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GEOPHYSICAL STUDIES OF ACTIVE GEOTHERMAL SYSTEMS IN
THE NORTHERN BASIN AND RANGE

Stanley H. Ward

Earth Science Laboratory/University of Utah Research Institute
420 Chipeta Way, Suite 120
Salt Lake City, Utah 84108

ABSTRACT

Most of the geophysical data in the public domain, acquired in exploration for high temperature geothermal systems in the Northern Basin and Range Province, have been reviewed. Sufficient data are available to compare 14 methods at 13 sites, but only 110 entries occur in the 14 by 13 matrix whereas 182 entries would have been optimum. Only four of the sites studied are believed to be capable of production of commercial electricity while three others probably will be placed in this category in the next few years. For three additional systems believed to be capable of commercial production, insufficient geophysical data are available, in the public domain, to permit review.

On a rating scale of 1 equals good through 4 equals poor, no geophysical method has a mean ranking of 1. Five methods rank about 2, six rank about 2.5, and three rank about 3, while none ranks 4. This ranking system is subjective, but uniformly applied to the question, "what contribution has the method made to understanding the reservoir or the presumed reservoir, at site X". The averages of the evaluations at each site are the rankings given above. No combination of any four methods has been successful at more than one site where "successful" means a ranking of 1 or 2. The most useful of the methods, judging by their average rankings, are heat flow, microearthquakes, gravity, resistivity, and self-potential. The least effective methods are earth noise, reflection seismology, magnetics, magnetotellurics and tellurics. The radiometric and induced polarization methods were excluded from the comparative study due to a paucity of data. The CSAMT method has been included with CSEM for purposes of this comparative study.

Twenty-three applications of geophysical techniques were judged to be good, 29 were judged to be fair, 48 were judged to be questionable, while 10 were judged to be poor. Only 52 of 110 entries in the 14 by 13 matrix were judged to be fair or good; i.e. 53 percent of the geophysical applications gave questionable or poor results. Of the 58 applications where questionable or poor ratings were assigned, 41 applications were judged to be simple failure of the geophysical method to solve the problem at hand. However, 17 of the

applications could have been better if improved technology or interpretation were available; some poor technology was applied in the middle of the 1970-80 decade.

Overall, this review provides a somewhat discouraging picture. This may be due in part to inadequate subsurface control and to inadequate survey design, execution and interpretation as a result of lack of experience. I would tend to use heat flow, microearthquakes, gravity, resistivity (or CSAMT), and self-potential methods at all prospects. Once these data were interpreted and correlated, I would then decide whether or not additional geophysical surveying was required and/or justified. All geophysical surveys should be designed and executed only after the geology has been mapped carefully, an integrated interpretation has been made of any and all other available earth science data, and one or more specific questions have been formulated for the survey to answer. The surveys should be designed with one or more conceptual geological models in mind and the density and extent of the geophysical coverage should be designed to provide adequate cover of the area dictated by such conceptual geologic models.

Within this manuscript, I have discussed the advantages and limitations of all of the geophysical methods considered; such a discussion is essential to their evaluation. Examples of overlapping geophysical data sets are given for three igneous-related geothermal systems and for two systems without obvious igneous relationships. No systematic difference in the past application to geophysical exploration for igneous-related systems versus those with no obvious igneous relationships is evident.

1.0 OBJECTIVES

Heat is the essence of a geothermal resource, and its role in the development of such resources, the theme of this symposium, is fundamental. However, this leads to the question of how much surface manifestation of heat is necessary to define a geothermal prospect worthy of exploration. Clearly, heat flow specialists at this conference will attempt to address this question directly. I have been asked to address it indirectly and to concentrate on the other geophysical methods used

in geothermal exploration and to provide a "common thread" among geophysical results for prospect-sized areas of 10 to 1000 square miles.

The objective of this paper, accordingly, is to focus on a comparative study of the problems and successes encountered with the gravity, magnetic, passive seismic, active seismic, self-potential, resistivity, passive electromagnetic, and active electromagnetic methods. Such a comparative study would be incomplete if I failed to draw conclusions as to the cost-effectiveness and preferred role of each of the methods listed above. At the risk of criticism from special interest groups, I shall draw such conclusions. The extent to which I am able to justify my conclusions must be judged by the individual reader in relation to his own experience. With the passing of time and further drilling, my conclusions undoubtedly will be modified. At least, however, I shall have provided a datum to which further analyses may be referenced.

For the purposes of this paper, I use Edmiston's (1982) definition of the Northern Basin and Range Province (Fig. 1). The number and evaluation of geothermal prospects in Figure 1 differs, however, from that given by Edmiston (1982). It is my intent to convey an overview of the contributions made by each geophysical method to understanding the geothermal reservoir of most of the prospects of Figure 1, independent of whether the prospect may now be classified as a discovery or otherwise. In this process, I shall use examples of geophysical data from only a few of the prospects. This permits the paper to be reasonably concise and yet illustrative. The zone of enhanced extension, shown in Figure 1, is my own interpretation.

2.0 DISTRIBUTION OF KNOWN HIGH TEMPERATURE RESOURCES

As noted by Edmiston (1982), "The northern Basin and Range Geologic Province of the western U.S.A. has been widely recognized as a highly prospective area for high-temperature geothermal reservoirs. Yet, only six apparent discoveries resulted from the drilling of 53 geothermal wildcat wells in this area from 1974 through 1981. This relatively lower success rate can be partly attributed to the difficulty of developing accurate geological and geophysical models in this area prior to drilling. However, it may also indicate that large, high-temperature geothermal reservoirs may be less common in this area than thought previously." Mansure and Brown (1982) support Edmiston's observations by forecasting (Figure 2) that the rate of drilling of geothermal wells through the year 2000, will be about the same for northern Nevada as for Roosevelt Hot Springs or Valles Caldera. If this forecast is correct, then the electric power generated by geothermal energy in northern Nevada will indeed be modest, i.e. of the order of 500 MWe. One might conclude, from the works of Edmiston (1982), Mansure and Brown (1982), and Benoit and Butler (1983), that for the whole of the northern Basin and Range Province there will be electrical production, by the year 2000, only at Beowawe, Coso, Desert Peak, Dixie Valley, Humboldt House, Roosevelt Hot Springs and Steamboat Springs. Soda Lake and Long Valley should possibly be added to this list. All of these potential resources lie in the eastern and western margins of the northern Basin and Range Province as Figure 1 shows; these are regions of enhanced crustal extension. Of course, moderate- to low-temperature geothermal resources are much more widespread.

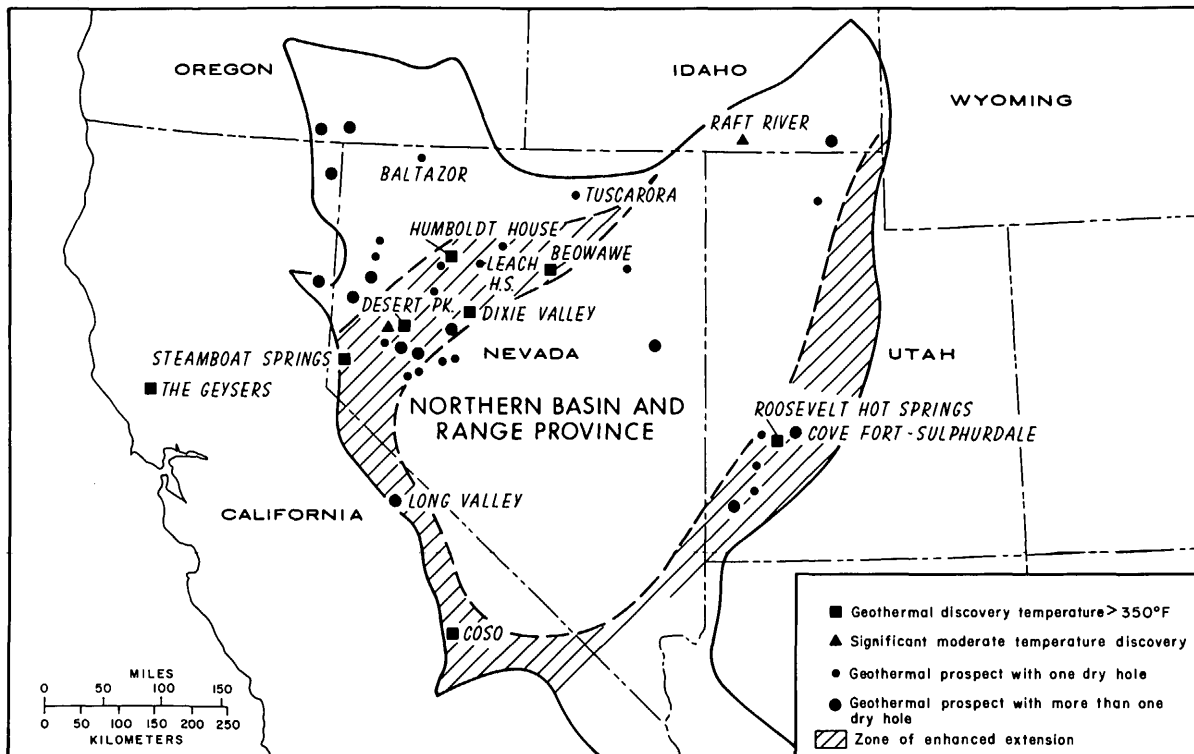


Figure 1. Map showing the location of geothermal discoveries and unsuccessful geothermal wildcat wells in the northern Basin and Range Province (after Edmiston, 1982).

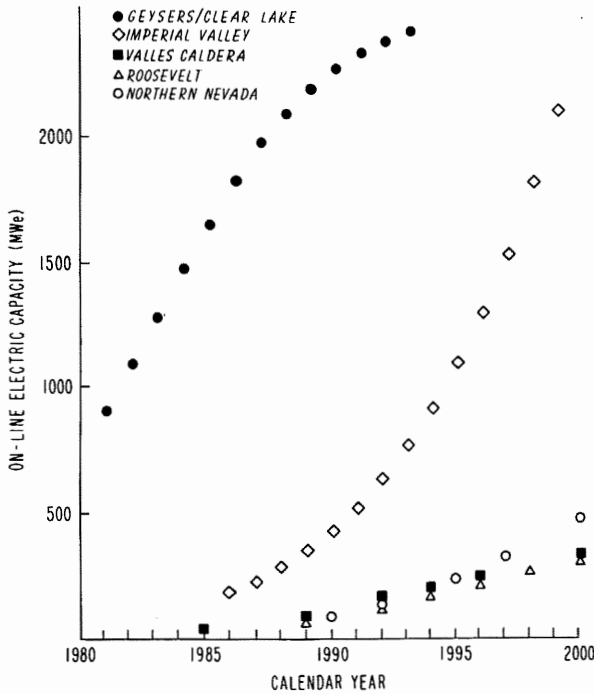


Figure 2. Forecast of geothermal electric power capacity (after Mansure and Brown, 1982).

Is this distribution of known, high temperature resources due to the geologic environment created by enhanced crustal extension, or is it because, in part, that we have not developed geophysical techniques capable of detecting and delineating high temperature resources when these resources are hidden? One would certainly have expected more than one exploitable high temperature resource along the eastern margin of the Basin and Range, if high heat flow, thin crust, and Quaternary volcanism are the key indicators of prime prospecting ground.

