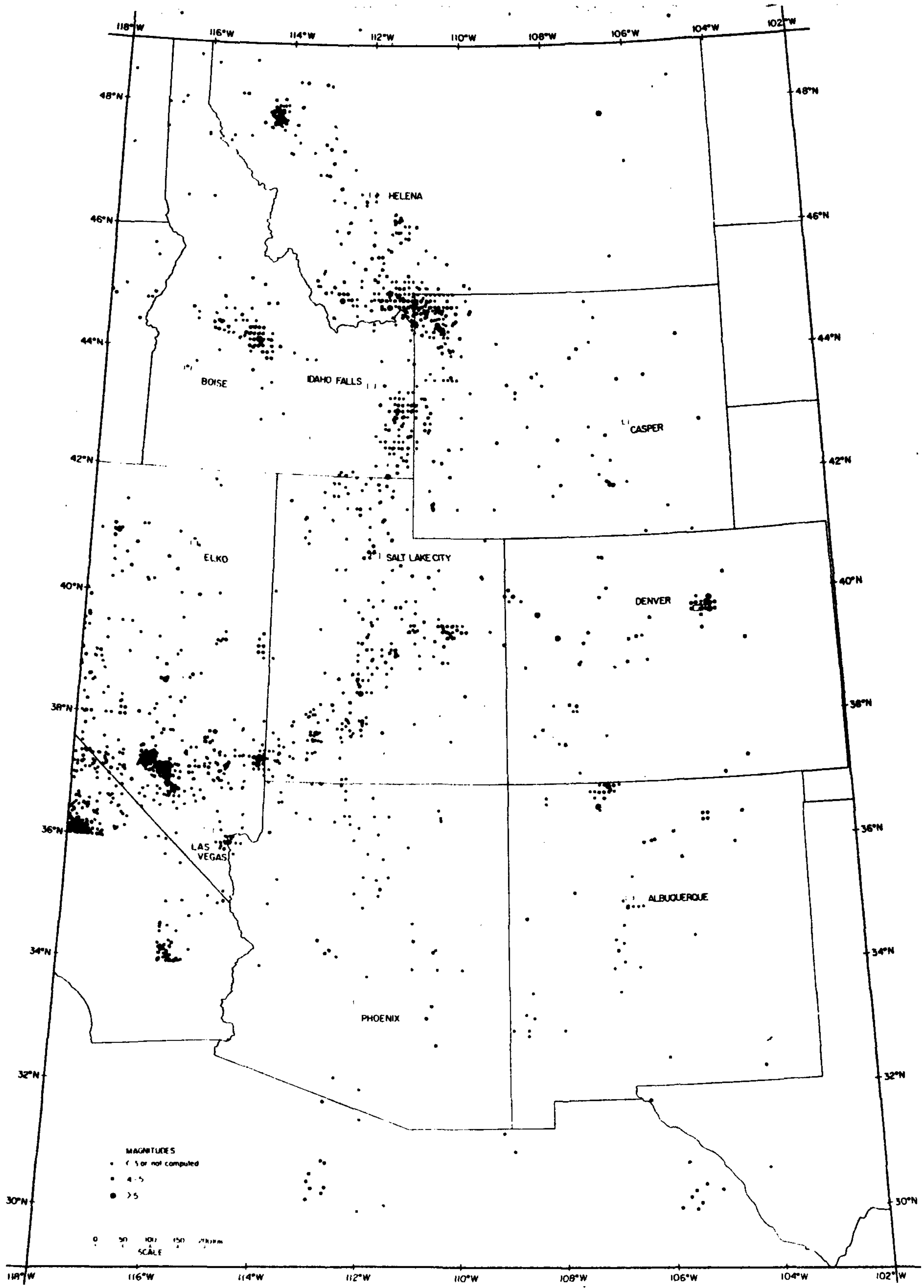

Figure 3. Epicenter map of the Intermountain seismic belt (from Arabasz and others, 1979)

North Bench Subdistrict

The North Bench subdistrcit consists of interfaced pre-Lake Bonneville mud-rock flows, Lake Bonneville deposits and Recent mud rock flows. To the east and south, these deposits grade into the City Creek Fan subdistrict and Lake Plain subdistrict deposits respectively. Generally the deposits consist of boulders, gravel and clay.

Cottonwoods District

The Cottonwoods district is bounded to the north by an abandoned channel of Big Cottonwood Creek and the East Bench fault, to the east by the "Wasatch Fault Zone", to the south by Dry Creek, and to the west by the Jordan River (figure 1). Sediments in the district have been derived from a number of sources, which are as follows: (1) glacial outwash and till, (2) lake deposits, including spits and deltas, and (3) alluvium and colluvium. The sediments primarily consist of gravel and sand. The gravel and sand predominate near the mountain front with the clay increasing toward the Jordan River. Depth to bedrock ranges from a few feet at the Bonneville shore level to more than 3000 feet beneath the Jordan River (Everitt, 1979).

Southeast District

The southeast district is bounded to the north by Dry Creek, to the east by the "Wasatch Fault Zone", to the south by the Traverse Mountains and to the west by the Jordan River (figure 1). A pediment, formed on the Oquirrh formation, extends northwest from the East Traverse Mountains into the valley and is covered by lakeshore sand and gravel. In the Jordan Narrows, gravel and clay have been logged to a depth of approximately 46 meters (150 feet), and are underlain by the Salt Lake Group of Slentz (1955). In most other areas of the district, unconsolidated sediments consist of Lake Bonneville spit sands and gravel and alluvial fan gravel, sand and clays. Depth to bedrock is from less than 305 meters (1000 feet) on the pediment to greater than 710 meters (2000 feet) at the Jordan River (Everitt, 1979).

West Slope District

The West Slope district is bounded to the south by the Traverse Mountains, to the west by the Oquirrh Mountains, to the north by a physiographic break in slope, to the northeast by the Granger fault scarp and to the east by the Jordan River (figure 1). This district includes a broad alluvial-pediment slope formed primarily on rocks of the Salt Lake Formation of Slenz (1955) and has been divided into two subdistricts which are as follows:

North Pediment Subdistrict

The North Pediment subdistrict consists of a thin layer of alluvial or lacustrine deposits overlying the lower units of the Salt Lake Group. Bedrock is considered to range from less than 305 meters (1000 feet) to more than 915 meters (3000 feet) in this subdistrict (Everitt, 1979).

South Fan Subdistrict

Deposits in the South Fan subdistrict consist primarily of gravel, boulders and clay of the Harker's fanglomerate and the Camp Williams units of the Salt Lake Group of Slentz (1955). The two units are, in turn, underlain by the Jordan Narrows unit. Depth to bedrock varies from less than 305 meters (1000 feet) to more than 915 meters (3000 feet) in this area (Everitt, 1979).

Northwest Lake Plain District

The Northwest Lake Plain district is bordered to the east by the Jordan River, to the south by the Oquirrh Mountains and the change in slope caused by the north boundary of the pediment extending east from the Oquirrhs, to the west and northwest by the Great Salt Lake and arbitrarily to the north by the Davis County Line (figure 1). The district is covered by Lake Bonneville bottom deposits. The underlying sediments are such that the district is divided into four subdistricts which are as follows.

Northwest Lake Plain Subdistrict

The Northwest Lake Plain Subdistrict consists of several thousand feet of lake clays with interbedded thin sand lenses. This unit is generally thought to extend to a depth of 700 meters (2300 feet) at which point approximately 152 meters (500 feet) of interbedded sand and andesite could be encountered which in turn is underlain by sand to 1067 meters (3500 feet). The Wasatch formation is thought to be present below this depth.

North Oquirrh Subdistrict

The North Oquirrh subdistrict consists of lake clay and silt, which thins to the south toward the Oquirrh Mountains while thickening to the north to approximately 137 meters (450 feet) at the Great Salt Lake. The clay and silt are underlain by a coarse angular gravel ranging from 46 to 137 meters (150 to 450 feet) in thickness. The gravel is, in turn, underlain by the Oquirrh formation.

South Margin Subdistrict

The South Margin subdistrict is underlain by approximately 30 meters (100 feet) of lake clay which, in turn, is underlain by 61 to 91 meters (200 to 300 feet) of alternating and variable thicknesses of gravel and clay beds which in turn are underlain by Oquirrh formation.

Mid-Jordan Subdistrict

The Mid-Jordan Subdistrict is underlain primarily by flood plain deposits of the Jordan River. Bedrock is known to be at a depth greater than 229 meters (750 feet) in this subdistrict; no known wells have penetrated bedrock in this area.

STRUCTURE OF THE JORDAN VALLEY

Faulting

The Jordan Valley is part of the Wasatch Front Valley physiographic subprovince of the Basin and Range physiographic province. The initial episode of block faulting which resulted in the elongated, parallel, north-south oriented mountain ranges with intervening basins, of which the Jordan Valley is one, occurred in Late Eocene (Eardley, 1955). Eardley has reported a second episode of block faulting which occured in the Pliocene. Also, a number of faults in the surficial valley deposits indicate that faulting has occurred in Recent time although no major earthquakes have been recorded in historical time.

The "Wasatch Fault Zone" (the major zone of recent faulting) in the Jordan Valley separates the Wasatch Range from the valley from Corner Creek, section 3, T. 3 S., R. 1 E., to Mount Olympus, section 14, T. 2 S, R. 1 E., (Plate I). North of this location the recent faulting (East Bench Fault) extends to the northwest (Plate I). Van Horn (1972) has mapped another fault that continues around the base of Mount Olympus and then northwest along the base of the range front, north approximately 6.4 kilometers (4 miles), to section 15, T. 1 S., R. 1 E. (Plate I). Movement on this fault is thought to have occurred more than 5,000 years ago (Van Horn, 1972). North of this point, Van Horn (1969) considered the faulting to have occurred prior to 3,000,000 years ago. The East Bench fault continues to section 33, T. 1 N., R. 1 E. (Van Horn, 1969). An older branch of this fault (pre-5,000 years old according to Van Horn, 1972) continues from section 3, T. 1 S., R. 1 E., northeast to section 33, T. 1 N., R. 1 E., thereby rejoining the younger fault segment. A number of faults have been located in excavations in northern Salt Lake City which indicate a possible continuation of a branch of the East Bench fault northwest, eventually adjoining the Warm Springs fault system located at the base of the Salt Lake salient.

The East Bench fault forms the eastern boundary of a visible inner graben in the Jordan Valley. The western boundary of this inner graben is what Marine and Price (1964) have mapped as the Jordan Valley fault zone, which is approximately 1 mile wide, and includes the Granger fault to the west and the Taylorsville fault to the east (Plate I). This fault zone is oriented northwest-southeast and extends from approximately section 11, T. 2 S., R. 1 W., in the south to section 17, T. 1 S., R. 1 W., to the north. Evidence of other faulting that occurred prior to recent time is apparent in the Jordan Valley to the south and west.

The Traverse Mountains are separated from the valley by a normal fault referred to by Marine and Price (1964) (Plate I) as the Steep Mountain fault. Normal faulting has also been mapped by Slentz (1955) along the base of the Oquirrh Mountains between the Pennsylvanian Oquirrh formation and the Tertiary Harker's fanglomerate from the Traverse Mountains to just south of Bacchus (Plate I). Slentz (1955) reports that in places the fanglomerate is down-faulted to the east and in other areas the fault is buried beneath the fanglomerate.

Tooker and Roberts (1961) have mapped Sevier Orogony thrust faulting at the north end of the Oquirrh Mountains (Plate I). Also, Van Horn (1975) has mapped a number of additional faults beneath the valley sediments based on geophysical investigations. (These fault locations are speculative and have not been included on Plate I).

Geophysical Investigations

Site Specific Gravity Surveys

Detailed gravity surveys were conducted by UGMS on two known low temperature geothermal resource areas in the Jordan Valley; the Warm Springs fault geothermal system and the Crystal Hot Springs geothermal system. The Warm Springs fault geothermal system is located at the western edge of the Salt Lake salient in the northern end of the valley (figure 4). The gravity survey consisted of 12 east-west oriented gravity lines with individual station spacings of 152 to 304 meters (500 to 1000 feet). Individual gravity lines were spaced from 0.4 to 1.2 km (0.25 to 0.75 miles) apart. One gravity profile was modeled using a three dimensional modeling program. The modeling indicates, from east to west, two faults, a deep alluvium-filled graben and a horst block; the easternmost fault corresponds to the Warm Springs fault. The model indicates the downthrown block of the Warm Spring fault to the west is covered by approximately 100 meters (328 feet) of alluvium (Murphy & Gwynn, 1979). The Hobo Springs fault (the second fault to the west) is also downthrown to the west and borders the aforementioned graben. This graben has an estimated depth of 1220 meters (4000 feet).

The Crystal Hot Springs geothermal system is located in the southern part of the Jordan Valley, southwest of the town of Draper (figure 4). An areawide gravity survey was conducted by orienting profiles perpendicular to the East Traverse and Wasatch Mountain ranges. Profiles were spaced at nearly 0.8 kilometer (0.5 mile) intervals with approximately 304 meter (1000 feet) intervals between individual stations. The area-wide survey provided a regional setting on which to base a more detailed gravity grid to better delineate the structure beneath the springs (Murphy, 1981). The detailed grid consisted of 290 gravity stations, spaced 350 feet apart and was centered on the thermal springs.

The regional gravity surface resulting from the area-wide gravity survey indicates normal range-front fault segments bordering the west and north edges of the Wasatch and Traverse Mountain ranges respectively (Murphy, 1981). Murphy (1981) states that in the vicinity of the thermal springs, these faults trend almost east-west and abruptly terminate a gravity high to the south. In other areas, the presence of northeast trending faults is indicated.

Modeling of the data suggests Crystal Hot Springs is located between two range front faults striking roughly east-northeast, and dipping to the northwest (Murphy, 1981). Drill hole data has indicated a third range front fault to the northwest. Murphy (1981) also points out that the structure between the southernmost two range front faults is quite complex, consisting of a number of small, tilted fault blocks.

The detailed gravity surveys indicate that the Jordan Valley is very complex structurally, consisting of smaller scale bedrock horsts and grabens beneath unconsolidated valley sediments within the valley-wide graben. Work by Everitt (1979) and Arnow and Mattick (1968) also indicate a complex graben system within the Jordan Valley.