3.0 GEOPHYSICAL METHODS FOR GEOTHERMAL EXPLORATION

In the Northern Basin and Range Province, the basic target for geophysical surveys in most cases is a fracture and fault system filled with thermal fluids. The reservoir itself may be relatively near surface as in parts of Roosevelt Hot Springs, or it may be 5,000 to 10,000 feet deep as at Beowawe or Desert Peak. Many geophysical techniques are only capable of delineating the top of the plumbing system and this may or may not contribute to understanding the reservoir (C.M. Swift Jr., pers. com.). With this thought in mind, let us turn to the geologic feature or features each method is usually directed toward, as listed in Table 1. A brief discussion of each target, in relation to the appropriate method, follows.

earth noise

Earth noise is associated with hot spring activity, fluid circulation, and fluid phase changes in geothermal systems, i.e. active hydrothermal processes, and can be a direct indicator of the presence of a geothermal system. This noise, and noise generated in other ways, may also be used to map the three dimensional distribution of velocity and attenuation (Iyer and Hitchcock, 1975; Liaw and McEvelly, 1979; Liaw and Suyenaga, 1982).

microearthquakes

Microearthquakes are associated with active faulting. They can occur more or less continuously, but more frequently they are episodic. Hypocenter locations give information on regions where faults are active and can help determine strike and dip. Fault-plane solutions can yield information on direction of movement. The three-dimensional distribution of seismic wave attenuation may be mapped. By measuring both compressional and shear wave velocities, Poisson's ratio can be computed; it is expected to be low over a vapor-dominated system and high over a fractured, liquid-dominated system. The source parameters seismic moment, stress drop, fault slip, and source radius may be estimated, but their value in reservoir definition and delineation is unknown at present. Minimum hypocentral depth of microearthquakes is useful in estimating the thermal regime surrounding geothermal reservoirs (Ward, R.W. et al., 1979, Majer and McEvelly, 1979).

teleseisms

If a sufficiently distant earthquake is observed with a closely spaced array of seismographs, changes in P-wave travel-time from station to station can be taken to be due to velocity variations near the array. Travel-time residuals are computed as the observed arrival time minus that calculated for a standard earth. A magma chamber beneath the geothermal system would give rise to low P-wave velocities and hence to late observed travel times (Iyer et al., 1979; Reasenberg et al., 1980; Robinson and Iyer, 1981).

refraction and reflection

The seismic refraction and reflection methods can be used to map the depth to the water table, stratigraphy, faulting, intrusions, and geologic structure in general. They may also yield the subsurface distribution of seismic P-wave and S-wave velocities, attenuations and Poisson's ratio. Detection of a characteristic attenuation or a "bright" spot, as found over reservoirs in petroleum exploration, would be a useful feature (Ward, R.W. et al., 1979; Applegate et al., 1981), but this has not been reported clearly for any prospect in the northern Basin and Range province.

gravity

Density contrasts among rock units permit

GEOPHYSICAL TARGETS IN GEOTHERMAL EXPLORATION

<u>METHOD</u>	<u>TARGETS</u>
SEISMIC	
EARTH NOISE	Active hydrothermal processes, distribution of velocity and attenuation.
MICROEARTHQUAKES	Active faulting, fluid filled fracturing.
TELESEISMS	Deep magma chamber.
REFRACTION	Structure, distribution of velocity and attenuation.
REFLECTION	Structure, distribution of velocity and attenuation.
GRAVITY	Structure, alteration, densification, intrusions, distribution of density.
MAGNETICS	Structure, hydrothermal alteration, intrusives, extrusives.
ELECTRICAL	
RESISTIVITY	Faulting, brines, hydrothermal alteration.
INDUCED POLARIZATION	Hydrothermal alteration.
CSEM & SCALAR AMT	Faulting, brines, hydrothermal alteration.
MT / AMT	Structure, deep reservoir, magma chamber?, partial melt in deep crust or upper mantle.
SELF POTENTIAL	Fluid and heat flow.
TELLURICS	Faulting, brines, hydrothermal alteration.
RADIOMETRIC	Alteration, ²²⁶ Radon, ²²² Radium.
HEAT FLOW	Reservoir temperature.

use of the gravity method to map intrusions, faulting, deep valley fill, and geologic structure in general. A subsurface distribution of densities can be obtained in the interpretation process. Densification of porous sediments gives rise to gravity highs over some geothermal systems (Isherwood and Mabey, 1978).

magnetics

Magnetic susceptibility contrasts in the subsurface permit use of the aeromagnetic and ground magnetic methods to map the distribution of magnetite. In some instances, this distribution may be related to rock type, and hence rock units can be mapped. This can be useful in structural studies. In some convective hydrothermal systems, alteration will lead to destruction of magnetite and hence the magnetic method can be used to map zones of active hydrothermal alteration. It can also be used to determine the depth to the Curie isotherm (Isherwood and Mabey, 1978).

resistivity, controlled source electromagnetics (CSEM), controlled source audiomagnetotellurics (CSAMT), tellurics and scalar audiofrequency magnetotellurics

These electrical methods are capable of delineating low resistivity associated with brine-saturated and hydrothermally altered rocks in geothermal systems. Usually, the brine and alteration occur predominantly along faults, so these methods may map faults controlling a fractured reservoir. Alternatively, they may map

a stratigraphic unit that contains thermal brines and/or alteration. By virtue of resistivity contrasts among rock units, each of these methods can map faults, stratigraphy, intrusions, and geologic structure in general, independent of the presence of brine or alteration (Hoover et al., 1978; Ward, S.H. and Sill, 1982).

induced polarization

The induced polarization method is theoretically capable of mapping the distribution of pyrite and clays, alteration products in convective hydrothermal systems (Zohdy et al., 1973; Risk, 1975a; Chu et al., 1983).

MT/AMT

The tensor magnetotelluric/audiofrequency magnetotelluric method is usually too expensive to be used for mapping the resistivity distribution in the shallow parts of a geothermal system. Hence, it is more logically used to map regional structure, to map the deeper parts of convective hydrothermal systems, to attempt to map magma chambers, and to detect and delineate zones of partial melt in the deep crust and upper mantle (Ward, S.H. and Wannamaker, 1983).

self-potential

Self-potential anomalies over convective hydrothermal systems arise from electrokinetic and

thermoelectric effects. Accordingly, the self-potential method can give information on fluid flow and on the distribution of heat sources in the subsurface. This measurement can be made in the static state or dynamically when an injection experiment is under way (Anderson and Johnson, 1976; Corwin and Hoover, 1979; Sill, 1982a,b,c; Sill, 1983a,b).

radiometric

Gamma-ray spectrometry may be used to map the areal distributions of ^{40}K , ^{238}U , and ^{232}Th . The gamma-ray peaks used are at 1.46 MeV for ^{40}K , 1.76 MeV (^{214}Bi) for ^{238}U , and 2.62 MeV (^{208}Tl) for ^{232}Th . If ^{226}Rn or ^{222}Ra are present in a geothermal system, they will be detected in the ^{214}Bi peak, since they also are daughter products of ^{238}U decay (Ward, S.H., 1981). "An examination of hot-spring waters in Nevada indicates the presence of [^{226}Rn and ^{222}Ra], in varying abundances, in spring systems where CaCO_3 is the predominant material being deposited. Systems where silica predominates are relatively low in radioactivity" (Wollenburg, 1975). The use of alpha-cup detectors for radon emanating from geothermal systems has been reported by Wollenburg (1975) and Nielson (1978).

heat flow

Measurement of thermal gradients in holes drilled to 100 m provide shallow temperature data which may be direct evidence of a major source of heat at depth. However, flattening or reversal of thermal gradients frequently occurs so that the data from shallow holes may be misleading. For this reason, temperature gradient holes of 1000 m or greater are frequently employed (Blackwell, this volume).

4.0 THE PRINCIPAL PROBLEMS WITH GEOPHYSICAL METHODS IN GEOTHERMAL APPLICATIONS

The following discussion covers the principal problems encountered in applying geophysical methods in geothermal exploration. To facilitate understanding the problems, the elementary basis for each method is described where considered necessary. The amount of space devoted to each method largely reflects the author's experience. For example, someone who has been directly involved in applying reflection seismology in geothermal exploration would undoubtedly expand upon the section I have written. However, it is expected that most of the significant problems encountered in applying geophysical methods are at least mentioned here.

earth noise

The velocity V , frequency f , wavelength λ , and wavenumber k , of seismic waves are related through the equations

$$V = f\lambda = \frac{2\pi f}{k} \quad (1)$$

It has been demonstrated by several workers (e.g. Liaw and Suyenaga, 1982) that hydrothermal processes deep in the reservoir radiate seismic body waves in the frequency band 1 to 100 Hz. Contours of noise power on the surface should delineate these reservoir-associated sources of noise, i.e. the noise radiating from a deep reservoir ought to be evident as body waves of high phase velocity (Liaw and McEvelly, 1979).

Noise in the 1 to 100 Hz band also arises in nearby sources such as freeway traffic, trains, rivers, canals, waterfalls, pipelines, wind, cattle, interfering seismic wavetrains, etc. There is also a distant source of noise of unknown origin. Thus, there is always an ambient noise background upon which any seismic radiation due to hydrothermal processes is superimposed. This ambient noise exhibits a diurnal variation, being lowest in the early morning hours. It may be, in part, related to the diurnal solar heating cycle.

It is well known that seismic noise amplitudes are usually higher over alluvium and soft sedimentary basins than over hard rock. (Iyer and Hitchcock, 1975). Thus, noise power anomalies may merely reflect a local increase in sediment cover. The noise in valley alluvium is dominantly of high-wavenumber, i.e. low velocity, (Liaw and McEvelly, 1979) and is propagated as fundamental-mode Rayleigh waves with the alluvium serving as a waveguide.

An array of geophones spaced on 100 m centers, as commonly used in ground noise studies in the past, would give spurious results because spatial aliasing folds high-wavenumber noise into low-wavenumber noise. The spatial aliasing results in the appearance of noise of erroneously high velocity, such noise being interpreted as body waves, whereas in reality they are surface waves of low velocity. From equation (1), one can see that aliasing of k implies aliasing of V .

If a geophone array is sufficiently dense that wavenumber aliasing is avoided, then the fundamental mode Rayleigh waves can be used to obtain the three-dimensional distributions of velocity and attenuation. These distributions may be used with the horizontal component of the vector wavenumber, via ray tracing, to locate a source region of radiating microseisms (Liaw and McEvelly, 1979). Random directions of propagation (i.e. random vector wavenumbers) are characteristic of low velocity waves (Iyer and Hitchcock, 1975).

Liaw and Suyenaga (1982) detected high-

velocity body waves at Beowawe, but did not detect any body waves at Roosevelt Hot Springs. Liaw and McEvelly (1979) failed to find any body waves at Grass Valley. Thus, even when full f-k analysis is performed, high-velocity body waves may not be detected because they are not generated in all geothermal systems.

microearthquake (MEQ)

Microearthquakes, those having magnitudes between -2 and +3, frequently are closely related spatially to major geothermal systems. Accurate locations of these earthquakes can provide data on the locations of active faults that may channel hot water toward the surface (Ward, P.L. et al., 1969; Lange and Westphal, 1969; Ward, P.L. and Bjornsson 1971; Ward, P.L., 1972; Hamilton and Muffler, 1972).

From the arrival times of P and S waves at 8 to 12 geophones, spaced several kilometers apart, the source location, i.e. hypocenter of a seismic event, is calculated. The locus of many hypocenters may define a fault zone. Using first motion polarities, a fault-plane solution may indicate the type of motion along the fault, i.e. dip-slip vs. strike-slip, etc.

P- and S-wave velocities are retrievable from microearthquake data. Majer and McEvelly (1979) report locally high P-wave velocities in the production zone at The Geysers as determined from refraction surveys. Gupta et al. (1982) used microearthquake data to obtain regional P- and S-wave velocities for The Geysers. Usually, detailed velocity models, obtained from refraction surveys, are used to control the hypocenter determinations of the microearthquakes.

Measurement of either the absorption coefficient or a differential attenuation number called the "Q" may reveal the presence of exceptionally lossy materials in a reservoir due to fluid-filled fractures, or it may reveal the presence of low-loss materials due to steam-filled fractures or to silica- or carbonate-filled fractures. Majer and McEvelly (1979) found a shallow, high Q in the production zone at The Geysers from refraction and microearthquake surveys while they found a deeper, lower Q from the refraction survey. Majer (1978) reported that a refraction survey yielded high Q at Leach Hot Spring due to silica densification of sediments. Gertson and Smith (1979) found high Q over the geothermal system at Roosevelt Hot Springs, using refraction data.