Valley-Wide Gravity Surveys

The structural complexity of the Jordan Valley initiated a gravity survey over the entire study area which consisted of 800 stations along 40 profiles at 0.4 to 0.8 kilometer (0.25 to 0.5 mile) intervals. This survey was designed to compliment the two site specific surveys and the work previously done by Cook and Berg (1961). The result of incorporating this data with the work previously done can be seen in the "Complete Bouguer Gravity Map" presented as Plate II. This map indicates that a number of major bedrock fault blocks may be present in the Jordan Valley. This could be significant because the borders of these aforementioned fault blocks could be conduits for geothermal water. Plate II indicates significant gravity lows in the Jordan Valley which could correspond to structural grabens whereas the

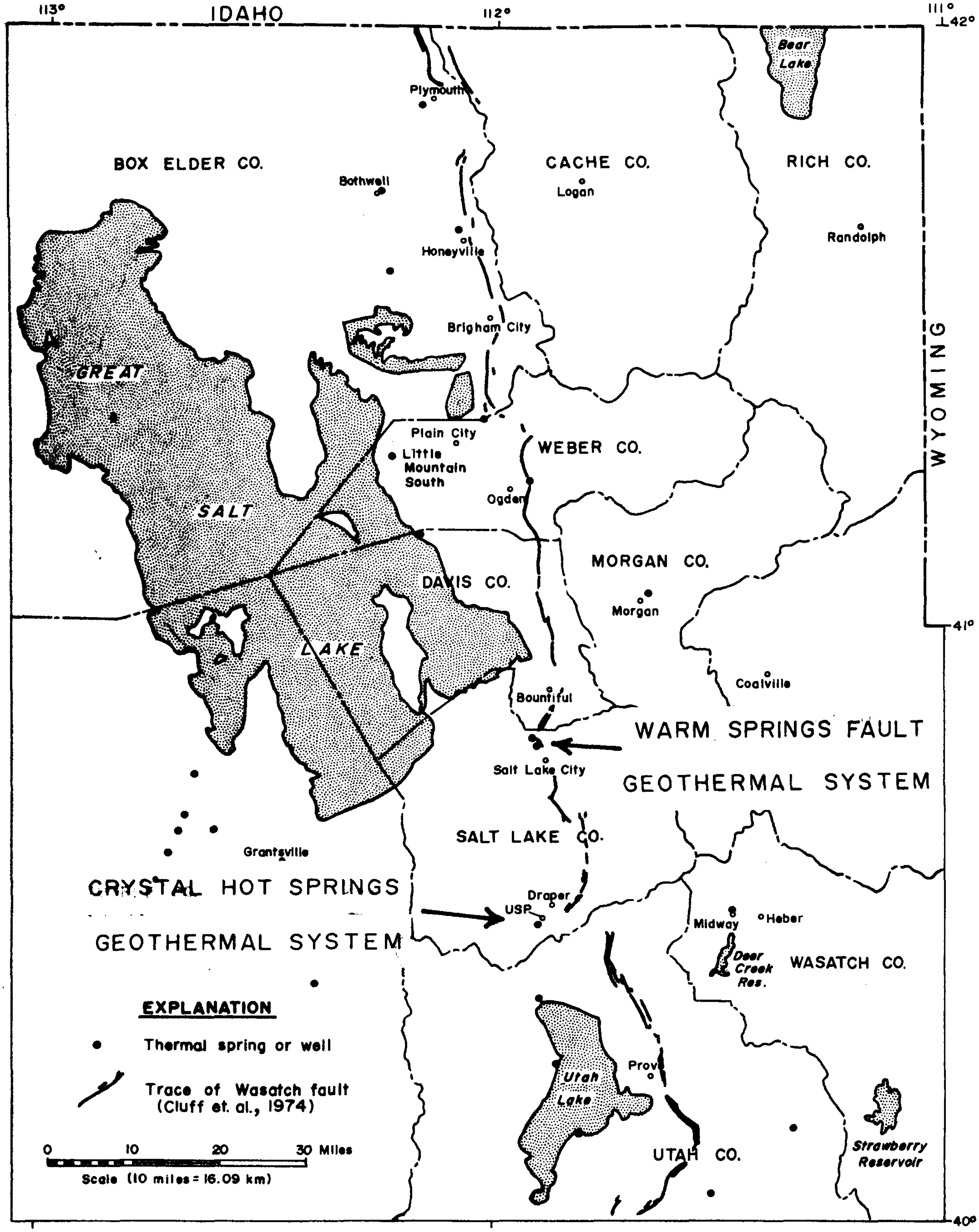

Figure 4. Location of the Warm Springs fault and Crystal Hot Springs geothermal systems, Salt Lake County, Utah

intervening gravity highs could indicate structural bedrock horsts.

Aero Magnetic Surveys

An aeromagnetic survey, 9.5 miles in length (north-south), 6 miles in width (east-west) and centered on Crystal Hot Springs, was flown to detail the complex magnetic surface of the area. Smith (1980) concluded that the resulting magnetic anomaly results from a series of magnetically susceptible intrusive and extrusive bodies that trend east-northeast and vary in depth from 2887 meters (7500 feet) to within 107 meters (350 feet) of the surface. The lower portions of the bodies are thought to be intrusive while the uppper levels may be either intrusive dikes and sills or extrusive flows (Murphy, 1981). Smith (1980) noted that many of the stacked prisms used to model the intrusives shared common edges which could indicate the presence of deep seated structures. One of these deep seated structures is present just north of the thermal springs and may be coincident with faults delineated on the basis of gravity data (Murphy, 1981). Results of this survey and preliminary modeling by Smith (1980) provides an understanding of the distribution of the magnetic susceptibility in the subsurface and a major normal range front fault north of the springs (Murphy, 1981).

GROUND WATER

Ground water in the Jordan Valley occurs in: (1) a large artesian aquifer, (2) a deep unconfined aquifer, (3) a shallow unconfined aquifer overlying the (artesian) confined aquifer, and (4) in local, perched unconfined aquifers. All are hydraulically interconnected to some extent, but the large artesian aquifer directly recharges the deep artesian aquifer forming the principal groundwater reservoir (Hely and others, 1971). The shallow unconfined aquifer overlies the confining layer for the artesian aquifer while the locally perched aquifers are in areas overlying the deep unconfined reservoir. This confining layer generally consists of clay, silt and fine sand, varying in thickness from 12 to 30 meters (40 to 100 feet), and lying between 15 and 46 meters (50 and 150 feet) below the surface (Hely and others, 1971). The shallow, unconfined aquifer extends over the same area as the confined aquifer while the perched aquifers are found primarily east of Midvale and west of Riverton overlying the deep unconfined aquifer (figure 5).

Principal Aquifer

The deep unconfined aquifer in the Jordan Valley is a principal recharge source for the artesian aquifer. The line dividing these two aquifers can only be approximately located due to shifts caused by response to changing rates of recharge and discharge (Hely and others, 1971) (figure 5).

The artesian aquifer consists of quaternary deposits of interbedded clay, silt, sand and gravel, all hydraulically interconnected; thin beds and lenses of fine-grained material up to 20 feet thick tend to confine water in each of the many individual beds of sand and gravel. The fine-grained material is slightly to moderately permeable and discontinuous, thereby allowing movement of water between the various permeable beds of sand and gravel (Hely and others, 1971). This confined aquifer attains a maximum thickness of more than 305 meters (1000 feet) in the northern part of the valley (Hely and others, 1971). For the most part his aquifer is underlain by Tertiary and pre-Tertiary deposits. In some areas the Tertiary deposits are permiable enough to yield water to wells (Hely and others, 1971).

Recharge and Movement

Recharge to the Jordan Valley ground water system comes from the following sources: (1) seepage from bedrock fractures in the adjoining mountains, (2) underflow in channel fill draining the adjacent canyons, (3) underflow from Utah Valley to the south through the Jordan Narrows and Ogden Valley to the northwest of the Salt Lake salient, (4) seepage from creek channels and the Jordan River, (5) seepage from unlined canals, (6) migration upward through fault systems, (7) direct precipitation, (8) seepage from irrigation and (9) seepage from tailings ponds.

Groundwater movement in the principal aquifer is generally northward toward the Great Salt Lake. Groundwater migrates laterally toward the Jordan River from both the east and west sides of the valley and subsequently migrates to the north (figure 5).

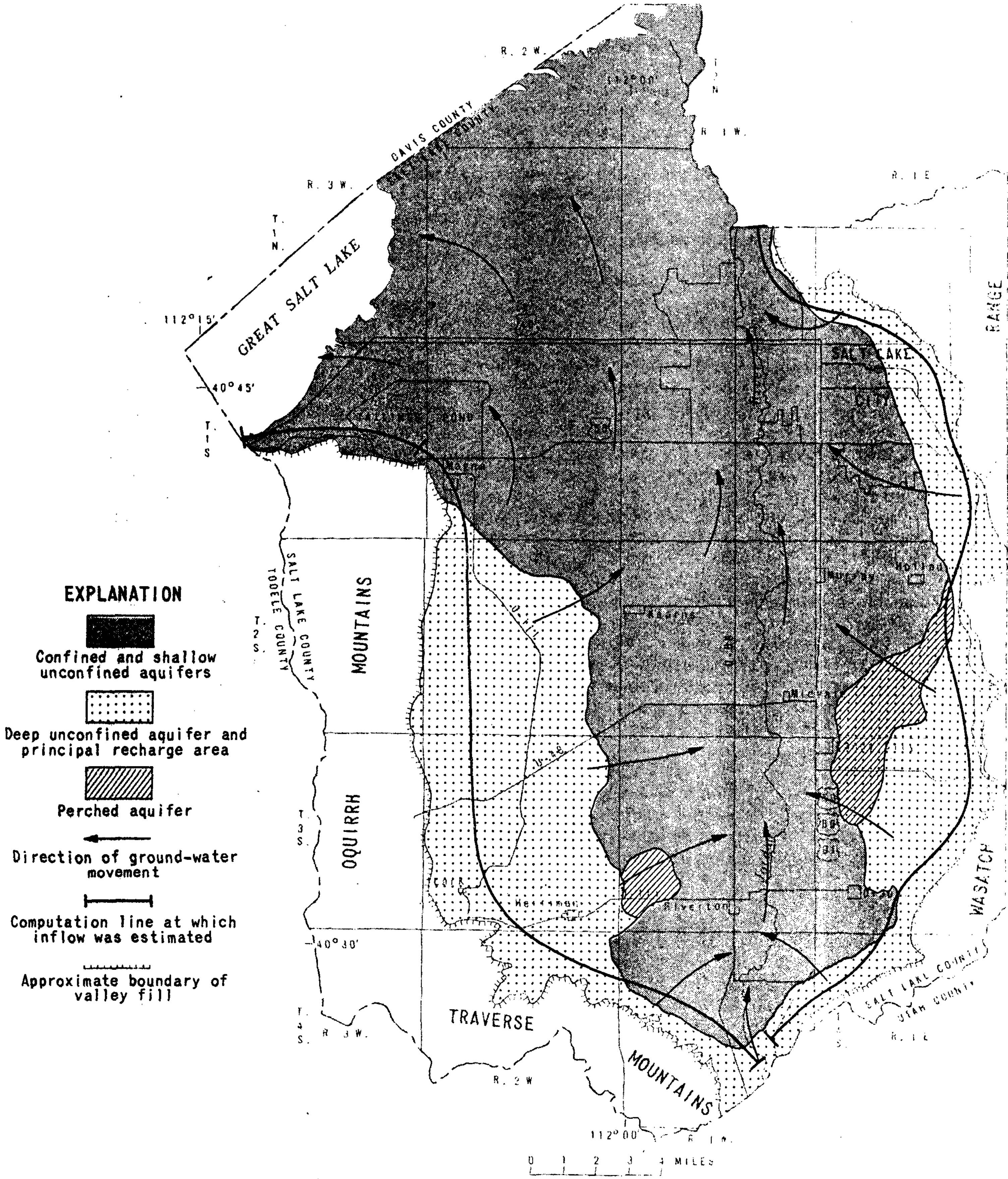

Figure 5. Approximate areas in which ground water occurs in confined, shallow unconfined, deep unconfined, and perched aquifers in Jordan Valley (from Hely and others, 1971)

An attempt was made to measure/sample wells that intercepted the principal aquifer. Where no wells of this type were available or accessible, shallower wells were used although these are few in number. Temperatures were recorded at 214 locations by Utah Geological and Mineral Survey (UGMS) personnel, an additional 9 temperatures were obtained from local municipal water departments, and 15 temperatures were provided by the Kennecott Copper Corporation. Of the 214 locations measured by UGMS, 5 were springs and the balance consisted of pumped or flowing wells.

Temperatures, both measured and acquired, range from 7.5° to 85°C; 182 of the 238 total measured range from 10.4° to 17.4°C. These temperatures are slightly higher than those of Marine and Price (1964) who found most temperatures between 7.8° and 59.4°C. This result is not unexpected since this study was designed to define potential geothermal areas and much data was collected in areas thought to be of above average temperature. Also, test wells have been drilled at Crystal Hot Springs, intercepting warmer water at depth, thereby increasing the upper limit to 85°C from the 59.4°C that had previously been measured in the surface ponds at this location.

Areas of Warm Water

Seventy-six percent (182 out of 238) of all temperatures measured or acquired in the Jordan Valley were 17.4°C or lower. For this reason, 18.0°C was designated as the low temperature limit in trying to delineate anomalously warm areas for further investigation. The result indicates six general areas of potential low temperature geothermal water in addition to a few isolated wells. These general areas are: (1) the northcentral valley area, (2) the area immediately north of the Oquirrh Mountains, (3) the Warm Springs fault geothermal area, (4) an east-west section of the central valley in the vicinity of Kearns, Murray and Holladay, (5) an area between Sandy City and Draper in the southeastern part of the valley, and (6) an area in the extreme southern part of the valley, including Crystal Hot Springs. These areas are presented in Plate III.

Northcentral Valley Area

Temperatures in this vicinity range from 19.3° to 28.1°C and are spread over a fan-shaped area covering approximately 85 square kilometers (33 sw. mi.). First indications from gravity surveys suggest this area is bordered by faulting which could, in turn, structurally control the location of warm water encountered in wells. If this is occurring, groundwater must be circulating to a minimum depth of 0.5 kilometers (0.3 miles) for the Basin and Range geothermal gradient to heat the water to the temperature recorded at the well heads. Marine and Price (1964) suggest an alternative theory of ground water reactions with the organic clays in the area, with the resulting temperature being a product of exothermic reactions.

North Oquirrh Area

This area, located immediately north of the Oquirrh Mountains, ranged in temperature from 21° to 29°C (Plate III). A possible source of this anomaly could be water circulating at depth and migrating up the Pony Express and Rio Grande thrust faults. The minimum depth of circulation is estimated to be 0.5 kilomters (0.3 miles).

Warm Springs Fault Area

The Warm Spring Fault geothermal system is located immediately west of the Salt Lake salient (Plate III). According to Murphy and Gwynn (1979) this system is controlled by water circulating to a minimum depth of 1.5 to 2.0 kilometers (0.9 to 1.2 miles) and migrating up the Warm Springs and Hobo fault systems. They indicate that the major springs tend to occur at the intersections of these major faults with older, minor structures striking roughly perpendicular. An anomalously warm well (19.4°C) is located approximately 1.6 kilometers (1 mile) south of the Warm Springs fault. This anomaly could result from a continuation of the Warm Springs or Hobo fault systems to the south.

Central Valley Area

A temperature of 18.8°C was recorded approximately 3.2 kilometers (2 miles) north of Kearns while two warm temperatures (18.5° and 21.0°C) were recorded approximately 2.4 kilometers (1.5 miles) to the southwest. The warm temperature recorded to the north could be controlled by the Jordan Valley fault zone. The two wells to the southwest were drilled to depths of between 305 and 335 meters (1000 to 1100 feet) in an area where wells are known to be receiving water from permeable zones in Tertiary deposits. These wells were drilled to a significantly greater depth than other, cooler wells in the area therefore indicating that the normal geothermal gradient may be producing these anomalously warm temperatures.

A second group of warm temperatures were measured in wells located in the Murray area (Plate III). Six wells produced temperatures ranging from 18° to 21°C.

Two wells with temperatures of 23.8° and 22.1°C are located in the Holladay area (Plate III). This anomaly may be attributable to warm water rising from depths of at least 0.5 kilometers (0.3 miles) along the East Bench fault.

Sandy City – Draper Area

Warm water was encountered in 9 wells located in the general area between Sandy City and Draper, Utah (Plate III). Six of these wells are located south of Sandy City and trend in a west-northwest direction with temperatures ranging from 18.0° to 48°C. The anomalous temperatures in two of these wells can be attributed to the normal thermal gradient.

Three other warm wells are located randomly in and northwest of Draper (Plate III). These wells produce water with temperatures of 19.2°, 21.7° and 23.7°C. Insufficient data is available at this time to speculate as to the source of this water.

Crystal Hot Springs Area

Five warm temperatures ranging from 28.5° to 85.0°C were measured at and in the vicinity of Crystal Hot Springs. Murphy and Gwynn (1979) and Murphy (1981) studied this area extensively and conclude that it is located between two range front faults, is underlain by smaller fault blocks and is supplied by warm water circulating to a minimum depth of 2.5 kilometers (1.55 miles). This water is heated by the normal geothermal gradient and rises along permeable fault zones and infiltrates into well-fractured quartzite located beneath the site.

Other Isolated Warm Temperatures

A temperature of 23.5°C was measured approximately 2.4 kilometers (1.5 miles) north of Sandy City. This temperature cannot be accounted for by the normal Basin and Range geothermal gradient and does not seem to be related to other anomalous temperatures in the area.

Three other anomalous temperatures were measured in the southwest part of the valley (Plate III). Locations of these wells were approximately 3.2 kilometers (2 miles) southeast of Crystal Hot Springs, just east of Herriman and in Rose Canyon (Plate III). The 18.6°C temperature recorded southwest of Crystal Hot Springs and the 21.0°C temperature measured in Rose Canyon can be accounted for by the normal geothermal gradient. At this time no explanation is available for the 19.0°C temperature recorded east of Herriman.

Chemistry

The Jordan Valley exhibits a complex ground water chemistry which is attributable to complicated stratigraphy and structure as well as low temperature geothermal activity within the area. Despite this inherent complexity, however, trends are apparent in this system which are indicative of anomalous chemistry that may be associated with geothermal activity.

A definite trend in the ground water seems evident from the Jordan Narrows north to Sandy City (Plate IV). Generally, the water in this area contains approximately equal amounts of sodium, calcium and magnesium cations with predominantly sulfate and chloride anions.

Four isolated locations are indicative of ground water high in sodium and chloride (Plate V). Three of these locations are in the area where Corner Creek, Little Cottonwood Creek and Big Cottonwood Creek plumes should predominate. However, due to the fairly deep graben present in this area (Everitt, 1979), the ground water is thought to circulate quite deep, increasing in temperature and dissolution time and consequently accumulating predominantly sodium and chloride ions. The fourth location is on the edge of this basin but exhibits similar chemistry. This could be caused by the deep circulation of geothermal ground water through a highly faulted area (Murphy and Gwynn, 1979).

The central part of the Jordan Valley displays chemistry typical of ground water in a resistate system. The groundwater contains approximately equal amounts of sodium, calcium and magnesium cations with predominantly sulfate and chloride anions (Plate IV). Total dissolved solids are greater than 1600 ppm (Plate V). These chemical conditions may be the result of the structural conditions within this portion of the valley; a possible explanation is as follows: The "Complete Bouguer Gravity Map" (Plate II) suggests this area is structurally complex. Plate II indicates that an east-west striking normal fault downdropped on the south is present, extending east from Bacchus approximately 6.4 kilometers (4 miles). This fault is truncated by a north-south oriented fault, downdropped on the east, which extends northward through the valley (Plate II). The increase in bedrock elevation to the north through this part of the Jordan Valley causes a decrease in the rate of groundwater migration and an increase in time of dissolution. East of Bacchus, the groundwater could be interacting with geothermal water migrating upward from depth along the fault system, thereby increasing in total dissolved solids.

Further north, a sharp transitional boundary exists, where the ground water changes to a predominantly sodium chloride system (Plate IV). This transition zone roughly parallels the stratigraphic change from the pediment located in the West Slope groundwater district to the Lake Bonneville clays of the Northwest Lake Plain groundwater district. The change in chemistry reflects the influence of the Great Salt Lake and the ion exchange capacity of the lake clays located in this area.

The chemistry of the groundwater in the north-central part of the valley is predominantly a sodium chloride system, but displays a significant increase in bicarbonate while exhibiting a decrease in calcium, magnesium, sulfate and total dissolved solids (Plates IV and V). Plate II indicates this area is a horst bounded by grabens to the north, south and west. A seismic reflection survey conducted by the U. S. G. S. also shows this area to be a bedrock high (Mower, 1968). According to Hely and others (1971), the low - total dissolved solids is caused by the migration of high quality water from the Ogden Valley to the north. The significance of the chemistry in relation to the warm water measured in this area is not understood at this time.

The area east of the Jordan River, extending from the Salt Lake salient south to Corner Canyon is recharged by high quality water from the canyons in the Wasatch Range. This results in low total dissolved solids and no significant anamalous water chemistry (Plates IV and V). This high quality water could mask any obvious chemical indication of possible geothermal activity in this area of the valley. The temperature map, however, does indicate warm temperatures both east and west of Murray (Plate III).

SUMMARY OF FINDINGS

Prior to the study of area-wide, low-temperature geothermal resources in the Jordan Valley, Murphy and Gwynn (1979) conducted studies of two known geothermal areas: (1) the Warm Springs fault, and (2) Crystal Hot Springs. These studies indicate meteoric water is being circulated to depth and heated by the ambient temperature derived from normal heat flow. This warm water migrates upward along permeable fault zones. Study of these two geothermal areas has proven important to the present investigation by providing insight into the controlling mechanism of low-temperature geothermal resources.

The present study being conducted in the Jordan Valley has provided a complete Bouguer gravity map in addition to water temperature and chemistry for 238 wells and springs (Plates II, III, IV, and V). Results of the gravity survey indicate a number of fault blocks are buried beneath valley sediments (Plate II). Since previous work has indicated that warm water migrates upward along permeable fault zones, the common borders of these horsts and grabens could provide conduits for warm water.

Water temperatures measured in the valley have provided further insight into areas of geothermal potential (Plate III). Four areas, in addition to the two known geothermal sites have warm water wells. These areas are designated as the: (1) Northcentral Valley area, (2) North Oquirrh area, (3) Central Valley area, and (4) Sandy City - Draper area.

The chemistry of the Jordan Valley is quite complex and not fully understood in regard to geothermal potential. Further investigation is required to discern the significance of the anomalies noted.

The two known geothermal areas are located where bedrock is within 305 meters (1000 feet) of the surface. In other areas of the valley, bedrock is covered by hundreds of meters (thousands of feet) of unconsolidated sediment which could conceal warm water anomalies due to lateral dispersion or dilution within the principal aquifer. Areas of low permeability could retard warm water flow, thereby allowing for cooling of the water prior to discharge in wells or springs. The area of the valley east of the Jordan River receives considerable cool, high quality ground water recharge from snow melt in the Wasatch Range. This water could appreciably cool possible warm water as well as notably improve the water quality of geothermal resources in that area.

FUTURE INVESTIGATIONS

Jordan Valley geothermal assessment has provided insight into the direction of future investigations to complete the study. The following is additional work to be undertaken within the following year.

(1) Gravity modeling to define the structure and geology in specific areas indicative of geothermal potential.
(2) Testing of geothermometry models to determine possible reservoir temperatures in areas with warm water wells.
(3) Gradient logging where accessible throughout the valley.
(4) Further investigation of the chemistry to determine, if possible, correlations between warm water and ion concentrations or types.

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Plate I

Sources of information:

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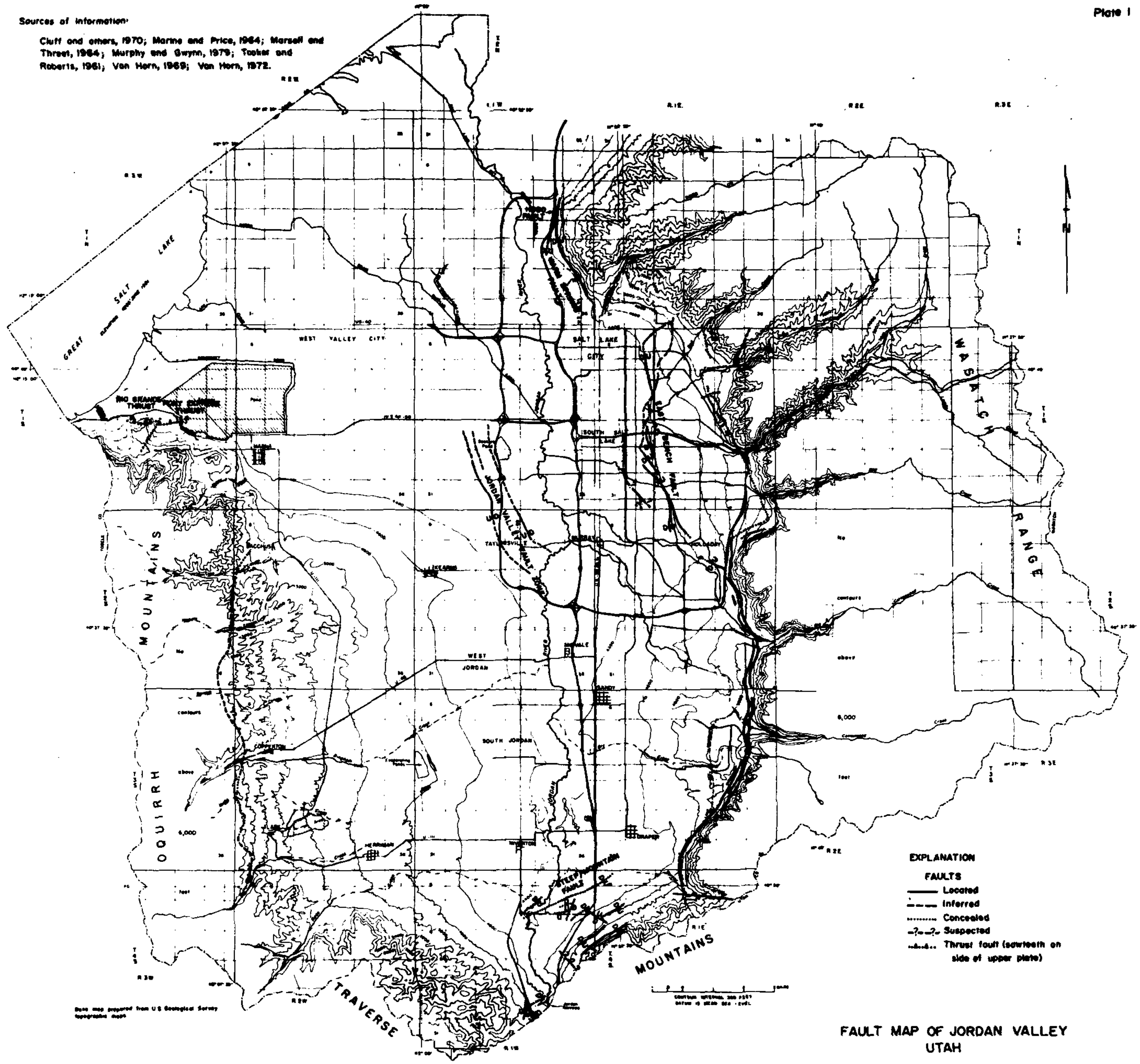
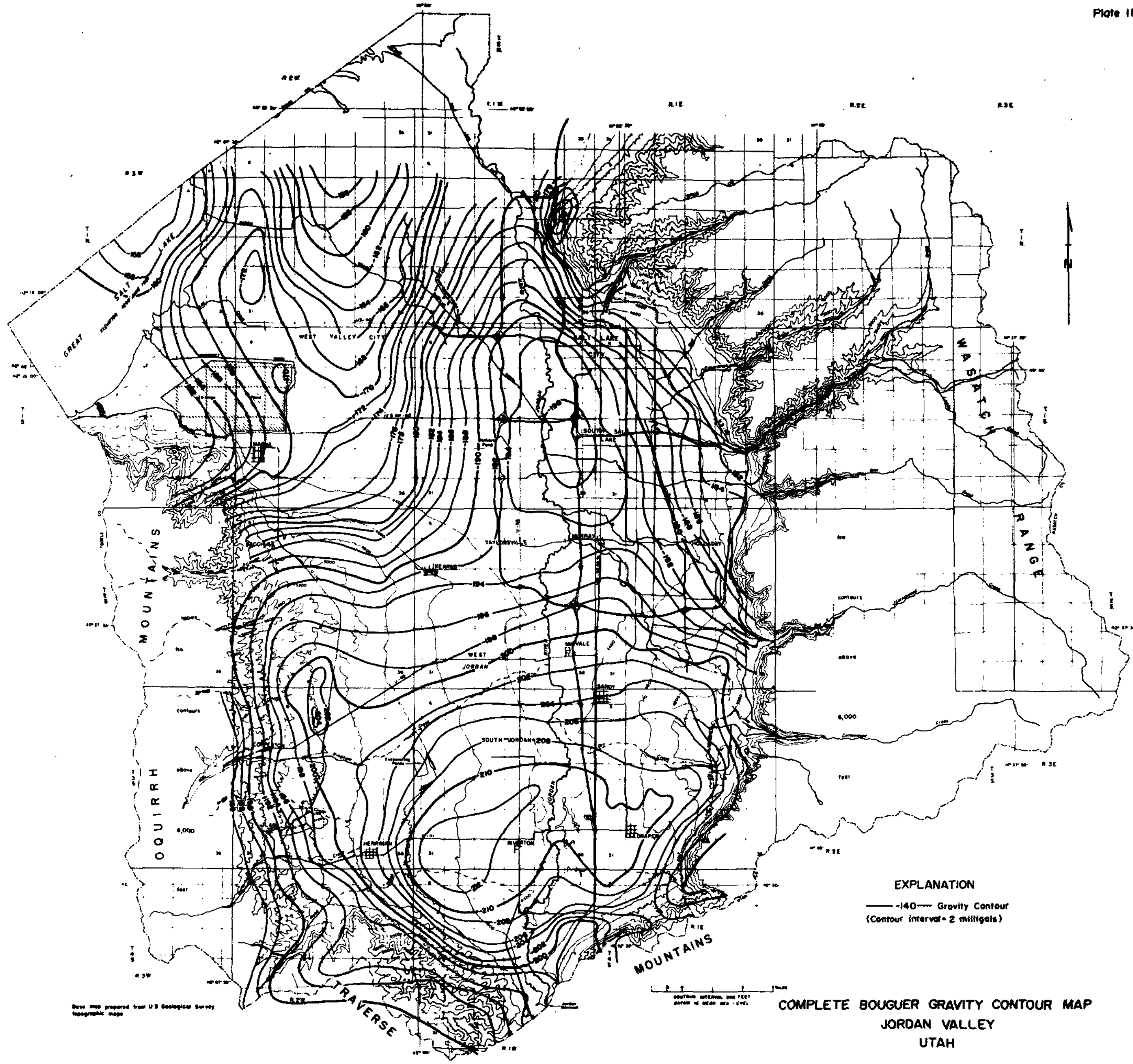
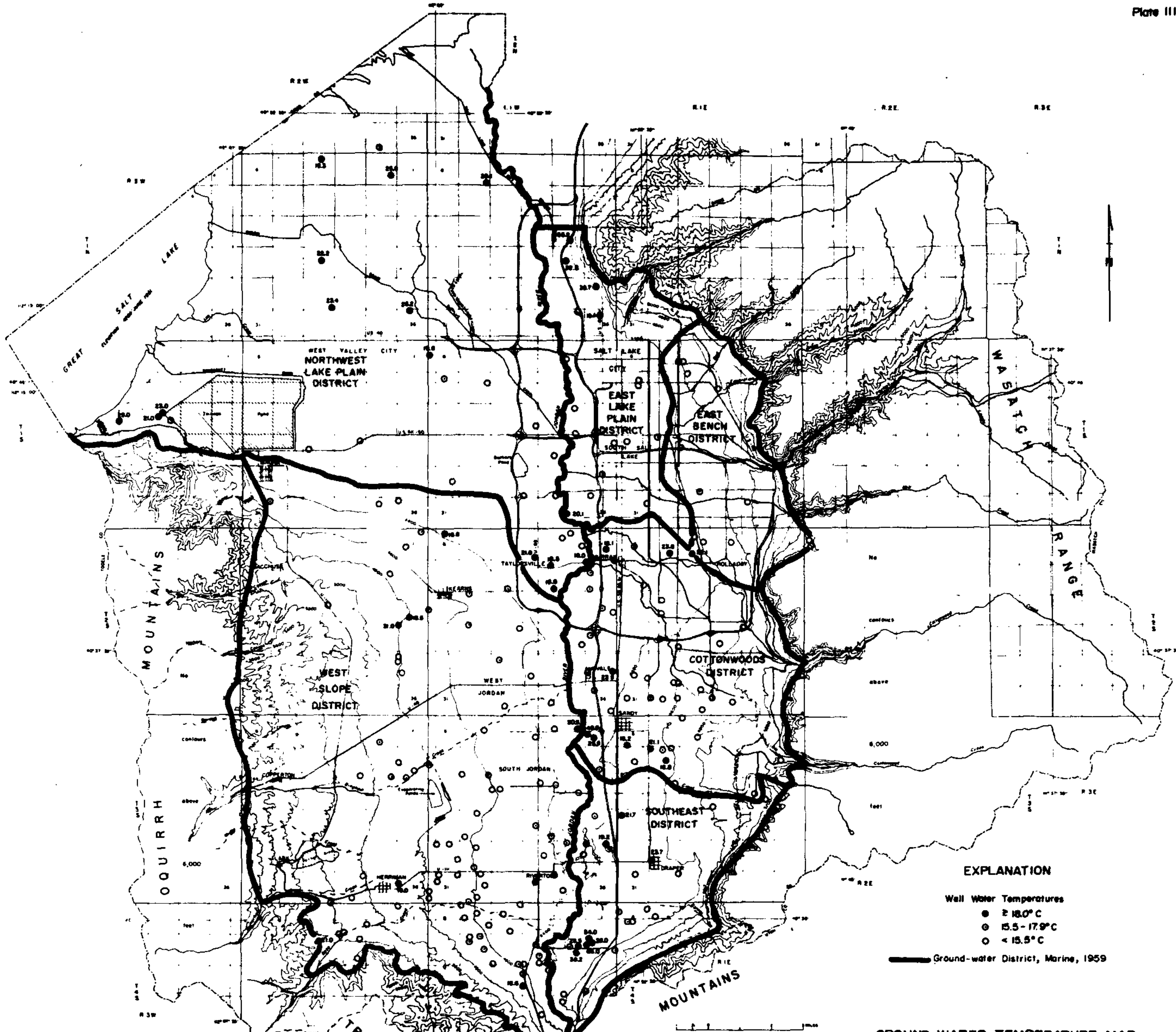