The ratio, K, of P-wave to S-wave velocity may be estimated using a Wadati diagram in which S-P arrival times are plotted versus the P-wave arrival time at many different stations for a single event. From such a plot, a value for Poisson's ratio may be found. Nur and Simmons (1969) observed, experimentally, that fluid saturation in rocks leads to high values of Poisson's ratio ($\sigma \geq 0.25$) while dry rocks exhibit low values of Poisson's ratio ($\sigma < 0.20$). Thus

determination of Poisson's ratio in MEQ surveys can conceivably result in determining whether a geothermal reservoir is vapor or water dominated (Combs and Rotstein, 1975; Majer, 1978; Majer and McEvelly, 1979; Gupta et al., 1982).

Table 2 lists values of Poisson's ratio for several geothermal systems. At Baltazor,

TABLE 2
POISSON'S RATIO

GEOTHERMAL SYSTEM	AUTHOR	POISSON'S RATIO
Baltazor	Senturion Sciences	0.22 ?
Coso	Combs and Rotstein	0.16 Steam
Grass Valley	Beyer et al.	0.25 to 0.30 No Anomalies
McCoy	Lange	0.15 Dry or Siliceous Volcanic Fill >0.35 Saturated Alluvium
Roosevelt Hot Springs	Ward	0.25 ?
The Geysers	Majer	0.15 to 0.24 Prod. Zone
	Gupta et al.	0.13 to 0.16 Prod. Zone
Tuscarora	Nicholl and Lange	to 0.35 Saturated Alluvium

Senturion Sciences, Inc. (1977) reported a Poisson's ratio of 0.22. Combs and Rotstein (1975) reported a Poisson's ratio of 0.16 for Coso, a system dominated by vapor, at least in its shallower parts. In Grass Valley, Poisson's ratios of 0.25 to 0.30 were found over the entire region with no apparent anomalies (Beyer et al., 1976). Lange (1980) presented a plan map at McCoy showing contours of Poisson's ratios ranging from 0.15 over the central volcanic fill area, indicating dry or siliceous competent material, to greater than 0.35 over saturated alluvium. P-wave advances accompanied the higher values of Poisson's ratio while P-wave delays accompanied the lower values of Poisson's ratio at McCoy.

I have analyzed microearthquake swarm data apparently related to an east-west fault at Roosevelt Hot Springs and found a Poisson's ratio of 0.25. For the few events directly beneath the geothermal reservoir, a Poisson's ratio of 0.24 was found. Majer and McEvelly (1979) found Poisson's ratios ranging from 0.15 to 0.24 over the production zone at The Geysers, with higher values outside of it. The low Poisson's ratio in part corresponds to a decrease in P-wave velocity. Gupta et al. (1982), in a more extensive study at The Geysers, noted Poisson's ratios of 0.13 to 0.16 over the production zone and values 0.25 and higher outside of it. Nicholl and Lange (1981) reported Poisson's ratios as high as 0.35 due to saturated, fractured, basin fill at Tuscarora. While low values of Poisson's ratio seem to be associated with vapor-dominated systems at Coso and The Geysers, low values at McCoy would appear to be due to silification. A lack of an anomaly in Poisson's ratio at Roosevelt Hot Springs indicates that this parameter may not always contribute useful information about the reservoir.

Amplitude calibration of the microearthquake recording system is essential for meaningful microearthquake surveys. If microearthquake data are recorded on magnetic tape, then amplitude spectra are readily obtained, from which the source parameters seismic moment, stress drop, fault slip, and source radius may be estimated (Majer and McEvilly, 1979). The usefulness of these parameters in reservoir assessment requires extensive study.

For any of the above analyses of microearthquake data, a good model of the subsurface velocity distribution is required. Lack of good velocity control is a principal problem in analysis of MEQ data. Some geothermal systems, such as Roosevelt Hot Springs, have a generally low, episodic level of occurrence of microearthquakes. Swarms of earthquakes occur, but in the intervals between them, insufficient activity may preclude any of the foregoing analyses. Indeed, one can record passive seismic data for a two- or three-week period or longer and come to the conclusion that the geothermal system is unimportant since it is not seismically active.

In some geothermal systems, a fault-plane solution may not be meaningful because the microearthquakes occur not on a single fault, but on several intersecting faults. Caution must be exercised in accepting fault plane solutions in such cases. Further, the active fault(s) might not be related to the fault zone(s) serving as the reservoir.

teleseisms

Steeple and Iyer (1976a,b) found relative P-wave delays of 0.3 sec at stations in the west central part of the Long Valley caldera. Reasenberg et al. (1980) recorded relative P-wave delays of 0.2 sec at Coso. Iyer et al. (1979) found relative P-wave delays as large as 0.9 sec at The Geysers. Robinson and Iyer (1981) reported relative P-wave delays up to 0.3 sec at Roosevelt Hot Springs. Lange (1980) deduced P-wave delays as large as 0.25 sec at McCoy. Berkman and Lange (1980) recorded relative P-wave delays and advances at Tuscarora; the delays were attributed in part to alluvium and in part to zones of hydrothermal alteration, while the advances were attributed to silicification along fracture zones.

While one can speculate that relative P-wave delays are caused by partial melts or magmas, as may be the case at Coso, Long Valley, and The Geysers, they can also be caused by alluvium, alteration, compositional differences, lateral variations in temperature or locally fractured rock (Iyer and Stewart, 1977). Wechsler and Smith (1979) suggest that the P-wave delays found by Robinson and Iyer (1981) at Roosevelt Hot Springs may well be due to fluid-filled fractures or to a compositional change. Thus, teleseismic P-wave delay studies will not always produce conclusive results.

refraction seismology

The seismic refraction method has been used mainly as a geophysical reconnaissance method for mapping velocity distributions and, hence, faults, fracture zones, stratigraphy, and intrusions (Williams et al., 1975; Hill, 1976; Combs and Jarzabek, 1977; Majer, 1978; Ackerman, 1979; Gertson and Smith, 1979). The seismic refraction method does not give resolution of structure as well as does the seismic reflection method. Sentiment today calls for performing seismic refraction at the same time as seismic reflection, with little added cost. Some attempts have been made to map velocity and amplitude attenuation anomalies, of both P- and S-waves, coinciding with a geothermal system (Goldstein et al., 1978b). Beyer et al. (1976), Combs and Jarzabek (1977), Majer (1978), and Gertson and Smith (1979) found anomalous velocities and amplitudes of refracted waves passing through the reservoir region. "The limited results at hand are not easy to explain and are contradictory" (Goldstein et al., 1978b). On the other hand, the potential contribution to the understanding and mapping of the reservoir seems large.

reflection seismology

The seismic reflection method provides better resolution of horizontal or shallow-dipping layered structures than any other method and, hence, is invaluable in mapping stratigraphic geothermal reservoirs of the Imperial Valley type. However, where the structure becomes highly faulted or folded, diffraction of seismic waves occurs at sharp corners and makes the task of interpreting structure difficult.

"Conventional [reflection] seismic surveys appear to give good definition of Basin and Range border faulting and depths to the base of alluvial fill at Roosevelt Hot Springs, UT, Soda Lake, NV, San Emidio, NV and Grass Valley, NV". "One seismic line which crosses the Mineral Mountains at Roosevelt Hot Springs shows little obvious lithologic or structural information within the range itself, or within the reservoir, but substantial structural information along the range front. At Beowawe, extensive and varied digital processing was ineffective in eliminating the ringing due to a complex near-surface intercalated volcanic-sediment section. Majer (1978) found reflection data extremely useful in delineating structure in Grass Valley, NV" (Ward S.H., et al., 1981). At Soda Lake, "in 1977, Chevron obtained modern, 1200% CDP seismic reflection coverage (12 line miles, 48 channel, explosion). The seismic data yielded a complex NE-SW trending graben from the shore of Soda Lake passing south of Upsal Hogback. The reflectors dip to the southwest, consistent with a small basin over the gravity low. The maximum depths of reliable seismic data are governed by a thin basalt unit and vary from 2,400-4,000 ft" (Swift, 1979).

Zoback (1983) has nicely demonstrated the use of seismic reflection data in mapping the style of

Ward

initial faulting, infill and subsequent slumping and faulting in some basins in the province.

It is clear from the above remarks that high-resolution reflection seismology is capable of mapping faults, fractures, and stratigraphy in the vicinity of a geothermal system. In the Northern Basin and Range province, it will be useful in mapping range front faults and sedimentary stratigraphy. However, reflection seismology will seldom be of value in outlining reservoirs developed in volcanic, metamorphic, or igneous rocks. It may become almost useless when a volcanic cover or an intercalated volcanic cover is present as at Beowawe. The expense of modern multifold high resolution seismic surveys has deterred their use in geothermal exploration.

gravity

Gravity surveys are used in geothermal exploration as a relatively inexpensive means of obtaining structure and thickness of alluvium. "In the Basin and Range Province, large gravity lows are associated with the low-density basin fill. The gravity anomalies are used to estimate the thickness of the basin fill and the gross structure of the basin, including the location of the major normal faults that are important in understanding a geothermal system. Geothermal-related anomalies in the basins are most commonly residual gravity highs that are interpreted to reflect densification of porous sediments, structural highs, or anomalous geometry of fault zones" (Isherwood and Mabey, 1978). Gravity lows are sometimes found over siliceous magma bodies (Isherwood, 1976). At other times, gravity highs are expected due to rhyolite domes and hydrothermal alteration (Hochstein and Hunt, 1970; Macdonald and Muffler, 1972). Goldstein and Paulsson (1979), Berkman and Lange (1980), and Edquist (1981) found gravity particularly useful in mapping range-front normal faults in the Basin and Range province.

In the above, as in all other geophysical applications, the gravity method is hampered by ambiguity in interpretation. This ambiguity can be reduced by using drill hole, seismic refraction, or seismic reflection data for control. If these additional data are not available, then lateral and vertical resolution of geological features can be quite uncertain.

magnetics

Magnetic surveys, either airborne or ground, have been conducted at many geothermal prospects. Their use can be either for structural mapping or for mapping of changes in the magnetization of rocks caused by hydrothermal fluids. Magnetic anomalies in New Zealand geothermal fields have been interpreted as being due to a conversion of magnetite to pyrite (Studdt, 1964). Such an effect would, of course, remain in extinct hydrothermal systems.

"Opinion has been divided on the usefulness

of magnetic surveys in geothermal exploration (Cheng, 1970; Banwell, 1970). The magnetization at different rock units may be quite variable, especially in volcanic areas" (Palmason, 1975). "We examine the data from broad magnetic surveys as part of the effort to determine the regional setting ----. Some geothermal anomalies appear to relate to regional magnetic lineaments and zones suggestive of structure within the basement. Regional magnetic data can also be used to estimate the depth to the Curie isotherm by analysis of the spatial wavelengths of the field" (Isherwood and Mabey, 1978).

My own analysis indicates that magnetics have seldom been of major importance in assessing a geothermal reservoir in the Northern Basin and Range province. An outstanding exception appears to occur at Coso, where a magnetic low caused by destruction of magnetite from hydrothermal alteration draws attention, when correlated with other methods, to the heart of the reservoir. A similar magnetic low occurs over a part of the hot spring area at Long Valley (Plouff and Isherwood, 1980), and is interpreted by Kane et al. (1976) as due to magnetite destruction. At the Cove Fort-Sulphurdale KGRA, aeromagnetic data were important for an entirely different reason; structural controls on the potential reservoir could be deduced rather inexpensively from aeromagnetic data.

resistivity

Geothermal reservoirs frequently exhibit low resistivities due to high temperature, enhanced porosity, salinity of the interstitial fluid, and alteration of feldspars and other silicate minerals to clay minerals.

The geothermal use of the Schlumberger and Wenner arrays have been referenced in Banwell and Macdonald (1965), Hatherton et al. (1966), Macdonald and Muffler (1972), Meidav and Furgerson (1972), Zohdy et al. (1973), Arnorsson et al. (1975), Gupta et al. (1975), Stanley et al. (1976), Tripp et al. (1978) and Razo et al. (1980). Dipole-dipole arrays were used in surveys reported by Klein and Kauahikaua (1975), Jiracek et al. (1975), McNitt (1975), Garcia (1975), Beyer (1977), Fox (1978b), Ward, S.H. et al. (1978), Patella et al. (1979, 1980), Baudu et al. (1980), Smith (1980), Wilt et al. (1980a,b), Edquist (1981), and Mackelprang (1982). The bipole-dipole array was first used in geothermal exploration by Risk et al. (1970) and subsequently studied by Bibby and Risk (1973), Keller et al. (1975), Risk (1975a,b), Williams et al. (1975), Beyer et al. (1975), Stanley et al. (1976), Jiracek and Smith (1976), and Souto (1978). The bipole-dipole array achieved early success over broad areas of resistivity lows caused by hydrothermal alteration (Risk et al., 1970; Hohmann and Jiracek, 1979), but it has subsequently fallen into disfavor because of its failure to produce distinctive anomalies over geothermal systems lacking a broad surface manifestation (Dey and Morrison, 1977; Frangos and Ward, 1980).