FAULT MAP OF JORDAN VALLEY
UTAH

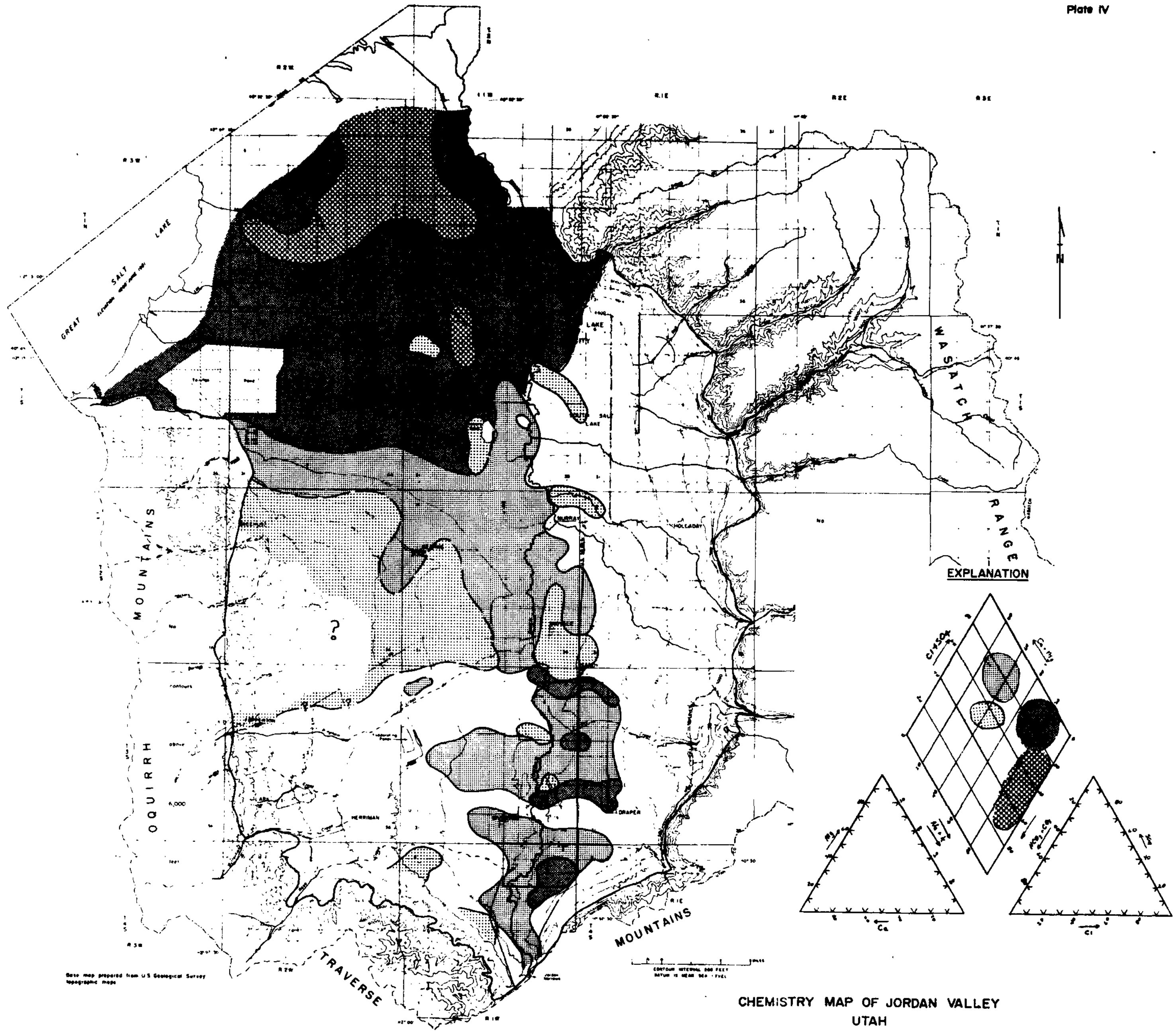


COMPLETE BOUGUER GRAVITY CONTOUR MAP
JORDAN VALLEY
UTAH

Plate III

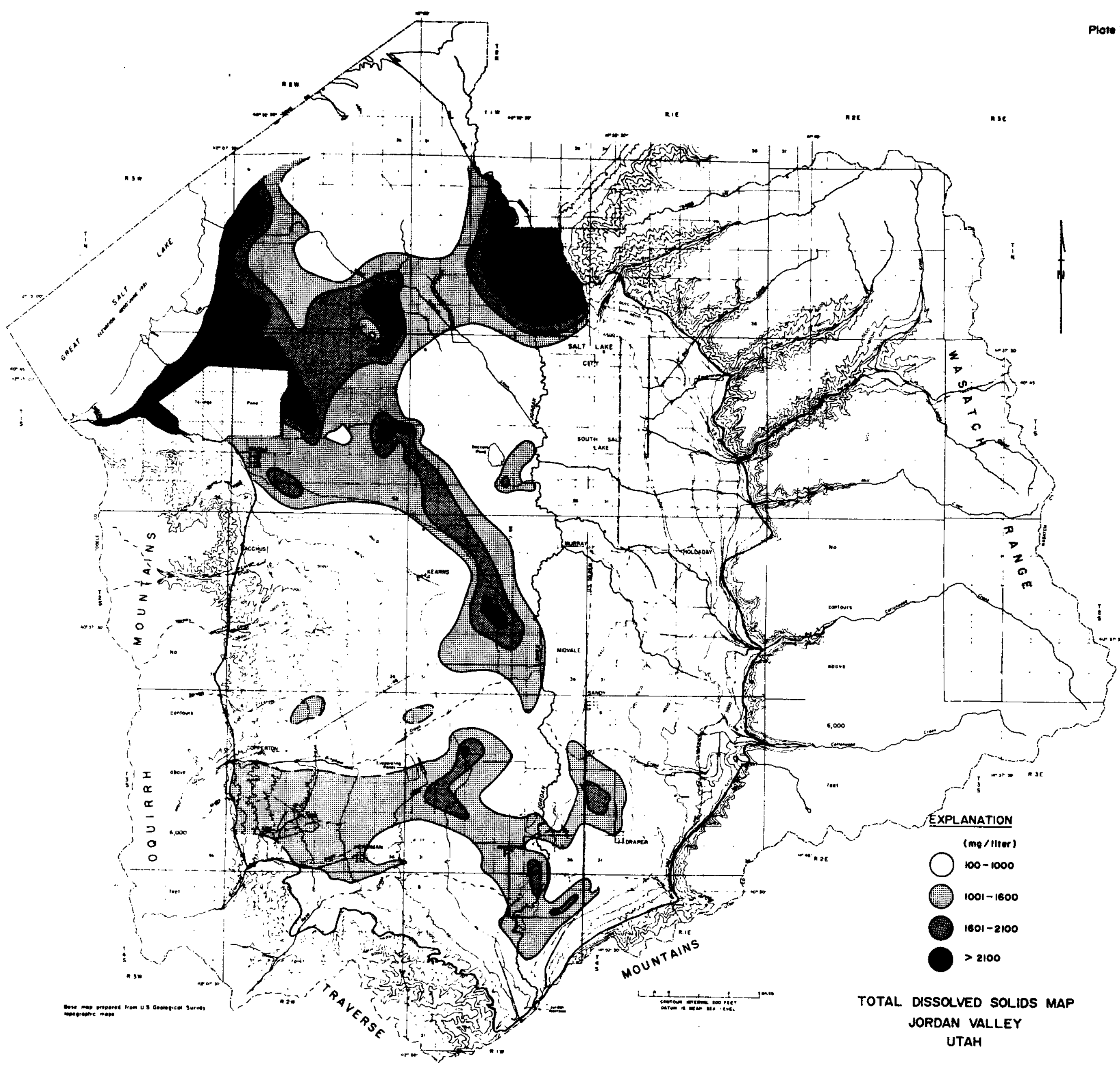


GROUND-WATER TEMPERATURE MAP
JORDAN VALLEY
UTAH



CHEMISTRY MAP OF JORDAN VALLEY
UTAH

Plate V



TOTAL DISSOLVED SOLIDS MAP
JORDAN VALLEY
UTAH

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GEOTHERMAL ASSESSMENT ACTIVITIES IN OREGON, 1979-1980, AND A CASE STUDY EXAMPLE AT POWELL BUTTES, OREGON

George R. Priest (Oregon Department of Geology and Mineral Industries)
Gerald L. Black (Oregon Department of Geology and Mineral Industries)
David D. Blackwell (Southern Methodist University)
David E. Brown (Oregon Department of Geology and Mineral Industries)

The Oregon Department of Geology and Mineral Industries (DOGAMI) has, for the last two years, been involved in two major geothermal resource studies. One project concerned regional assessment of the northern and central Cascade Range, while the other was aimed at site-specific assessment of various areas for direct-use (low- to moderate-temperature) geothermal resources. The Cascade Range study was recently expanded to include the southern Cascade Range of Oregon. Study areas for both assessment programs are outlined on Figure 1.

The Cascades project consisted of two major investigations. The Mount Hood project entailed detailed resource assessment around Mount Hood including drilling of a 1,837 m well at Old Maid Flat. Most of the Department's effort was directed at the other major investigation which involved collection and interpretation of regional gravity, aeromagnetics, heat flow, water chemistry, and geologic data on the Cascades. Dr. Richard Couch of Oregon State University produced an aeromagnetic map and free-air, complete Bouguer, and residual gravity maps of the range to nearly complete his coverage from previous studies. Only the aeromagnetic map set is not complete, owing to lack of data in the northern Cascades. Heat-flow analysis involved drilling 22 150 m temperature gradient holes and scrounging for additional water well data. The geologic studies included paleomagnetic studies (James Magill and Allan Cox of Stanford University), regional lineament analysis (subcontracted to specialists), and local detailed mapping by G.R. Priest and assistants in areas with special tectonic significance.

During the low- to moderate-temperature assessment project, spring chemistry, heat flow, geophysics, geology, and water chemistry data were compiled for twelve areas in Oregon (Figure 1). Nine of these areas have been selected for extensive detailed assessment, primarily on the basis of their proximity to population centers, and thus greater probability for utilization (Figure 1). A literature search was conducted for each of the areas, and temperature gradients and water compositions have been measured in most of the twelve areas (Brown, Black, and McLean, 1980a, b; Brown, McLean, and Black, 1980a, b, c: Brown, McLean, Priest, Woller, and Black, 1980; Brown, McLean, Woller, and Black, 1980; Peterson, Brown, and McLean, 1980).