The Schlumberger array is the most convenient one for depth sounding, i.e. estimation of the thicknesses and resistivities of the layers of a horizontally layered earth. The dipole-dipole array is used for continuous sounding-profiling, i.e. determination of both lateral and vertical variations in resistivity.

Several problems arise with the Schlumberger and dipole-dipole arrays (Ward and Sill, 1982):

- 1) Natural electric fields constitute noise for resistivity surveys and these result in slower productivity, higher costs, data of poorer quality, or in the extreme, no data at all. Modern data processing reduces but does not eliminate this problem.
- 2) Cultural features such as fences, powerlines, and pipelines redistribute current from the transmitter electrodes of the resistivity array; spurious resistivity anomalies result.
- 3) Strong noise voltages are present in the vicinity of powerlines. While notch filtering in the receiver will reduce this noise, it does not eliminate it.
- 4) Conductive overburden, generally in the form of porous alluvium or weathered bedrock, tends to prevent current from penetrating to the bedrock. Hence detection of bedrock features is less certain when overburden is present than when it is absent. When the overburden is of irregular resistivity, the geologic noise produced by the near-surface features may readily obscure the anomaly due to the target in the bedrock. Of course, this is common to all geophysical methods. Anomalies due to geological heterogeneities of no geothermal significance can also obscure, or partly obscure, the anomaly due to a geothermal system. In the Northern Basin and Range province these usually arise in Quaternary alluvial valley fill, and in salt playas.
- 5) Much geothermal exploration is done in mountainous terrain where topography can produce spurious resistivity anomalies. In a recent study, Fox et al. (1980) showed that a valley can produce a large, spurious resistivity low which could easily be misinterpreted as evidence for a buried conductor. Similarly, they showed that a hill can produce an apparent resistivity high.

In general, topographic effects are important where slope angles are 10° or more for slope lengths of one dipole or more. The solution to the problem is to include the topographic surface in numerical models used for interpretation.

- 6) The practical limitation on the depth of exploration of the Schlumberger array is the large separation needed between current electrodes. A large current electrode

separation usually means that lateral resistivity variations outside the array will affect the measurements, thus rendering interpretation difficult. The dipole-dipole array minimizes this difficulty, but introduces a new one; a dipole-dipole array is sensitive to shallow lateral resistivity variations beneath the array whereas a Schlumberger array is not so sensitive (Palmason, 1975).

The depth of exploration of resistivity arrays is difficult to assess, but values in the range 0.5 km to 1.5 km are common.

controlled source electromagnetic (CSEM) and controlled source audiomagnetotellurics (CSAMT)

The transmitter for a CSEM method consists of either a loop of wire or a grounded bipole. Either source is energized by one or more frequencies in the audio frequency range or by a step current. One or more orthogonal components of magnetic and electric fields are recorded by a receiver located at a distance from the transmitter. The resistivity of the earth is measured by noting the phase and amplitude relationship of the voltage in the receiver to the current in the transmitter. This relationship is termed the impedance and is defined as

$$Z = \frac{V(t + \phi)}{I(t)} \quad , \quad (2)$$

in which both the voltage V and the current I are functions of time. The received voltage is shifted in phase by ϕ relative to the current in the transmitter.

Keller (1970) reviewed the applications of active and passive electromagnetic methods in geothermal exploration. His article constituted a baseline for reference to controlled-source electromagnetic methods (CSEM) in geothermal environments. Subsequent to Keller's review, a number of articles have appeared which illustrate the success and failure of these methods in geothermal exploration. Included are the articles by Lumb and MacDonald (1970), Jackson and Keller (1972), Keller and Rapolla (1974), Jacobson and Pritchard (1975), Morrison et al. (1978), Tripp et al. (1978), Kauahikaua (1981), Wilt et al. (1980c,d), Wilt et al. (1981a,b), Goldstein et al. (1982), and Keller et al. (1982).

The method will suffer all of the problems listed under the *resistivity* section above.

CSAMT is a subset of CSEM, or is a subset of AMT in which the transmitter is a grounded bipole. It is the only CSEM method that does not utilize a loop source. Two orthogonal, horizontal components of electric and magnetic field are measured (as in magnetotellurics). It offers advantages over resistivity methods in that it is faster and suffers less from the effects of lateral resistivity variations when providing

Ward

sounding information (Ward, S.H., 1983). The method is used for combined sounding-profiling.

scalar audiomagnetotellurics (AMT)

The AMT method utilizes natural electromagnetic fields in the 10 Hz to 20 kHz band. These natural fields arise in atmospheric lightning discharges worldwide. Hoover et al. (1976), Hoover and Long (1976), Hoover et al. (1978), and Long and Kaufman (1980) described reconnaissance AMT surveys. Keller (1970), Whiteford (1975), Williams et al. (1975), Isherwood and Mabey (1978), and Jackson and O'Donnell (1980) have also reported on its use in geothermal exploration.

Two orthogonal magnetic-field components and two orthogonal electric-field components are measured and the scalar apparent resistivities are computed from the formulae

$$\rho_{a_1} = 0.2T \left| \frac{E_x}{H_y} \right|^2, \quad (3)$$

and

$$\rho_{a_2} = 0.2T \left| \frac{E_y}{H_x} \right|^2, \quad (4)$$

in which the electric field is measured in mv/km and the magnetic field in gammas; T is the period in seconds. The apparent resistivities are then given in ohm-meters.

The method suffers from all of the problems listed under resistivity, except natural field noise. It exhibits two other problems, however. The first is that the natural fields occasionally are too weak to obtain useful information. The second, and far more important, is that the simple formulae of equations (9) and (10) are totally inadequate for interpretation in two- and three-dimensional terrains, in which the tensor AMT method should be used. However, the scalar AMT method has proven useful for reconnaissance surveys of the type performed by USGS personnel who have fostered its use in this country. Very little application of the method has been made by industry, probably because equipment for the method has not been available commercially.

The CSAMT method is a substantial improvement over scalar AMT insofar as the direction of the inducing fields can be controlled, thus simplifying interpretation in two- and three-dimensional environments. The uncertainty over the strength of the fields used in AMT disappears when CSAMT is used.

MT/AMT

The tensor MT/AMT method utilizes natural electromagnetic fields in the 10^{-4} Hz to 10^4 Hz band. Below about 1 Hz these fields arise in geomagnetic perturbations brought about by

interaction of the solar wind with the main geomagnetic field. Above 1 Hz the natural fields arise in atmospheric discharges worldwide. Three components of magnetic field and two components of electric field are measured. Data acquisition, processing and interpretation must be treated in highly sophisticated ways whose theory is beyond the scope of this paper.

Papers describing application of the tensor MT/AMT method in geothermal areas include Hermance et al. (1975), Hermance and Pedersen (1977), Stanley et al. (1977), Goldstein et al. (1978a), Morrison et al. (1979), Dupis et al. (1980), Gamble et al. (1980), Musmann et al. (1980), Ngoc (1980), Wannamaker et al. (1980), Aiken and Ander (1981), Berktoold (1982), Berktoold and Kemmerle (1982), Goldstein et al. (1982), Hutton et al. (1982), Martinez et al. (1982), Stanley (1982) and Wannamaker et al. (1983). A comprehensive review of data acquisition, processing, and interpretation for the method, plus a full discussion of the problems it encounters in geothermal exploration, has been prepared by Ward, S.H. and Wannamaker (1983). In the following I seek to summarize their review of these problems.

1) Source dimensions

All of the formulation for interpretation of MT/AMT data over one-, two-, or three-dimensional earths assumes that the MT fields are propagated as plane waves. This assumption was the source of much controversy in the early days of MT, but Madden and Nelson (1964) showed that the field is usually plane wave at frequencies greater than 10^{-3} Hz in mid latitudes.

At frequencies below 1 Hz, the primary concern appears to be whether or not the fields due to equatorial and auroral electrojet ring currents in the E-layer of the ionosphere can be treated as planar. Hermance and Peltier (1970) and Peltier and Hermance (1971) have studied the effects of such ring currents. They conclude that in conductive environments, the plane-wave assumption is valid in the frequency range 10^{-4} Hz to 1 Hz. However, significant errors can occur at frequencies less than 10^{-1} Hz in areas where high resistivities are encountered if measurements are made within 500 km of the position vertically beneath the electrojet.

At frequencies above 1 Hz, the proximity of lightning discharges becomes important. Bannister (1969) studied the fields radiated from a vertical electric dipole over a homogeneous earth and concluded that the plane-wave assumption is valid for distances greater than seven skin depths from the source.

If however, the plane wave assumption is not valid, then the extra field components associated with non-planar waves will be processed so as to produce bias in MT estimates.

2) Random noise

Random noise may arise in a) the electrodes for E-field measurement via chemical disequilibrium, b) movement of the E-field wires in the earth's magnetic field when wind agitates them, c) movement of the H-field sensors in the earth's magnetic field due to wind or seismic activity, d) microphonics in the H-field sensors due to any motion, e) thermal noise in the E- and H-field preamplifiers, f) quantization noise in A/D converters, g) non-linear behavior of the total recording system, h) sporadic departure from plane wave propagation, and i) sporadic cultural noise due to power lines, telephone lines, rail electrification, pipeline corrosion protection, radio interference, and the power sources in the recording instrumentation. The processing system must be designed to minimize, evaluate, and place statistical limits on errors introduced into MT transfer functions by random noise.

3) Systematic noise

Most of the noise sources described in 2) above are also capable of introducing systematic noise into estimates of the MT transfer functions. As Stodt (1983) points out, the systematic noise must be treated independently of the random noise in any statistical evaluation of noise in MT data. As a result of systematic noise, biased estimates of the MT transfer functions result. To attempt to eliminate this problem, the use of a remote reference has become common practice (Gamble et al., 1979a,b).

4) Geological noise due to overburden

In areas where there is irregular conductive overburden, current channeling into a patch of deeper or more conductive overburden will produce anomalies even to the lowest frequencies. Unless these anomalies are interpreted via 2D or 3D modeling, they can be mistaken for deep-seated features. Wannamaker (1983a) illustrates these effects.

5) Topography

The effect of topography on the results of an MT survey may be significant. Anomalous secondary electric and magnetic fields result. Topography must be included in the numerical modeling used in interpretation to prevent topographic effects from being interpreted as subsurface effects.

6) Depth of exploration and detectability

Depth of exploration is often stated to be one skin depth, δ , where

$$\delta = 500 \sqrt{\frac{2}{\omega \mu \sigma}} \quad (5)$$

and σ = the conductivity of the earth, ω is the angular frequency of the signal under consideration, while μ is magnetic permeability. This simplification is misleading, because noisy data

or surface geological noise can obscure the responses of deep bodies. However, with care in both data acquisition and interpretation, depths of exploration well in excess of 100 km can be achieved for infinite interfaces.

For 2D or 3D bodies, depth of exploration can be considerably less. Newman et al. (1983) have explored the possibility of detecting deep magma chambers with MT. If the magma chamber is electrically connected to a highly conducting layer below it, the magma chamber probably will not be detected. On the other hand, if the basal half-space is resistive or if the earth is not layered, the magma chamber is more readily detected.

self-potential

In principle, self-potential surveys are very simple: two non-polarizing electrodes, a length of wire, and a D.C. voltmeter are all the equipment needed to perform a survey. However, much attention must be paid to details if the desired reproducibility of ± 5 mV is to be achieved. Two methods of moving the electrodes along the traverse line are used; they are *leapfrog* and *long wire* methods. In the former, the back electrode is leapfrogged past the forward electrode for each move. Only a short wire is required. In the long wire method, the back electrode is left fixed and the forward electrode is moved farther and farther away. A long length of wire is then required.

Corwin and Hoover (1979) reviewed the self-potential method in geothermal exploration. Other pertinent references to the use of the self-potential method in geothermal exploration include Zohdy et al. (1973), Corwin (1975), Anderson and Johnson (1976), DeMouilly and Corwin (1980), Hoover (1981), Sill (1982a,b,c), and Sill (1983a,b). Noise in self-potential surveys arises in telluric currents, electrode drift, topographic effects, variations in soil moisture, cultural noise, vegetation potentials, and electrokinetic potentials due to running surface water (Ward, S.H. and Sill, 1982). The cultural noise sources include: radiated fields from power lines, telephone lines, and electrified rails; corrosion potentials from pipelines, fences, and well casings; and spurious potentials from corrosion protection systems associated with pipelines. The following discussions of these problems follow from Ward, S.H. and Sill (1982).

1) Telluric currents

Time-varying voltages induced in the earth by the geomagnetic field, of frequencies within the passband of the voltmeter, may reach several hundred mv/km over resistive terrain (Keller and Frischknecht, 1966). These time-varying voltages constitute noise which inhibits repeatability of a measurement of the steady-state self-potentials. The magnitude of this noise is proportional to the separation between the two electrodes and accordingly is largest for the *long wire* method.

Ward

If telluric noise is dominant, then the *leapfrog* method is preferred.

2) Electrode drift

A voltage will be measured across an electrode pair if either or both of the electrodes are not in equilibrium. Departure from zero electrode potential will occur if the electrolyte in the non-polarizing electrode is diluted or contaminated by groundwater or if there is a temperature differential between the two. These effects will vary with time, moisture content of the soil, and ambient temperature. Repeated checks of drift must be made periodically by placing both measuring electrodes in a bath of electrolyte solution and connecting them in order to establish equilibrium. Leapfrog surveys usually result in irregular electrode drift so that long wire surveys are preferred where electrode drift is the dominant source of noise.

3) Topography

Topographic relief will distort self-potentials and this effect must be taken into account in interpretation of field data. Another topographic effect is due to the movement of shallow groundwater. More negative potentials are sometimes correlated with an increase in elevation with observed gradients as large as -6 mv/m (Hoover, 1981).

4) Variations in soil moisture

As noted by Corwin and Hoover (1979), variations in soil moisture often give rise to self-potential variations, with the electrode in the wetter soil usually becoming more positive. Watering of electrodes to improve electrical contact can produce the same effect, but even worse, electrokinetic potentials are generated as the water moves through the soil. Electrode watering should be avoided for self-potential surveys in geothermal areas because the geothermal anomalies frequently are small, so that reducing noise to a minimum becomes essential.

5) Cultural noise

We have noted above some sources of cultural noise encountered when performing S.P. surveys. All of them can lead to potentials, which vary with time, while corrosion potentials and potentials from corrosion protection systems additionally will produce spurious anomalies. Results from self-potential surveys in a developing or developed geothermal field will be quite different from results obtained before well casings were installed.

6) Vegetation potentials

Trees, shrubs, and grasses produce potentials which are commonly of order 10 mv. A technique used to reduce this noise, and noise due to varying soil moisture, involves making five measurements in a star about each observation

point. Four readings are offset about 3 m north, south, east, and west of the central point. The five readings are then averaged.

7) Electrokinetic potentials from moving non-thermal water

Potentials generated by the flow of non-thermal surface and subsurface water, i.e. electrokinetic potentials, constitute a noise source in geothermal exploration, and may be a major cause of topographic noise (Corwin and Hoover, 1979) because of water flowing downhill beneath the surface.

tellurics

"--- the telluric method is mainly suitable for reconnaissance of horizontal resistivity variations. It is based on the assumption that telluric currents flowing in extensive sheets are affected by lateral variations in the resistivity structure, which can be caused, for example, by variations in geological structure or by hydrothermal systems. The method requires the simultaneous measurement of the telluric electric field at two stations. From the ratio of the amplitudes of the electric field at the two stations, inferences may be drawn about variations in the underlying resistivity structure. By keeping the base station fixed and moving a field station about, one can thus map resistivity variations in a qualitative way." (Palmason, 1975)

The method has been used in geothermal exploration by Beyer (1977), Isherwood and Mabey (1978), Jackson and O'Donnell (1980), and others. It appears to be a convenient method for regional surveys in order to detect areas worthy of more detailed exploration by resistivity methods (Palmason, 1975).

The method suffers from a number of problems which have already been described under MT/AMT as follows: random noise, geological noise due to overburden, lack of resolution, and effects of topography. Its worst problem is that it is a semi-quantitative method at best. However, Beyer (1977) advocated its use for northern Nevada because of its simplicity, low cost, and ease of interpretation.

heat flow

I will not dwell on the problems encountered with temperature gradient and heat flow measurements since that subject will be treated by D. D. Blackwell in this volume.

5.0 ILLUSTRATIVE RESULTS

Long Valley

Figure 3 presents the generalized geology of the Long Valley caldera, California from Bailey et al. (1976). I wish to draw attention to the resurgent dome depicted by early rhyolite flows,

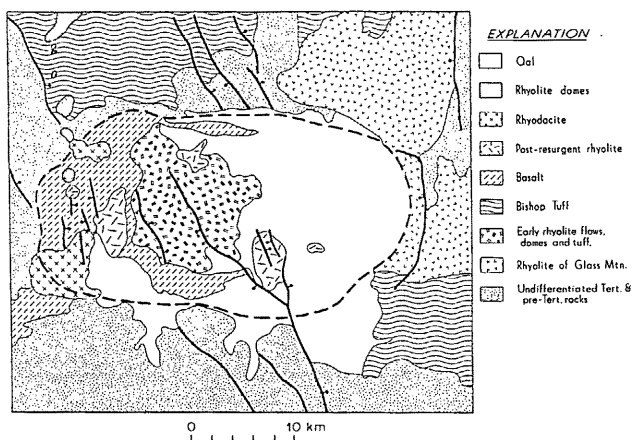


Figure 3. Generalized geologic map of Long Valley caldera (from Bailey et al., 1976).

domes and tuff, and to the medial graben transecting the resurgent dome in a northwest direction. Heat flow has been measured in 29 drill holes from 50 to 300 m deep within the caldera. An additional 11 holes outside, but adjacent to the caldera, were drilled to 150 to 300 m and heat flow was measured in them (Lachenbruch et al., 1976a,b). Temperature at 10 m depth for the drill holes within the caldera is contoured in Figure 4. An outline of the area of

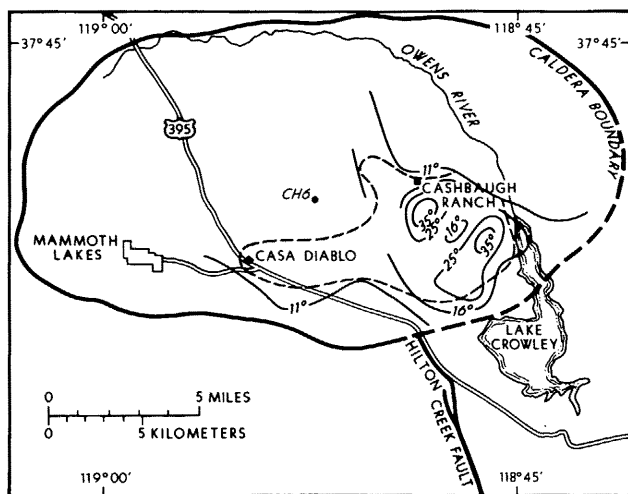


Figure 4. Temperatures at 10 m depth, Long Valley caldera (from Lachenbruch et al., 1976a). The area of hot spring activity is superimposed as a dashed line.

hot springs is included in the figure. There is an obvious correlation between shallow temperature measurements and hot spring activity. Note that drill hole CH6 is the only drill hole within the resurgent dome; it was isothermal at 111°C from about 150 m to 300 m. Ground deformation, enhanced seismicity, and fumarolic activity in the last 5 years, and particularly in the last 2 years, testifies to a modern increase in tectonism and presumed magmatism. The enhanced seismicity appears to lie south of the resurgent dome. Figure 5 shows earthquakes occurring in the region

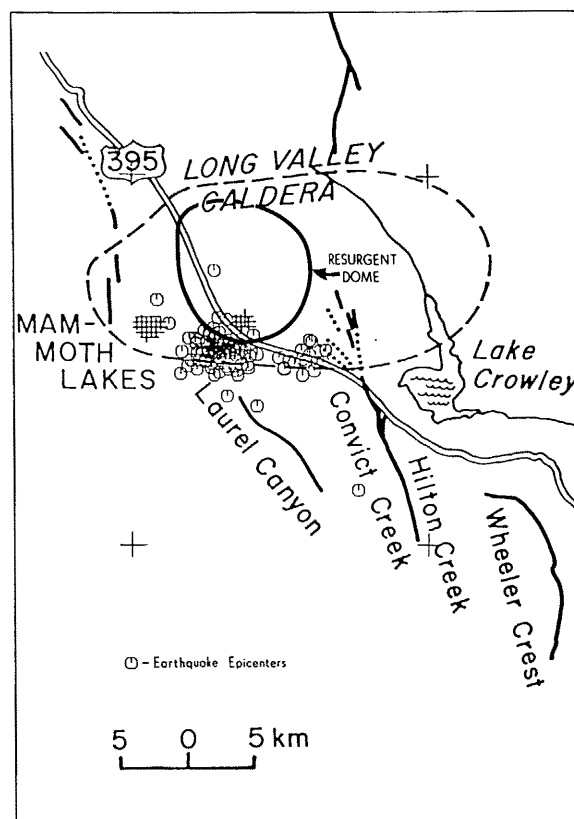


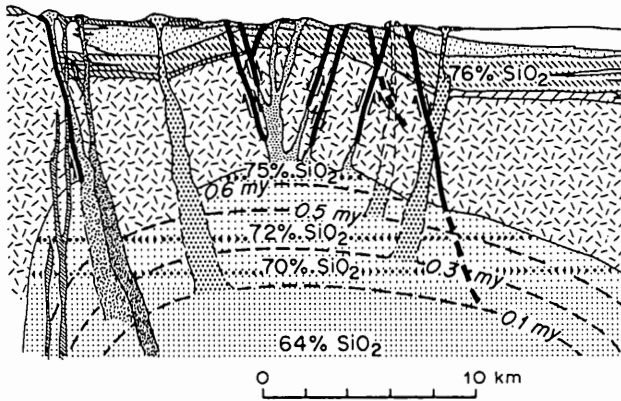
Figure 5. Mammoth Lakes earthquakes July, 1981, (from Miller et al., 1982).

in July, 1981 (from Miller et al., 1982). A tongue of magma is believed to have risen to within 2 to 3 km of the surface beneath the epicenter of the earthquakes shown in Figure 5. Many more earthquakes than are shown in this figure have occurred in the last several years. Ryall and Vetter (1983) indicate that such earthquakes occur as far as 7 km north from the south rim of the caldera and as much as 15 km south of it.

The resurgent dome has bulged upward in recent years (Miller et al., 1982). Steeples and Iyer (1976a) reported a zone of significant

teleseismic P-wave delay, beneath much of the resurgent dome. Hill (1976) found evidence for the roof of a magma chamber in late P-wave arrivals which he identified as reflections from a low-velocity horizon at a depth of 7 to 8 km. Ryall and Ryall (1981) found P- and S-wave attenuation which they attributed to a magma chamber beneath the Long Valley caldera. Ryall and Vetter (1983) reported on the absence of S waves at stations north of the caldera, for events occurring south of it; a magma chamber within the caldera is postulated to account for these observations; the postulated magma chamber roughly coincides with the area of hot spring activity shown in Figure 4.