Detailed assessment of the nine priority areas involves more extensive water sampling and temperature gradient measurement than in the other areas. In some cases (e.g., Corbett-Moffett, Parkdale, Harney Basin, La Grande, Vale-Ontario, Lakeview, Belknap-Foley, and Willamette Pass), original geologic maps are being produced to aid in interpretation of the heat flow and water data. In the highest priority areas, temperature gradient wells have been or will be drilled (e.g., Corbett-Moffett, Powell Buttes, Harney Basin, Lakeview, La Grande, and Willamette Pass). The most extensive drilling was conducted at Powell Buttes, which the Department identified as a "blind" geothermal anomaly in 1978. The investigation of the Powell Buttes area is a good example of the Department's geothermal exploration strategy and will be discussed in some detail in a later report by Gerald Black of DOGAMI.

Rumors of warm wells in the Powell Buttes area caught the attention of the Department in 1977, and preliminary temperature gradient measurements in 1978 indicated that the area had abnormally high temperatures at shallow depths. Subsequent gradient measurements in 1979 and drilling of eight 150 m temperature gradient holes in 1980 revealed a major temperature gradient anomaly with gradients as high as 164°C/km (Figure 2). A 461 m well was then drilled into the anomaly in 1981 to test the temperature at depth (PB-1, Figures 2 and 3). It must be emphasized that no drilling was done in the area until the project geologist was sure that he had very good geologic maps of the area, and all available temperature gradients from water wells.

All data from the project were then carefully analyzed during the spring of 1981. It became clear that several models could explain the elongate gradient anomaly on the southwest side of the buttes:

1. Forced convection along a fault or fracture system.

2. Lateral flow of warm water at shallow depths.

3. Juxtaposition of rocks of strongly contrasting conductivity. In this third case, the gradient anomaly would then not imply a hydrothermal convection system.

Careful measurement of thermal conductivity of drill cuttings and cores allowed detailed heat flow modeling of the temperature data. This technique put important quantitative constraints on the proposed models. Although heat flow computations are not yet complete, it appears that either model 2 or 3 or some combination of them could explain the anomaly. The temperature profile in PB-1 (Figure 4) shows a sharp break in slope at the contact between the Oligocene basement rocks (Clarno? Formation) and the highly porous Mio-Pliocene Deschutes Formation. This break in slope could be caused by very slowly moving waters within the lower part of the younger rocks (Brown, Black, and McLean, 1980b), or by the low measured conductivity of the dry upper part of the porous Mio-Pliocene sequence, or some combination of these factors. In any case, geothermetric estimates of the reservoir temperatures of warm (30° to 40°C) shallow water in the thermal anomaly indicate that original water temperatures were similar to measured temperatures (Brown, Black, and McLean, 1980b). This further suggests that the warmth of these waters is obtained from shallow conductive heating rather than deep convection. It is possible, however, that the shallow thermal water has moved so slowly from depth that it reequilibrated with lower-temperature, near-surface rocks, thus masking effects of a much higher temperature reservoir.

Figure 5 illustrates how isotherms would be generated through the Powell Buttes area by assuming a deep heat flow of 3.0 H.F.U. (considered normal for

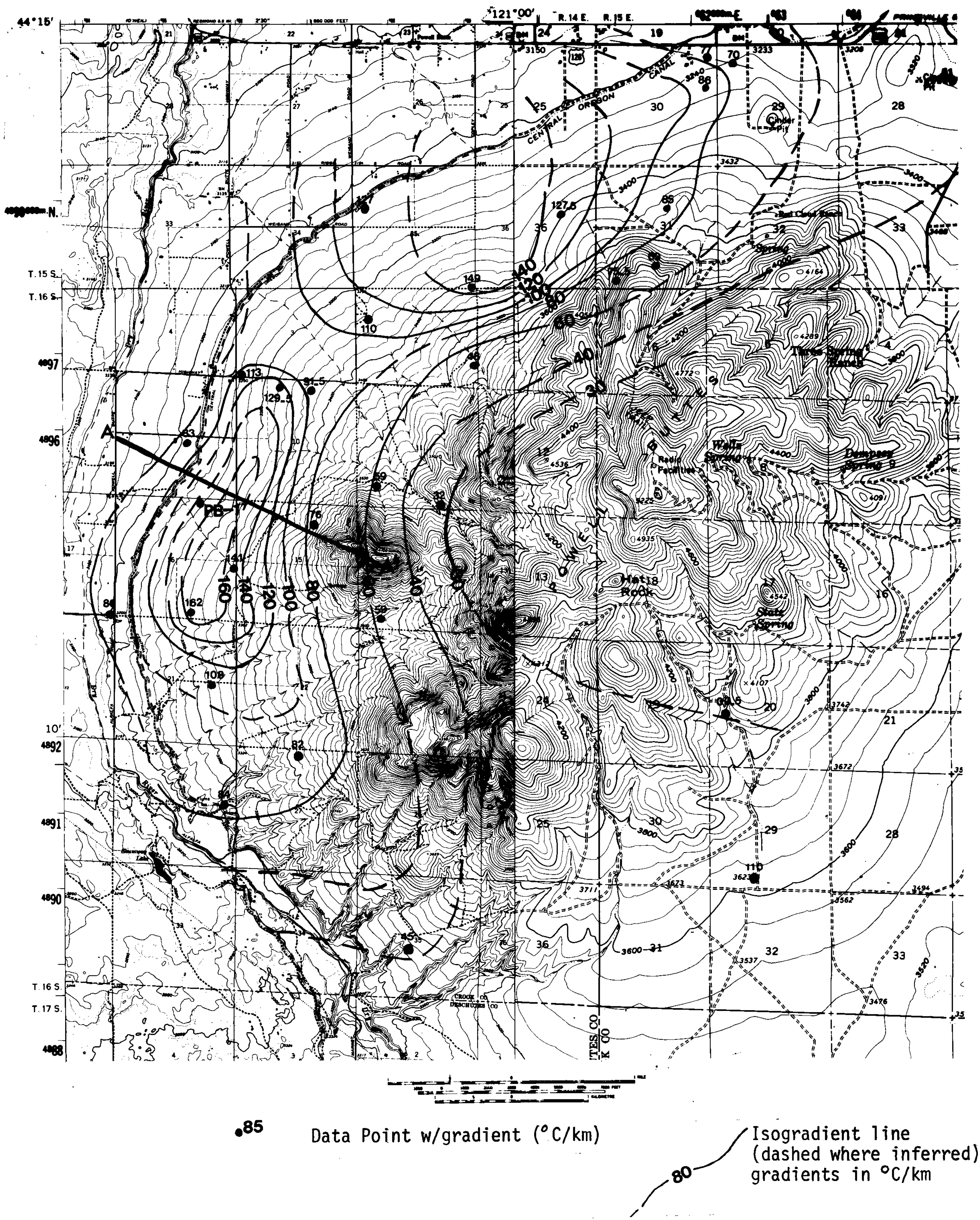

Figure 2: Isogradient map of the Powell Buttes area, Oregon. (From Brown, Black, and McLean, 1980b)

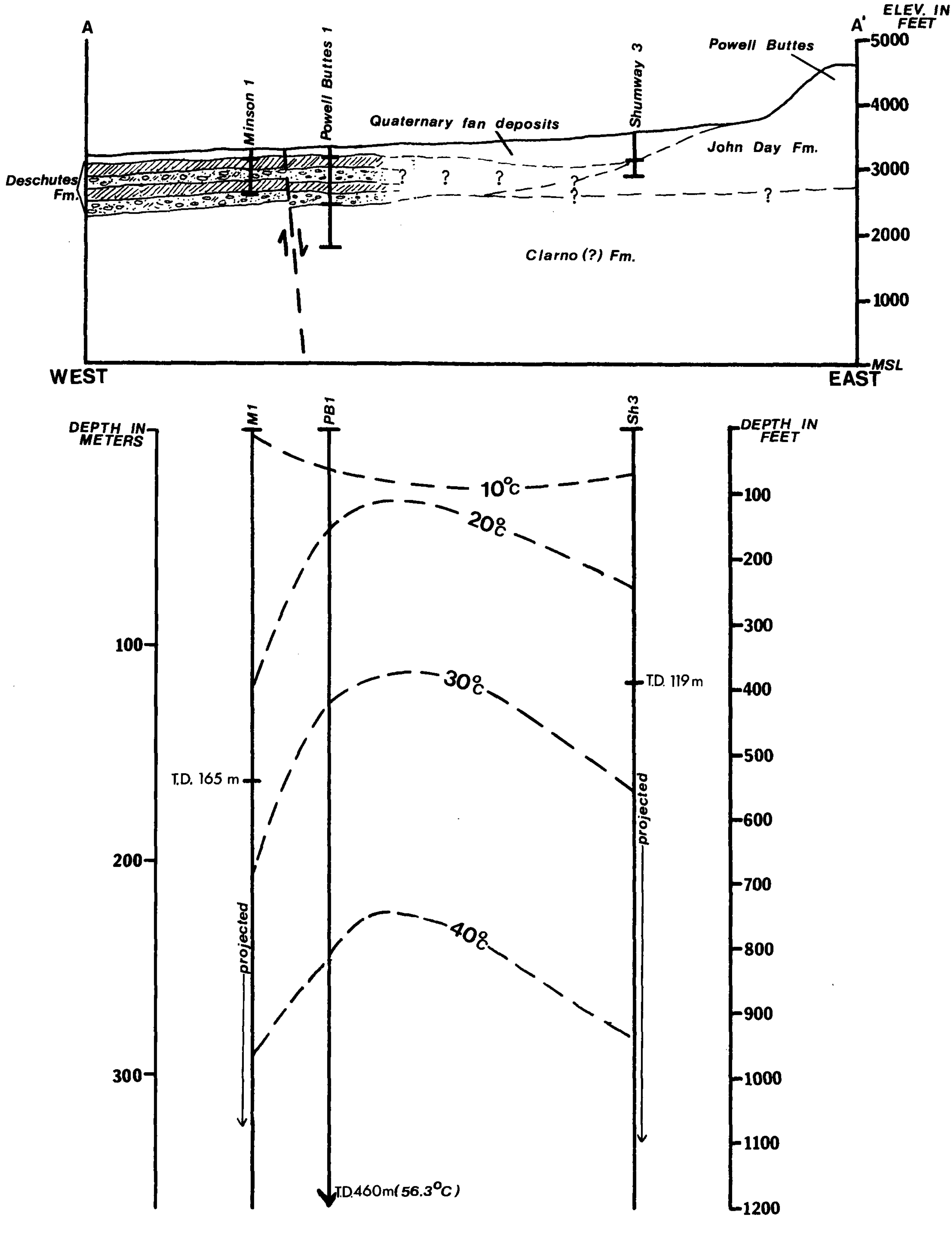

Figure 3: Geologic cross-section and isothermal plot through drill holes. (Modified from Brown, Black, and McLean, 1980b)

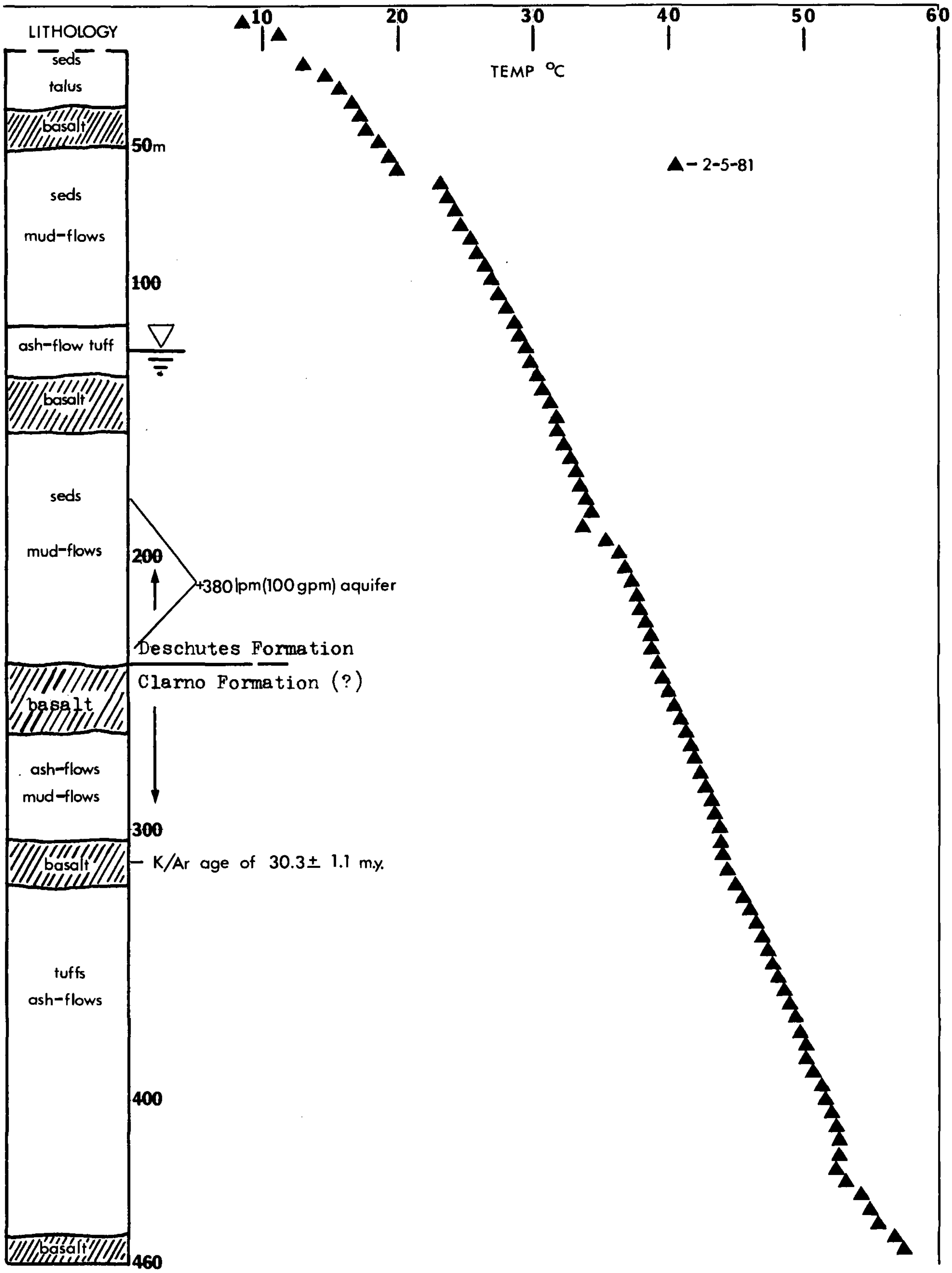

Figure 4: Lithologic and Temperature Log of Powell Buttes No. 1 Intermediate Gradient Hole