Figure 6 (from Bailey et al., 1976) portrays a schematic east-west cross section through Long



SCHEMATIC CROSS-SECTION

EXPLANATION

- ALLUVIUM, GLACIAL DEPOSITS, AND CALDERA FILL
- HOLOCENE, RHYOLITE-RHYODACITE
- LATE BASALTIC ROCKS
- RIM RHYODACITES
- MOAT RHYOLITES
- EARLY RHYOLITES { TUFFS: FINE DOTTED
FLOWS: COARSE DOTTED
- BISHOP TUFF
- RHYOLITE OF GLASS MTN. { DOME FLOWS: FINE LINED
TUFFS: COARSE LINED
- TERTIARY VOLCANIC ROCKS
- JURASSIC-CRETACEOUS GRANITIC ROCKS

Figure 6. Schematic east-west cross section through Long Valley caldera and its underlying magma chamber (from Bailey et al., 1976).

Valley caldera. As noted by Muffler and Williams (1976) and by Sorey et al. (1978) the prevalent USGS view has been that the geothermal reservoir lies in the Glass Mountain rhyolites and the Bishop Tuff, the two earliest volcanics to infill the caldera.

The caldera fill above the Bishop Tuff consists of a variety of rhyolitic flows and tuffs, rhyodacitic flows, basaltic flows, and in the eastern half, lake, marsh, and periglacial sediments. These rocks are considered to be impermeable except along faults and hence do not form part of the reservoir (Muffler and Williams, 1976). The gravity data of Kane et al. (1976) and the seismic section of Hill (1976) have been used by USGS personnel to define the volume of the proposed reservoir to be 450 km³.

Resistivity and AMT surveys mapped two near-surface low-resistivity zones associated with hydrothermally altered rocks and/or hot saline water (Hoover et al., 1976; Stanley et al., 1976).

Nine geothermal test wells were drilled near Casa Diablo and one well was drilled about 5 km farther east by Magma Power Company. One well was drilled by Republic Geothermal, Inc. about 4 km south of Cashbaugh Ranch. Union Geothermal drilled two wells, one near a clay pit in the north end of Little Antelope Valley, and one near Casa Diablo. All of the above wells have failed to discover a high temperature geothermal reservoir. None of the wells has been sited near the center of the resurgent dome, especially where transected by the medial graben. The self-potential data of Figure 7 draw attention to the whole of the area of the resurgent dome. W.R. Sill (pers. comm.) advises that cold water inflow from the west-northwest and outflow in the area of the thermal springs could account for the dipolar S.P. pattern. The cold water inflow could readily mask the center of a deep convective hydrothermal system possibly occurring beneath the resurgent dome.

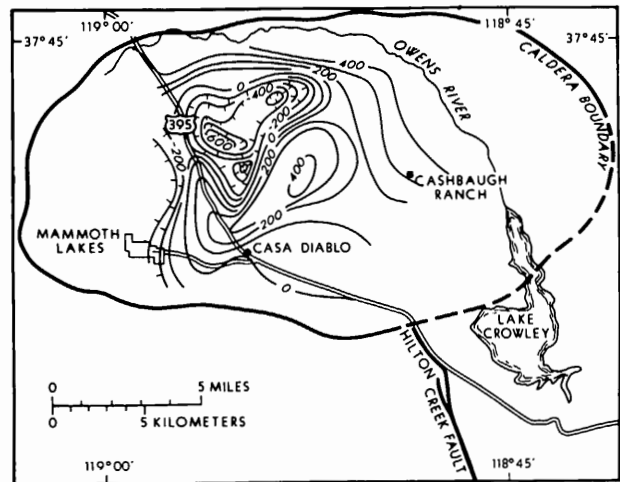


Figure 7. Self-potential map of Long Valley. (after Anderson and Johnson, 1976).

The magma chamber, as determined by the teleseismic P-wave delays, lies directly beneath the resurgent dome. The cross section of Bailey et al. (1976), Figure 6, schematically illustrates a potential fault-controlled reservoir within and below the caldera infill and beneath the resurgent

dome. The problem with this notion is that for calderas elsewhere (e.g. Baca) the resurgent dome may be a poor exploration target because of lack of permeability (D.L. Nielson, 1983, pers. com.).

The resistivity and AMT data draw attention to areas south and east of the resurgent dome, while the aeromagnetic data draw attention to a small region of magnetite destruction around Casa Diablo. Surface manifestations of hydrothermal activity are abundant in these areas. I have downgraded shallow heat flow, magnetics, resistivity, and AMT at this geothermal prospect because drilling to date has not found a commercial reservoir where these methods have produced anomalies. I have up-graded self-potential on the speculation that a commercial reservoir may be found in the central part of the resurgent dome. Gravity and seismic refraction were very useful in delineating the Glass Mountain Rhyolites and Bishop Tuff in the floor of the caldera and were instrumental in quantifying the USGS model of the reservoir. Teleseismic P-wave delays, S-wave attenuation, and S-wave delays seem quite important and reliable indicators of a magma chamber in this setting. Earth noise so far has failed to help us delineate a reservoir. The earthquake epicenter map of Figure 5 may be delineating an east-west fracture zone up which a tongue of lava is believed to have recently intruded. What bearing this would have on a commercial high-temperature reservoir is unknown at present.

Evaluating geophysical methods at Long Valley is a questionable pursuit at present because we do not know where a commercial high-temperature reservoir is located if, indeed, it exists. However, the analysis has been made on the assumption that either the Bishop Tuff or the deeper parts of the medial graben will prove to be commercial. Note that the recent bulging of the resurgent dome would indicate that the medial graben may now be in extension.

Coso

Figure 8 portrays the generalized geology of the Coso geothermal area (Hulen, 1978). "The oldest rocks exposed at Coso are intermediate to mafic metamorphic rocks of uncertain age intruded by dikes and pods of quartz latite porphyry and felsite, and by a small stock of Late Cretaceous (?) granite. These rocks are locally overlain by Late Cenozoic volcanic rocks, which include the domes, flows, and associated pyroclastic deposits of the Coso rhyolite dome field."

"Principal structures in the geothermal area are older high-angle faults of uncertain displacement trending northwest, west-northwest, and east-northeast, and younger high-angle faults with a normal component of displacement trending north-northwest, north-northeast, and (subordinately) northeast. Active surface thermal phenomena and hydrothermal alteration are concentrated along the younger northerly-trending faults, especially where these faults intersect

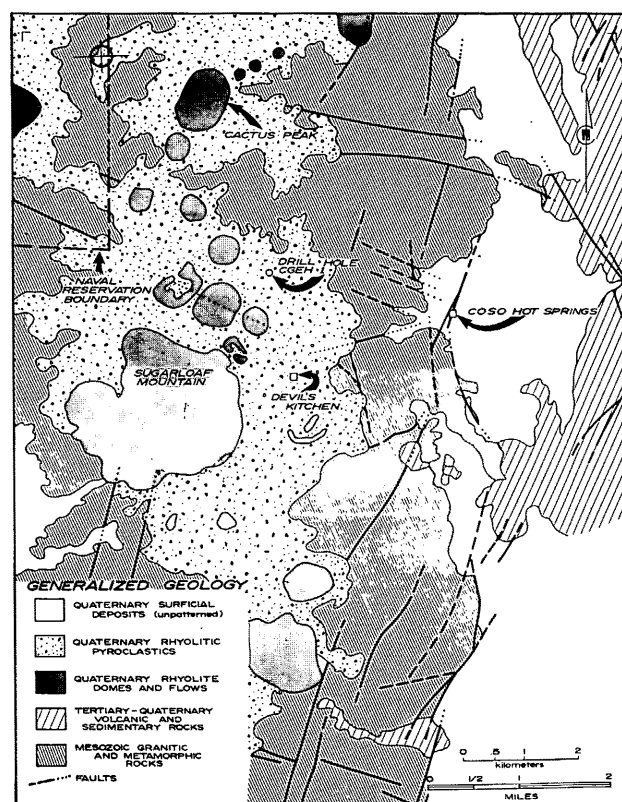


Figure 8. Generalized geologic map of Coso Hot Springs KGRA. (after Hulen, 1978).

older structures. Deep thermal fluid flow at Coso will be controlled entirely by structural permeability developed in otherwise tight and impermeable host rocks."

"Surface alteration at Coso is of three main types: (1) clay-opal-alunite alteration, (2) weak argillic alteration, and (3) stockwork calcite veins and veinlets, which are locally associated with calcareous sinter. Clay-opal-alunite and weak argillic alteration are typically developed around active thermal emissions. These are almost entirely restricted in distribution to an east-northeast-trending belt roughly one mile in width and four miles in length. Calcareous alteration is much more widely distributed, but is confined to a broad zone of anomalous geophysical response interpreted as evidence for a concealed geothermal reservoir" (Hulen 1978).

A low level aeromagnetic map of the area is given in Figure 9 (after Fox, 1978a). The anomalous low which extends southeast from Devil's Kitchen and Coso Hot Springs is attributed, in part, to magnetite destruction by hydrothermal fluids (Fox 1978a).

A resistivity low mapped in the vicinity of Devil's Kitchen and Coso Hot Springs (Figure 10) is attributed, in part, to hydrothermal alteration and hot brines in fractures (Fox, 1978b). The

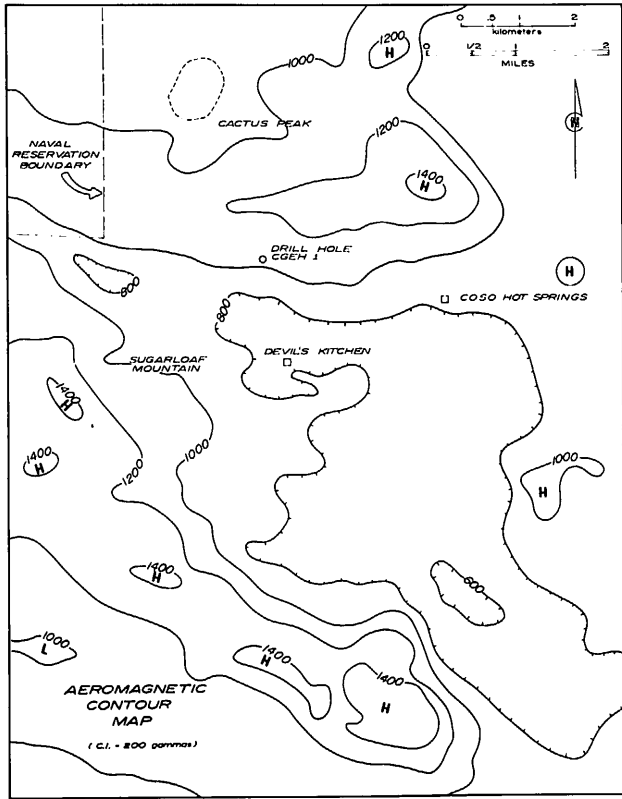


Figure 9. Aeromagnetic residual total field intensity contour map of Coso. (after Fox, 1978a).

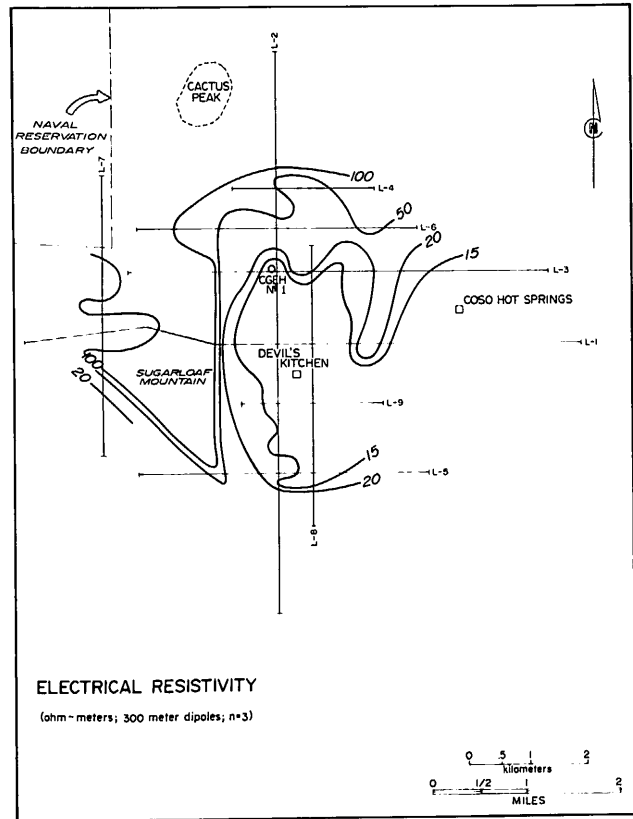


Figure 10. Third separation apparent resistivity map of Coso obtained with 300 m dipole-dipole survey. (after Fox, 1978b).

dipole-dipole array with 300 m dipoles was used. Much the same zone of low resistivity was mapped by AMT earlier. Figure 11 displays the 7.5 Hz AMT apparent resistivity map produced by Jackson and O'Donnell (1980). A telluric J-value map, also produced by these authors, was not nearly as effective in depicting the area of interest around Devil's Kitchen and Coso Hot Springs.

No one of the above geophysical techniques by itself created a confined target for drilling. However, when the resistivity and aeromagnetic data were combined with shallow heat flow data and a map of the hydrothermal alteration, then a target in the general vicinity of Devil's Kitchen was clearly delineated as shown in Figure 12 (from Hulen, 1978). Prior to Hulen producing Figure 12, drill hole CGEH-1 was drilled as a dry hole. Subsequently, all six wells have been producers. The importance of overlapping several data sets prior to spotting the first well on a geothermal prospect is vividly illustrated by this example.

I have not found that earth noise, teleseisms, gravity, or tellurics have contributed much to understanding the reservoir. The earth noise data of Figure 13 (after Combs, 1980) could be aliased and noise power peaks near Coso Hot Springs may, in part, be due to Rayleigh waves in deep alluvium. The teleseismic evidence of a 0.2

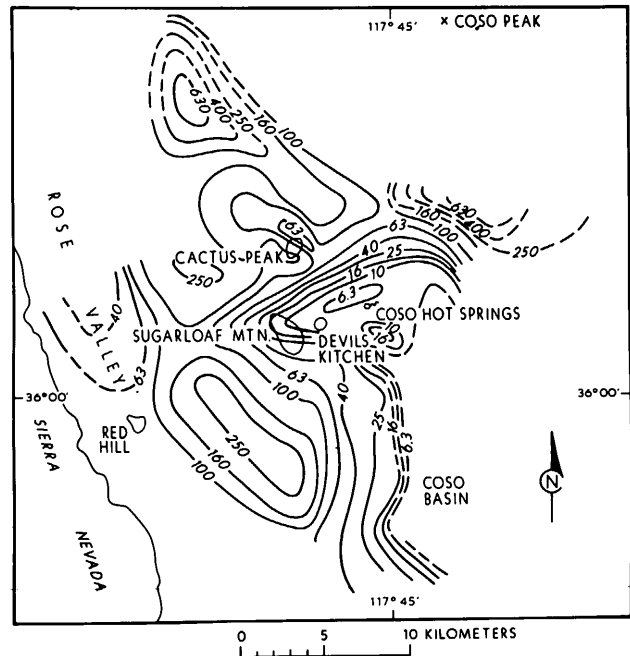


Figure 11. The 7.5 Hz AMT apparent resistivity map of Coso. (after Jackson and O'Donnell, 1980).

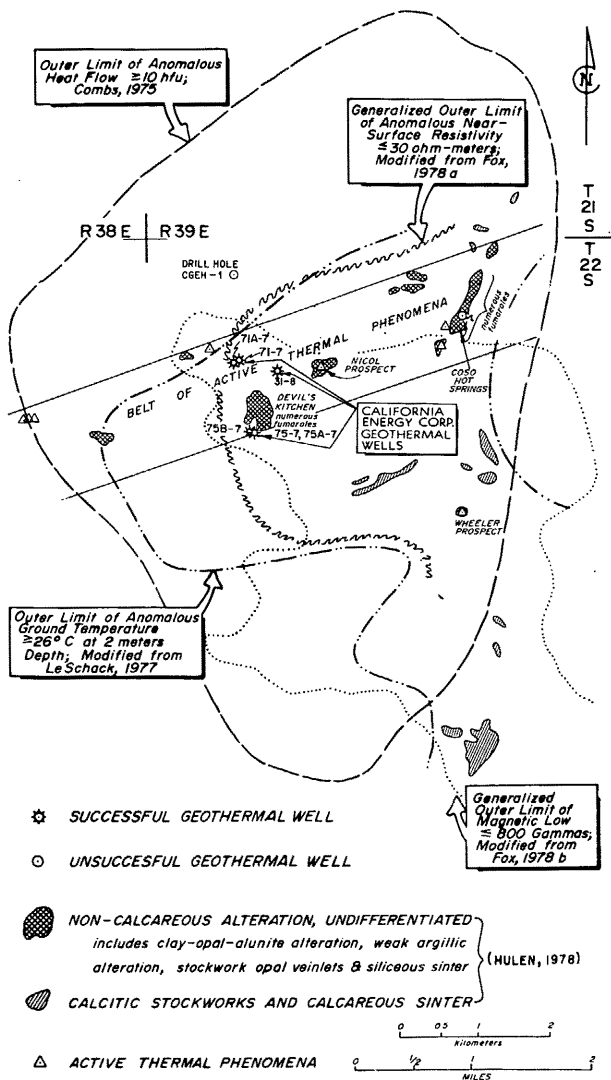


Figure 12. Generalized alteration and geophysical map. (after Hulén, 1978).

sec P-wave delay is not convincing. The gravity data are relatively featureless in the region between Devil's Kitchen and Coso Hot Springs. The telluric data are too much affected by Coso Basin to be of good quality in the region of interest. Microearthquakes were highly scattered throughout the area as Figure 14 illustrates (after Combs and Rotstein, 1975). No major faults were deduced from the data by Combs and Rotstein. However, the important contribution of microearthquakes at this prospect lies in calculation of a Poisson's Ratio

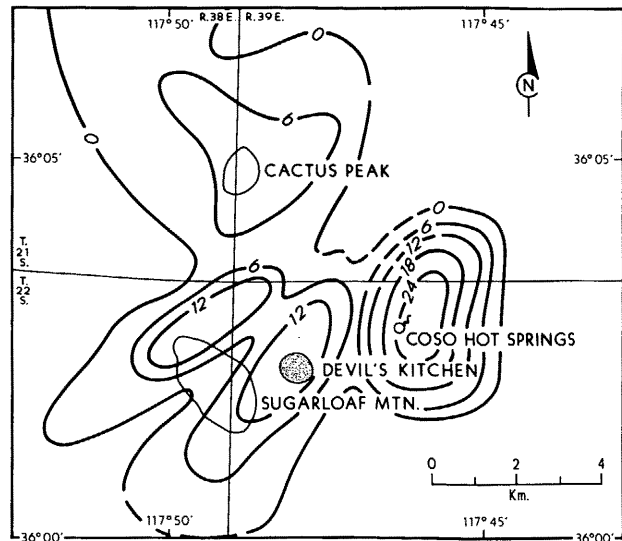


Figure 13. Earth noise power expressed in decibels, Coso. (after Combs, 1980).

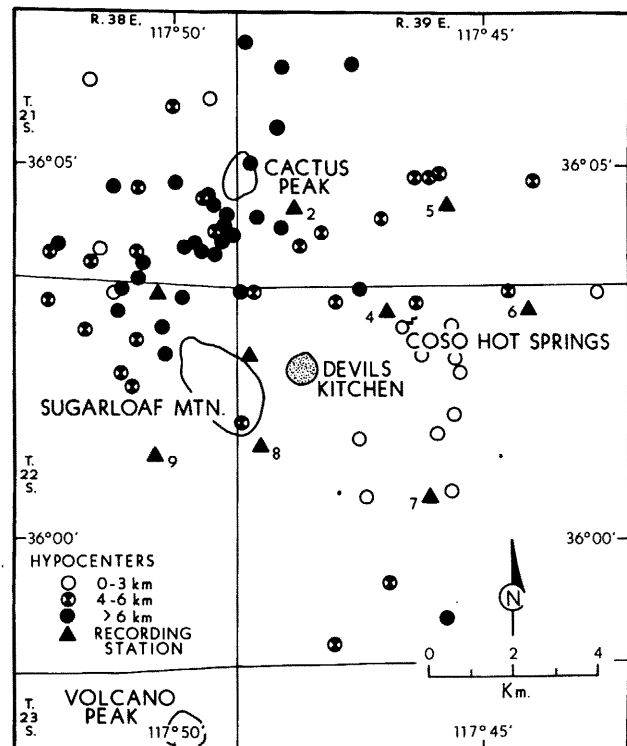


Figure 14. Epicenters of 78 microearthquakes associated with the Coso geothermal area. (from Combs and Rotstein, 1975).

of 0.16, an indicator of a vapor-dominated system. Based on current knowledge, the shallower parts of the reservoir appear to be vapor dominated (J.A. Whelan, 1983, pers. com.). The refraction method was useful at Coso in suggesting "the existence of a localized body of low velocity material at depth, possibly a magma chamber." (Combs and Jarzabek, 1977)

Roosevelt Hot Springs

Figure 15 portrays the generalized geology at the Roosevelt Hot Springs KGRA (after Sibbett and Nielson, 1980). From Ross et al. (1982) we extract the quotations which follow.

"The geothermal system is a high-temperature water-dominated resource, and is structurally

controlled with permeability localized by faults and fractures cutting plutonic and metamorphic rocks."

"The oldest unit exposed in the area of the geothermal system is a banded gneiss which was formed from regionally metamorphosed quartzofeldspathic sediments." "The rock is compositionally heterogeneous and contains thick sequences of quartzofeldspathic rocks. The unit also contains metaquartzite and sillimanite schist layers which have been differentiated in the more detailed geologic study (Nielson et al., 1978)."

"The Mineral Mountains intrusive complex is the largest intrusive body exposed in Utah. Potassium-Argon dating and regional relationships suggest that the intrusive sequence is middle to

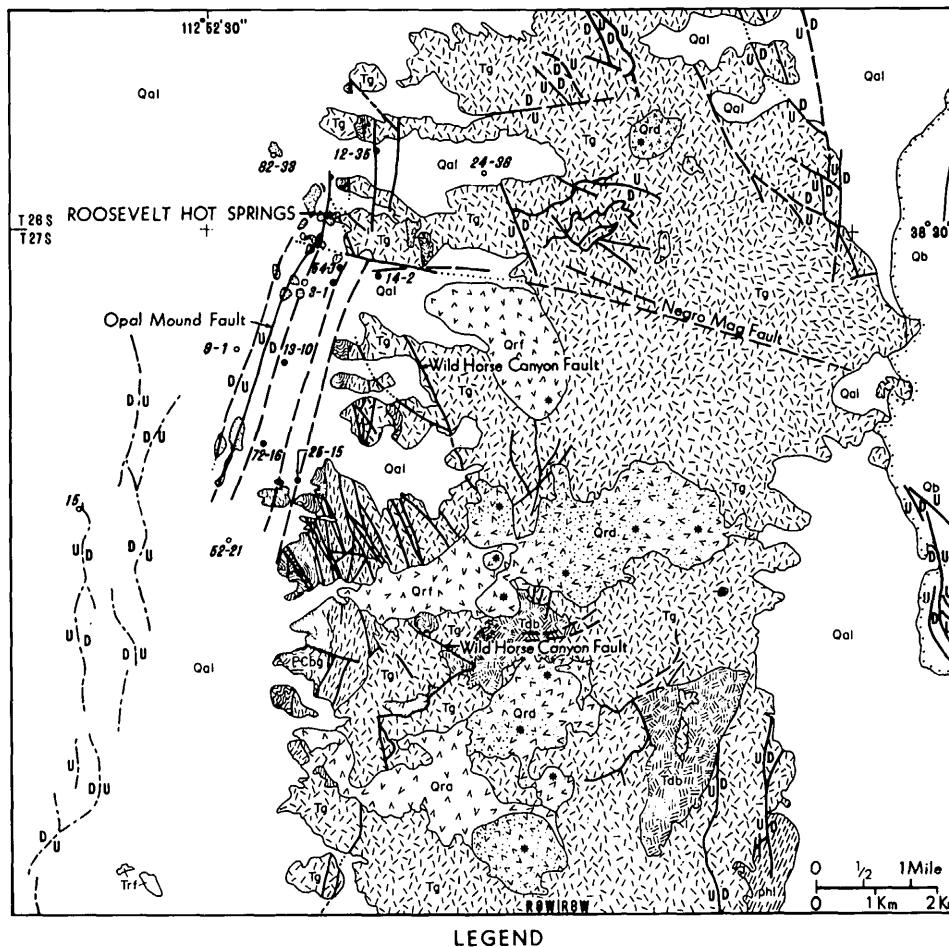


Figure 15. Geologic map of Roosevelt Hot Springs KGRA and vicinity. (after Sibbett and Nielson, 1980).

late Tertiary in age. In the vicinity of the geothermal system, the lithologies range from diorite and granodiorite through granite and syenite in composition.