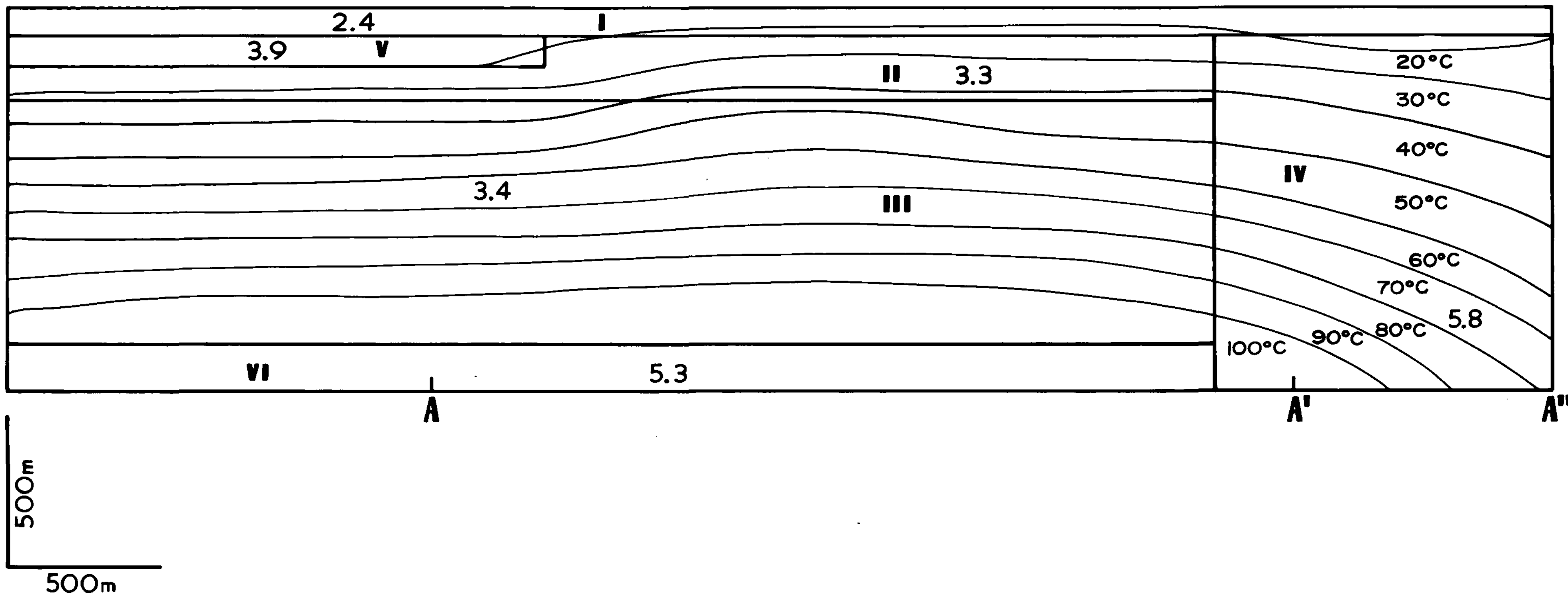

Figure 5. Finite Difference Thermal Conductivity Model of Powell Buttes

Cross section is through PB-1 (16S/14E-16Aba) and parallel to A-A'. The thermal conductivities (in cal/cm sec ^{o}C) of the blocks are weighted averages of measurements made on cutting samples from PB-1. 10^{o} isotherms were generated after 9909 years (500 iterations) from an input of 3.0 HFU at the base of the model. The temperature of Block V was maintained at 16^{o}C to simulate high water flow rates in the basalt aquifers to the west of the buttes.

this region) over a period of about 10,000 years. The computer program takes into account the measured conductivity of the stratigraphic sequence and shows how the highly conductive rhyolitic dome on the east side of the anomaly could combine with the heat sink of water-saturated younger rocks on the west side to produce a purely conductive thermal anomaly. Preliminary calculations indicate that a similar thermal model could be generated by very slow flow of warm water through permeable surface rocks. It must be emphasized that these models are mathematical possibilities and do not constitute proof.

The models should be tested by additional intermediate-level (600 m) drilling below the insulating Mio-Pliocene cap rocks. The relatively high absolute temperature of the anomalous area also indicates that relatively high temperatures can be realized at moderate depths (e.g., 150°C at about 1840 m or about 6,050 ft.). Deepening PB-1 to about 1,220 m (4,000 ft.) would help to assess the permeability of deeper reservoir rocks to evaluate the resource. Deep drilling should, however, be done with the realization that the older rocks (chiefly Clarno Formation) tend to be very impermeable in other areas where their hydrologic properties have been studied (e.g., Robison and Laenen, 1976). It might, therefore, be wiser to aim deep drilling at areas likely to have considerable fracture permeability. Such areas might be outlined by a careful program of intermediate-level drilling aimed at mapped faults and lineations within the thermal anomaly. Hopefully, areas with deep convection along fault and fracture zones would produce thermal anomalies which could be seen by intermediate-level drilling. In any case, the currently defined 30° to 40°C hydrothermal system can be used by local residents for a variety of direct-use applications.

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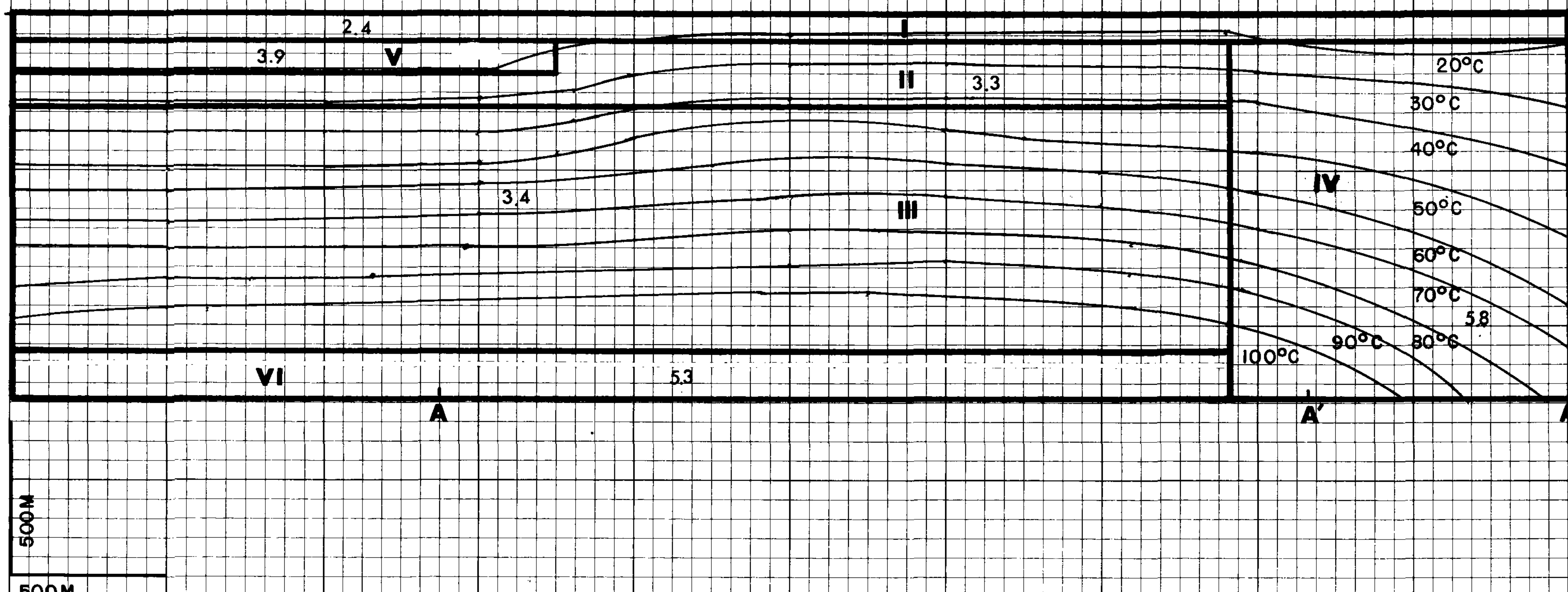
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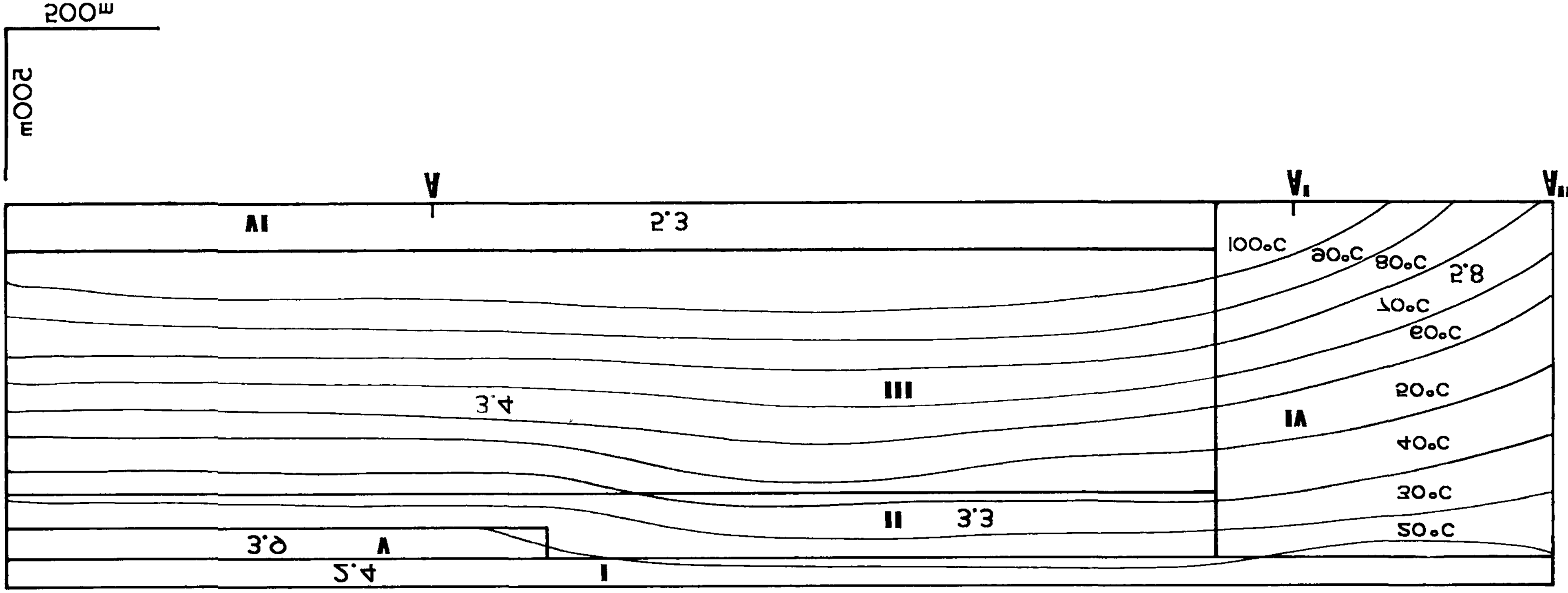

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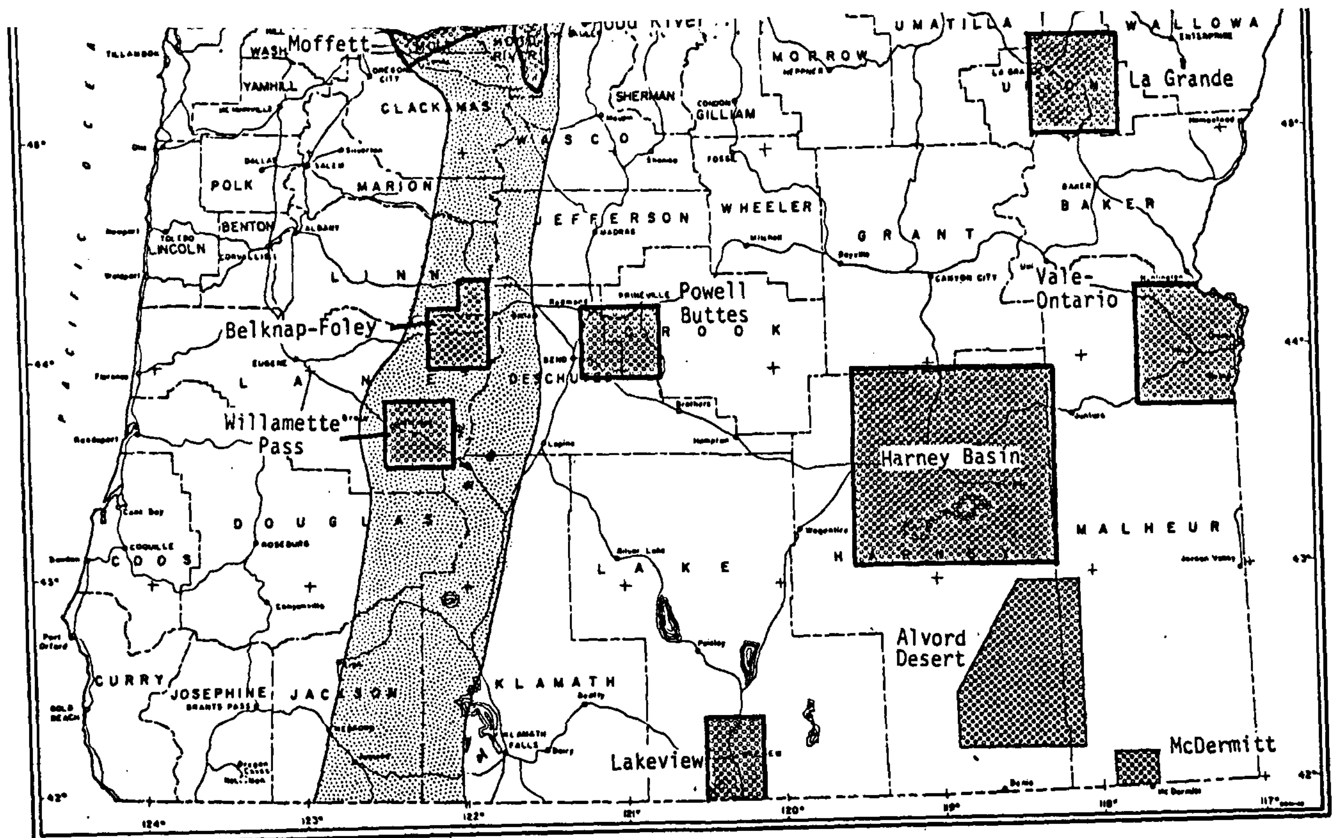
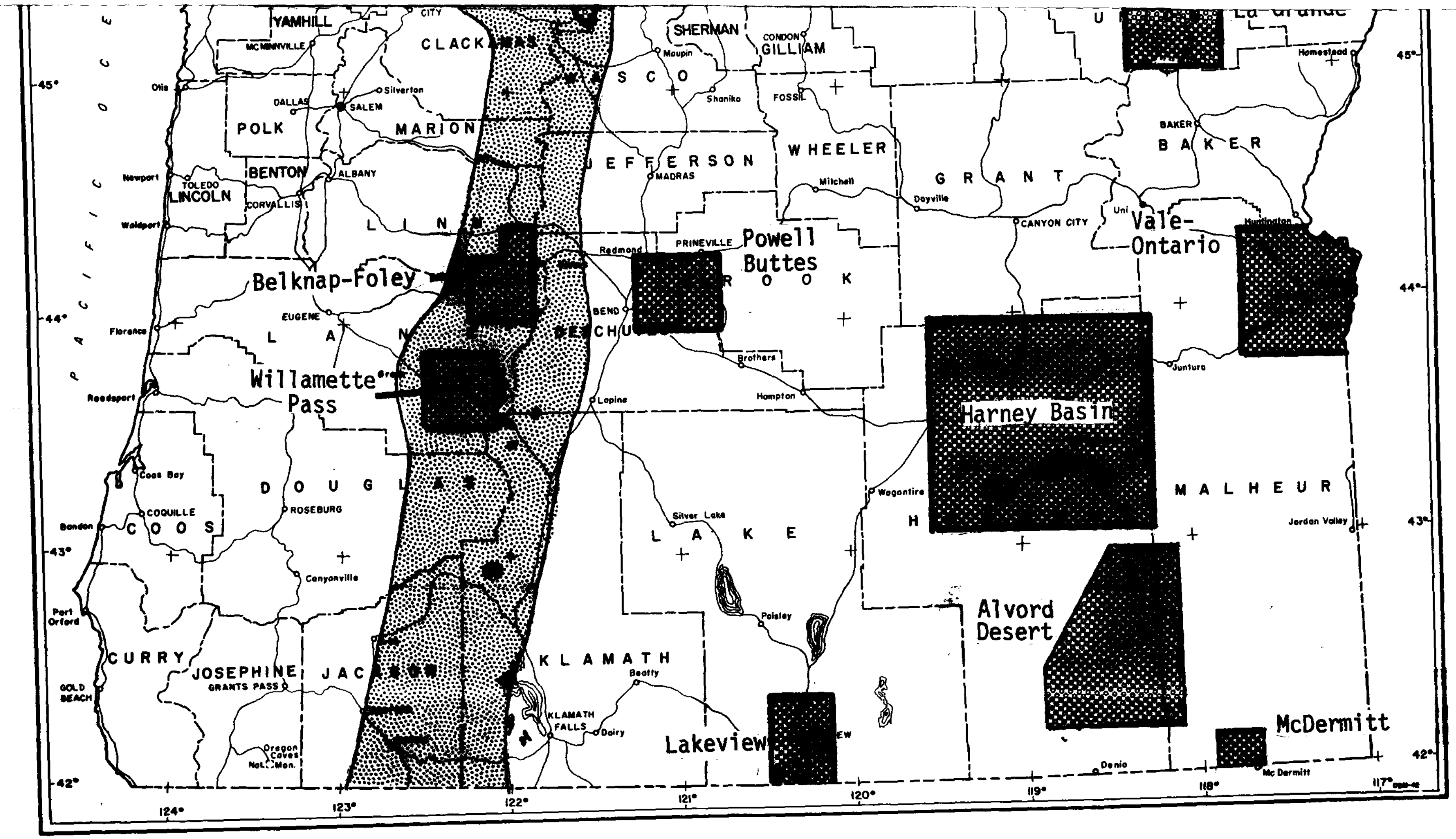


Low-to Moderate-Temperature Resource Assessment Area (target areas for detailed assessment have bolder outline).

Cascades Study Area.

FIGURE 1. Major study areas for the Oregon Department of Geology and Mineral Industries geothermal project.


La Grande
Vale-Ontario
Powell Buttes
Belknap-Foley
Willamette Pass
Harney Basin
Alvord Desert
Lakeview
McDermitt
YAMHILL
CLACKAMAS
SHERMAN
GILLIAM
WASCO
POLK
MARION
BENTON
LINCOLN
LINN
JEFFERSON
WHEELER
GRANT
BAKER
CROOK
LANE
DESCHUTES
DOUGLAS
COOS
CURRY
JOSEPHINE
JACKSON
KLAMATH
LAKE
HARNEY
MALHEUR
PACIFIC OCEAN
Salem
Eugene
Bend
Redmond
Prineville
Madras
Roseburg
Coos Bay
Klamath Falls
Canyon City
Baker
Burns

...RCEIVED ON LANDSAT IMAGERY OF CENTRAL TEXAS-- ...IONS TO GEOTHERMAL RESOURCE ASSESSMENT[1]

...Woodruff, Jr. and S. Christopher Caran

Linear trends of various geologic, geochemical, and geophysical phenomena are commonly observed in conjunction with geothermal resources. Examples include the geographic control of warm springs in western Virginia by fracture zones that trend normal to prevailing Appalachian structures (Geiser, 1979). Similar cross-cutting relations occur in the Ouachita Mountains at Hot Springs, Arkansas (Bedinger and others, 1979). Lineaments have also been frequently employed as a tool to delineate hydrothermal manifestations (both as ore deposits and as thermal resources) in the Cordillera of western North America. Presumably the coincidence of hydrothermal phenomena and lineaments relates in a general way to structural/tectonic influences on heat flow and on the migration of fluids. Simply put, active tectonic zones are areas of crustal discontinuities (thin crust in rift zones, for example), locally high heat flow, and marked seismicity; some of the surface manifestations of active tectonism in these areas may be perceived as lineaments. Yet even for relict areas of tectonic disturbance (for example, "dormant" orogens), there commonly are thermal expressions that are thought to result from deep circulation of waters along fractures and steeply dipping beds. Such areas are also often denoted by lineaments. However, as observed by Steeples and others (1979), some of these presumed "dormant" areas may still be more active than is generally recognized; microseismicity, for example, may accompany local heat-flow anomalies, and if hydrologic conditions are favorable, a geothermal resource may occur.

[1]Publication authorized by the Director, Bureau of Economic Geology, The University of Texas at Austin.

The work by Steeples and coworkers is an important point of departure for our discussion, because that study dealt with a buried orogen, the Nemaha Ridge of the Mid-Continent. Moreover, the buried structure is also denoted by a trend of lineaments at the earth surface (McCauley and others, undated). Buried structural trends, then, may have surface expression through lineaments, and such features may be favorable loci for geothermal resources because of either locally high heat flow or heat convection by upwelling waters.

As part of a statewide assessment of geothermal resources in Texas, we evaluated lineaments perceived on Landsat images. The logic behind this effort is that, except for the Trans-Pecos area, Texas largely comprises terrain that is underlain by flat-lying sedimentary rocks. Fundamentally, our guiding hypotheses have been: 1.) geothermal anomalies accompany structural discontinuities; and 2.) structural discontinuities--even buried (or dormant) features--may be subtly expressed at the ground surface. A synoptic overview afforded by Landsat images provides a means to perceive large-scale (albeit subtle) features that indirectly indicate structural, hydrologic, and thermal anomalies.

We present here a case study that shows the convergence of seemingly diverse phenomena. As in the aforementioned studies of the Nemaha Ridge, we focus on a buried orogen--the Ouachita structural trend in Central Texas, a foundered hinge zone between the Texas Craton and the downwarping Gulf Coast Basin (fig. 1). Our aim is to show that lineaments may provide evidence for buried structures, which, in turn, apparently control the location of geothermal anomalies.

We developed a method for perceiving lineaments statewide that entailed each of three observers[2] independently viewing 51 Band-5 Landsat images for

[2]For the study area, the third person was Gary E. Smith. Eric J. Thompson assisted in these efforts and worked subsequently on numerical evaluation of these lineaments.

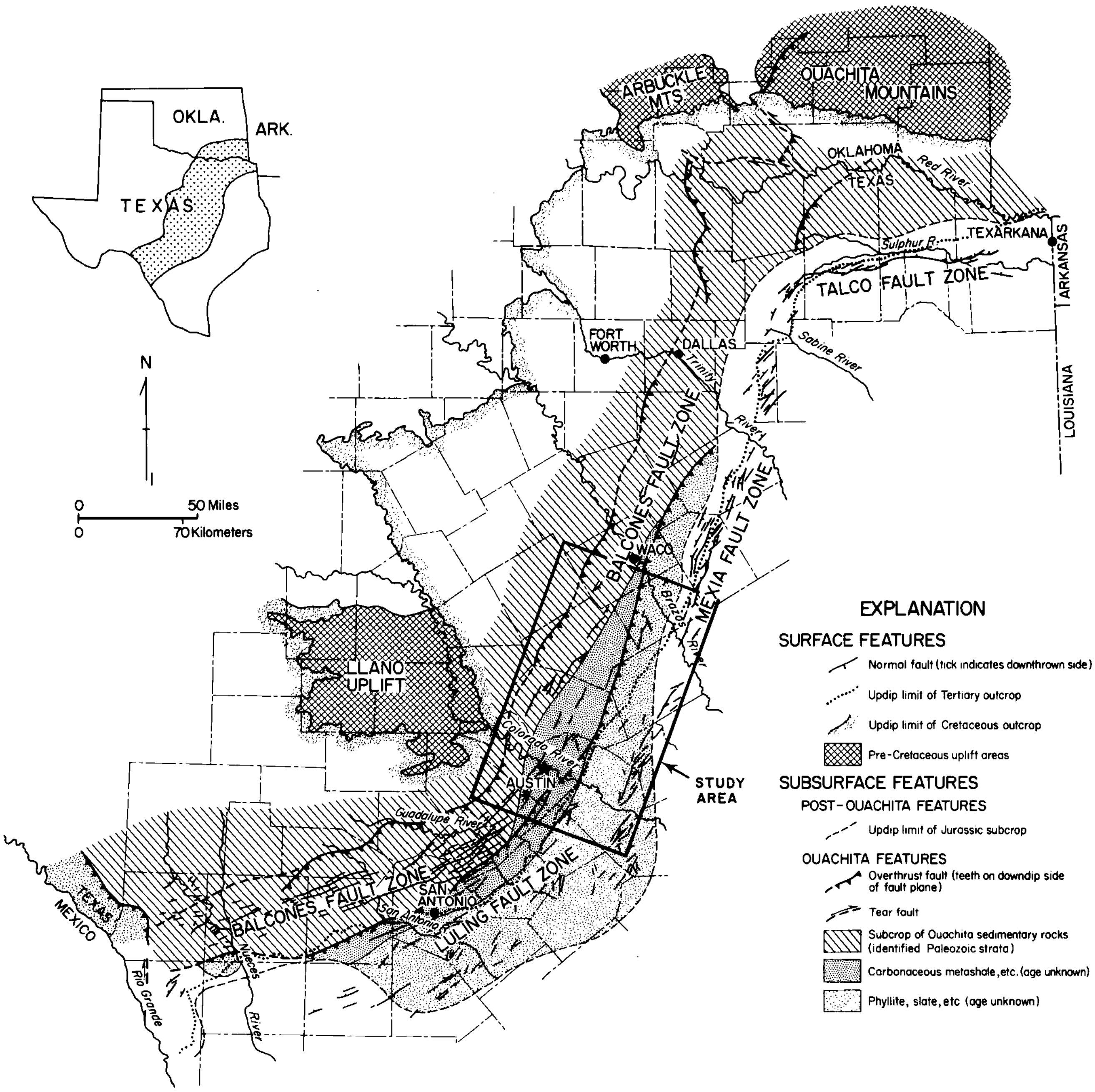


Figure 1. Balcones/Ouachita structural trend, Central Texas and location of lineament study area.

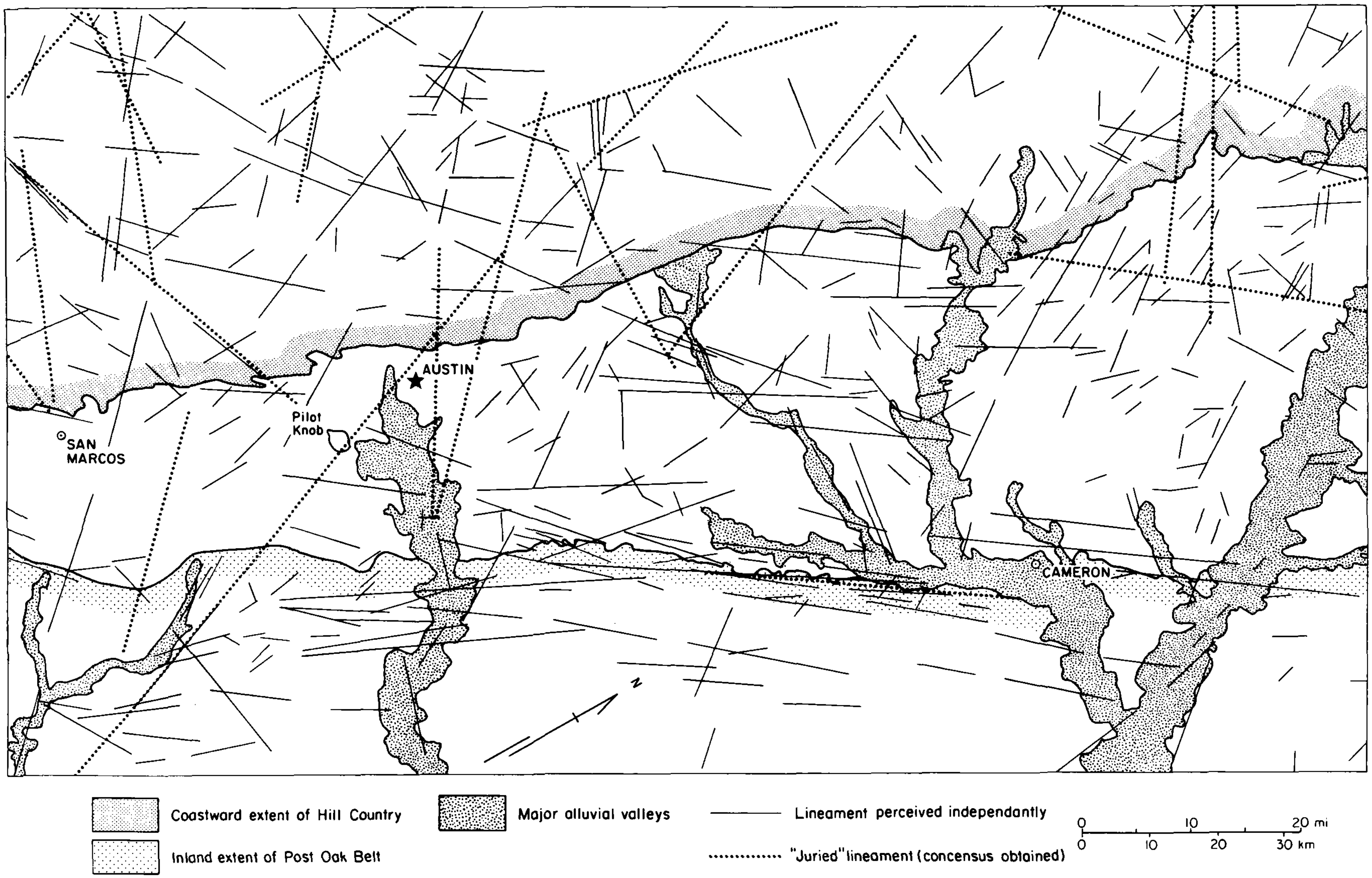


Figure 2. Lineaments perceived in study area; note, Blackland Prairie is the physiographic area between the Hill Country and the Post Oak Belt.

two periods of 30 minutes each (see Caran and others, 1981). For a pilot study in Central Texas we also conjointly viewed several mosaicked images and thus mapped large-scale, throughgoing figures as "juried" lineaments (fig. 2) along with the other linear features that we perceived independently. These two operations resulted in our perceiving more than 400 lineaments in an area of approximately 8680 km^2. We consider the lineaments, thus perceived, to be "raw data" without particular value until they are interpreted. In short, there is probably a high "noise to signal" ratio in a depiction of this kind. The interpretation of these data should allow a better discrimination of the salient information (signal) from the random background (noise).

Two sets of features stand out in this depiction of lineaments in Central Texas. One set trends oblique to the strike of stratigraphic units (that is, oblique to the boundaries of the Hill Country and the Post Oak Belt as depicted in figure 2). The other set aligns roughly parallel to the prevailing strike.

The oblique-trending lineaments compose mainly "juried" lineaments, although there are also families of the generally shorter features perceived by individual observers that, in the aggregate, produce orientations oblique to strike. We have no hypotheses on the implications of these features, except to note that Pilot Knob, a Cretaceous marine volcanic plug southeast of Austin, lies along the intersection of two of these large lineaments (fig. 2). Also, the northwest orientation exhibited by some of these features parallels a predominant trend of the Brazos River alluvial valley which, when depicted at a regional-scale, may be a giant lineament extending for over 120 miles oblique to the structural and depositional strike of the region.

The lineament trend that is parallel to strike of strata is the expected set--especially along the Balcones Fault Zone. This is because of the abrupt discontinuity in bedrock, soils, vegetation, and land use that occurs along this structural trend. There are similar but more subdued surface expressions along the contacts of most formations along the Gulf Coastal Plain of Texas. The initial implication of this set of lineaments is that we have merely rediscovered the Balcones Fault Zone, or we have perceived the contacts of mapped stratigraphic units. If these findings account for the entire significance of the strike-parallel lineaments, then that family of lines is clearly trivial. We intend to demonstrate, however, that we have, instead, perceived a previously unrecognized zone of structural dislocation that has implications on the location of geothermal resources in Central Texas.

Of the lineaments parallel to strike, one set is especially prominent. These features compose a northeast-trending family of lines along the boundary between the Post Oak Belt and the Blackland Prairie. The lineaments lie mainly along the alluvial reaches of Brushy Creek near its confluence with the San Gabriel River and the Little River system. We have named the feature the Brushy Creek Lineament. Actually, several lineaments make up this zone. They include the coincidence of a major (juried) lineament and a high density of parallel, smaller features that align with relict and modern stream reaches, a straight drainage divide, a west-facing topographic escarpment, and the contact between the Midway and Wilcox Groups of Eocene age. This stratigraphic contact is also responsible for the major physiographic break between the Post Oak Belt and the Blackland Prairie and the attendant changes in soils, vegetation, and land use.

The intriguing thing about this particular part of the Post Oak Belt/Blacklands border is its remarkable linearity. In fact, this family of lines is much more strongly expressed in our lineament survey than is the Balcones fault-line escarpment, which is an area of similar physiographic importance but one clearly documented as a structural zone. Yet, no major structural discontinuity has been previously documented between the Post Oak Belt and the Blackland Prairie;

there, only a single short fault is mapped at the surface along this trend (Barnes, 1974). It is our thesis that this lineament zone is the surface expression of a deep-seated structural disturbance of a significance similar to the Balcones fault trend farther west.

To test this thesis we collected several types of independent evidence that bear on subsurface structures in the area (fig. 3). We have located subsurface faults displacing the basal Cretaceous Hosston Sand and the (shallower) Edwards Limestone (see Woodruff and McBride, 1979). We also obtained additional well data and constructed a new map showing faults displacing the Buda Limestone (a prominent subsurface datum on electric logs) along the trend of the Brushy Creek Lineament.

Several other structurally related phenomena also converge along this lineament trend. They include buried igneous plugs and associated oil fields, the updip subcrop limit of Jurassic rocks (the first indication of marine conditions along the western margin of the Gulf of Mexico during Mesozoic time), and the proximity of major basement discontinuities in the Ouachita rocks as mapped by Flawn and others (1961). In short, the convergence of these diverse structural data indicate that the Brushy Creek Lineament zone delimits the eastern part of the Ouachita hinge, just as surface faults of the Balcones system roughly delimit the western margin of that hinge.

The eastern margin of the Balcones/Ouachita structural trend is important in a geothermal context because this zone marks a major change in orientation of depositional systems--from dip-fed fluvial sand bodies on the west to strike-fed lagoonal and marine, sand, mud, and carbonate deposits on the east. This change in depositional systems (documented by Woodruff and McBride, 1979) marks the deepest part of a hydrologic system that allows ready access of meteoric recharging waters to depths sufficient for markedly increased water temperature

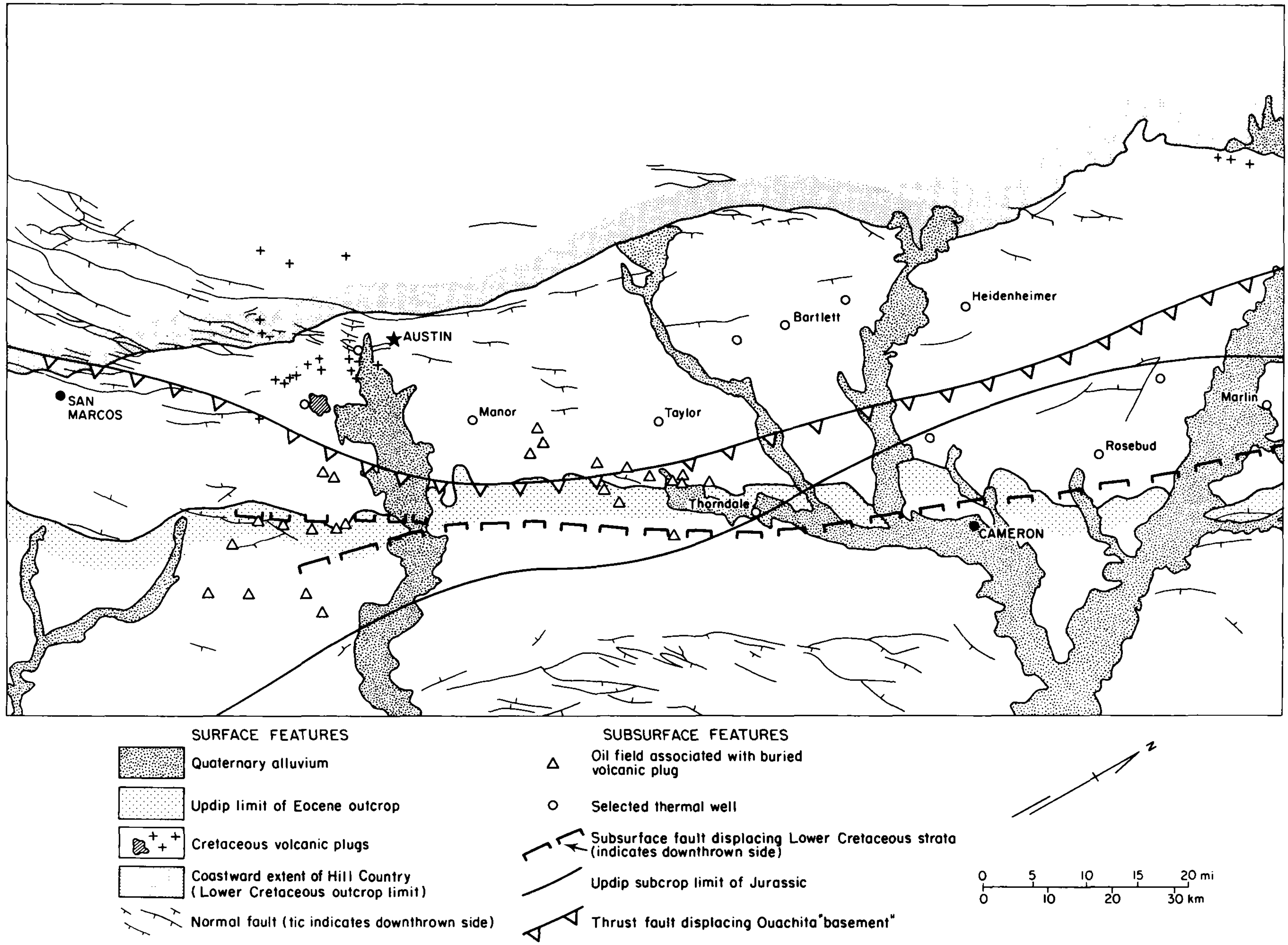


Figure 3. Surface and subsurface structural features in study area; note convergence of features near the Brushy Creek Lineament southwest of Cameron.

(given prevailing geothermal gradient) while maintaining low to moderate concentrations of dissolved solids.

A contour map of geothermal gradients across the study area shows gradient anomalies to generally align along the trend of the Brushy Creek Lineament (fig. 4). These anomalies may not indicate a zone of high heat flow (as would be expected in an area of active tectonism) but instead may indicate a locus of upwelling waters. The alignment of oil fields argues for this interpretation, in that the igneous plugs provide the preferred avenues for upward flow of waters and entrained hydrocarbons. As pointed out by Plummer and Sargent (1931), such areas of upwelling waters and hydrocarbons are part of expected basin-wide hydrologic interactions that also include geothermal gradient anomalies and the occurrence of warm (often saline) waters at a relatively shallow depth.

There are several thermal water wells in our study area. No discernible trend exists, however, because the locations of these wells are dictated by the prior siting of towns for which the wells supply water. On the basis of this lineament survey, we conclude that an exploration program for hydrothermal waters of potable quality should focus on the western side of the lineament zone in order to tap downward flowing recharge systems within the dip-oriented depositional facies. Hotter waters may occur on the eastern side of the lineament zone, but these would largely be upwelling waters from deep within the Gulf Coast Basin, and thus salinity would probably be quite high.

In summary, lineaments may provide a tool for locating "blind" geothermal resources, because they are subtle indications of subjacent tectonic disturbances in areas covered by flat-lying rocks. Lineaments may be the surface expressions of fractures propagated upward through undisturbed strata. Such fractures may provide enhanced permeability avenues for downward flow of recharging

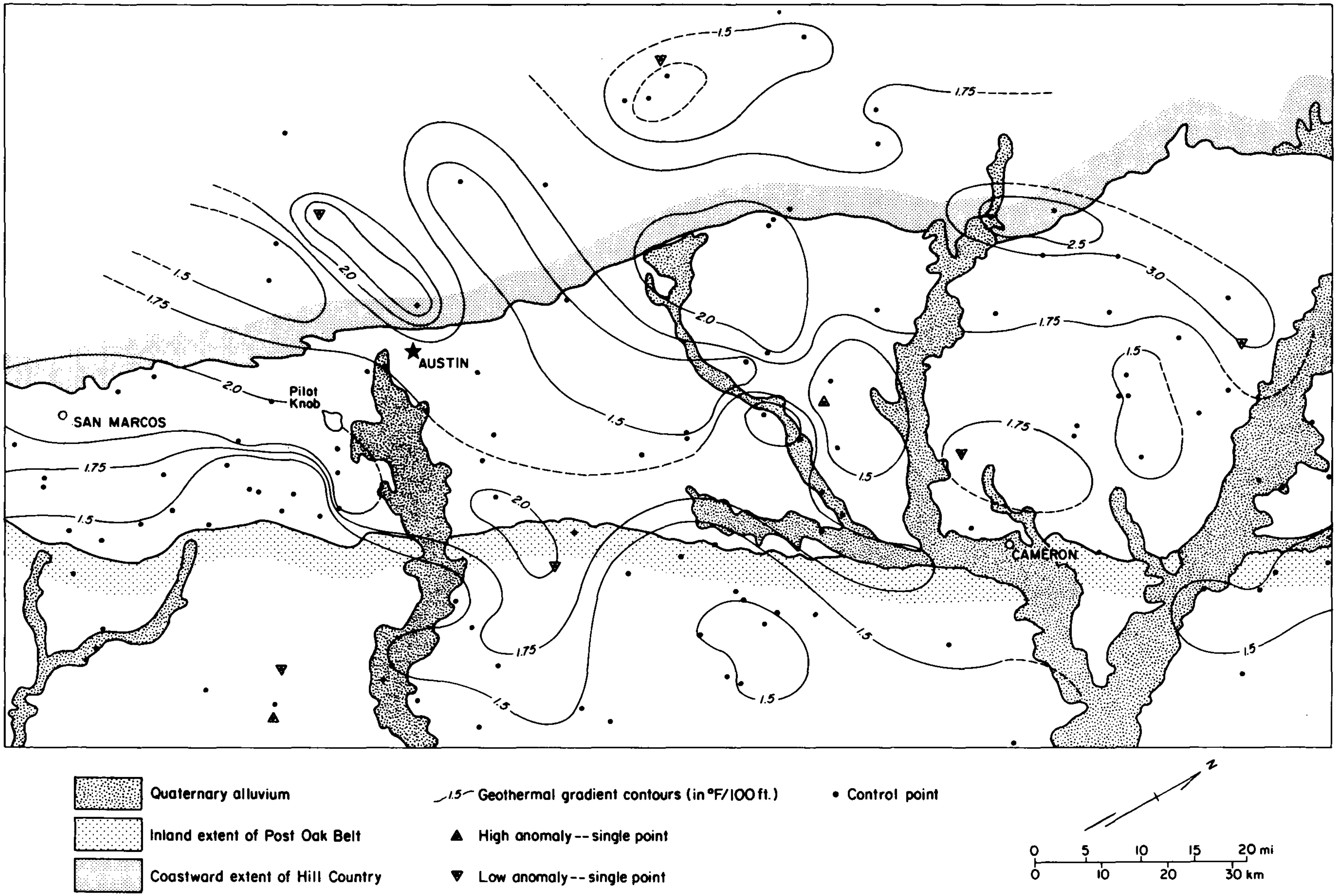


Figure 4. Geothermal gradient contours across lineament study area.

waters to a depth sufficient for thermal enhancement above mean annual air temperature. Also, such areas of high lineament concentrations may mark the loci of upwelling of hot (and often saline) waters from deep within a sedimentary basin.

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