"Rhyolite flows, pyroclastics, and domes were extruded along the spine of the Mineral Mountains 800,000 to 500,000 years ago (Lipman et al., 1977). The flows and domes are glassy, phenocryst-poor rhyolites. The pyroclastic rocks are represented by air-fall tuff and nonwelded ash-flow tuffs."

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

The hot spring deposits are depicted in Figure 16. A soil survey within the area of

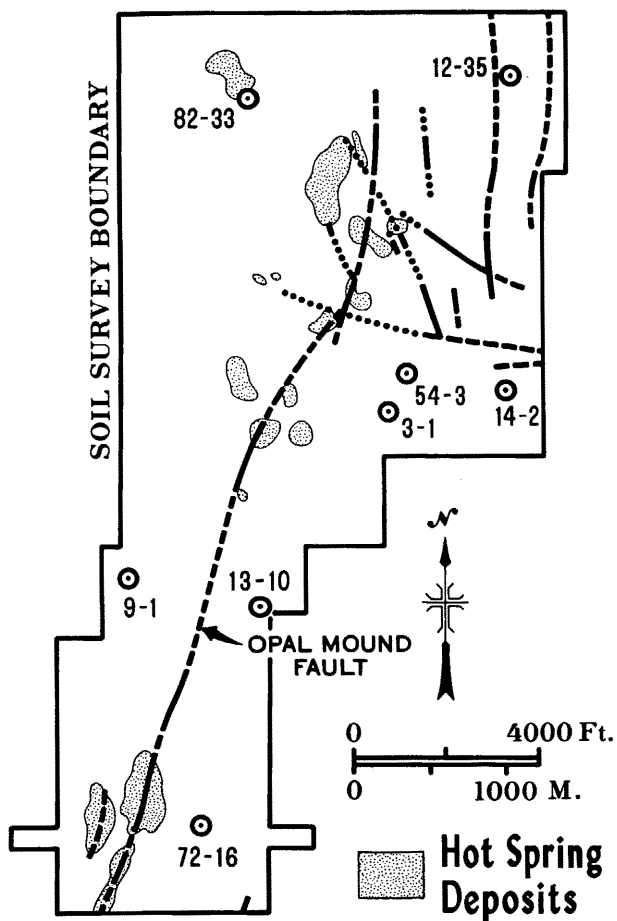


Figure 16. Hot spring deposits, Roosevelt Hot Springs. (after Bamford et al., 1980).

Figure 16 revealed anomalous mercury and arsenic. Hydrothermal alteration in the geothermal system and the adjacent Mineral Mountains is localized along faults and fractures.

"The hydrothermal alteration produced assemblages of quartz + chlorite + epidote + hematite. Hematite is commonly found as specularite veinlets and, where genetic relationships can be observed, hematite mineralization follows sulfide mineralization."

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

Figure 17 shows the heat flow contours, expressed in mWm^{-2} , over the main prospect area at Roosevelt Hot Springs KGRA. The southern lobe, up to Negro Mag fault, is a legitimate expression of the geothermal reservoir as it is currently known. The northern lobe, northwest of the intersection of the Negro Mag fault and the Opal Mound fault, may in part be an expression of leakage of hot water northwestward out of the reservoir. Wells 82-33 and 24-36 are dry, while well 12-35 is productive.

The resistivity contours of Figure 18 correspond with the heat flow high very nicely. The resistivity low is primarily due to surface conductivity of clay minerals produced in hydrothermal alteration (Ward, S.H. and Sill, 1976). Leakage of brine northwestward out of the reservoir would appear to be saturating valley alluvium leading to low resistivities beyond the heat flow high. The CSAMT apparent resistivities at 32 Hz, shown in Figure 19, also directly correlate with the zone of high heat flow. Quantitative two-dimensional modeling of the resistivity and CSAMT data has yielded high-resolution targets for drilling; such resolution is not afforded by modeling the heat flow data.

The self-potential map of Figure 20 reveals a number of closures, some of which are positive and some negative. Quantitative interpretation of the self-potential data has not yet been made and its contribution to understanding the reservoir is not known.

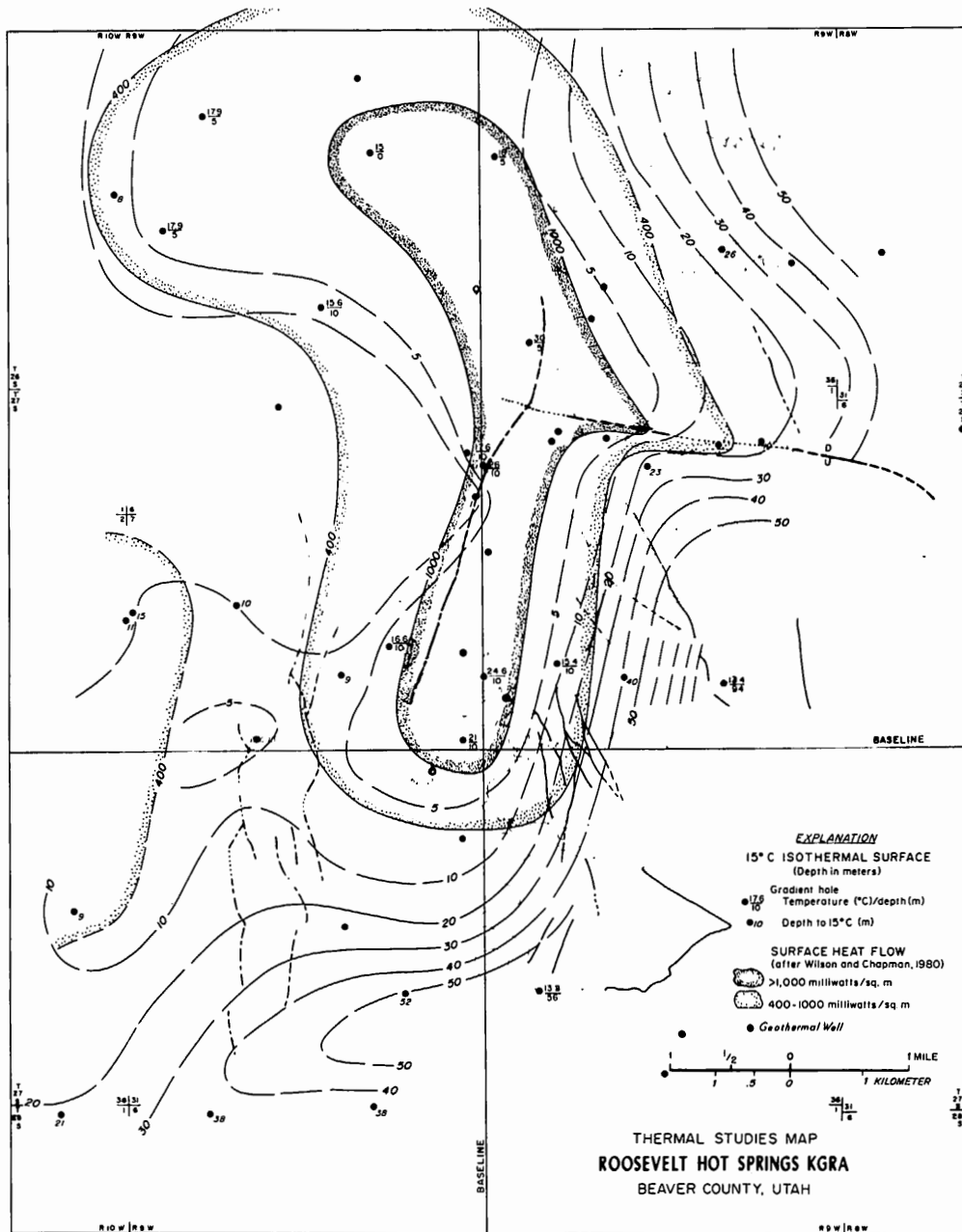


Figure 17. Thermal studies map, Roosevelt Hot Springs KGRA. (from Ross et al., 1982).

Very few microearthquakes have been found beneath wells currently known to be productive. Activity in the region is episodic. During one episode of increased activity, the numerous hypocenters of Figure 21 were located; most occur at a depth of about 5 km. They seem to depict an east-west fault along B-B'. Whether the hypocenters lie on a southward dipping Negro Mag fault, or whether they are associated with another east-west fault is unknown. Research on these microearthquakes and their structural significance is continuing. The Wadati diagram of Figure 22

was developed from the origin times and S- and P-arrival times. A Poisson's Ratio of 0.25 results. Hypocenters occurring beneath the reservoir as currently known, while few in number, yielded a Poisson's Ratio of 0.24. Thus the Poisson's Ratio does not reflect the fact that the reservoir is liquid dominated. It is too early to say whether microearthquakes will contribute to knowledge of the reservoir; to this date they have not.

Douze and Laster (1979) detected no earth

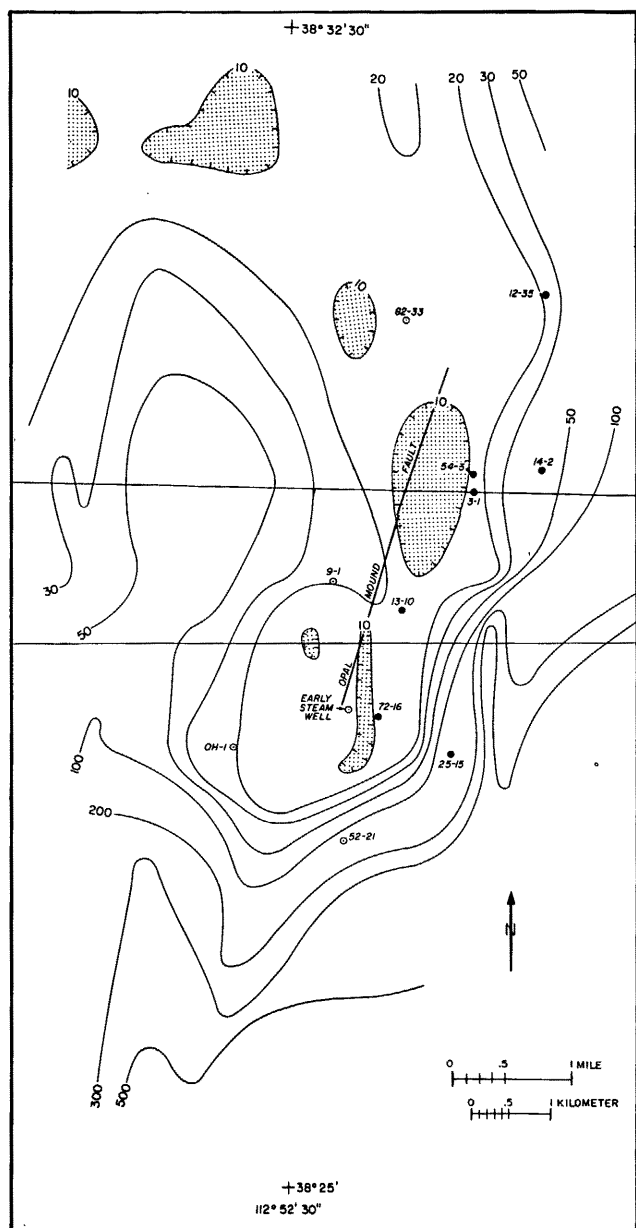


Figure 18. First separation resistivity map from 300 m dipole-dipole survey, Roosevelt Hot Springs KGRA. (after Ward and Sill, 1976).

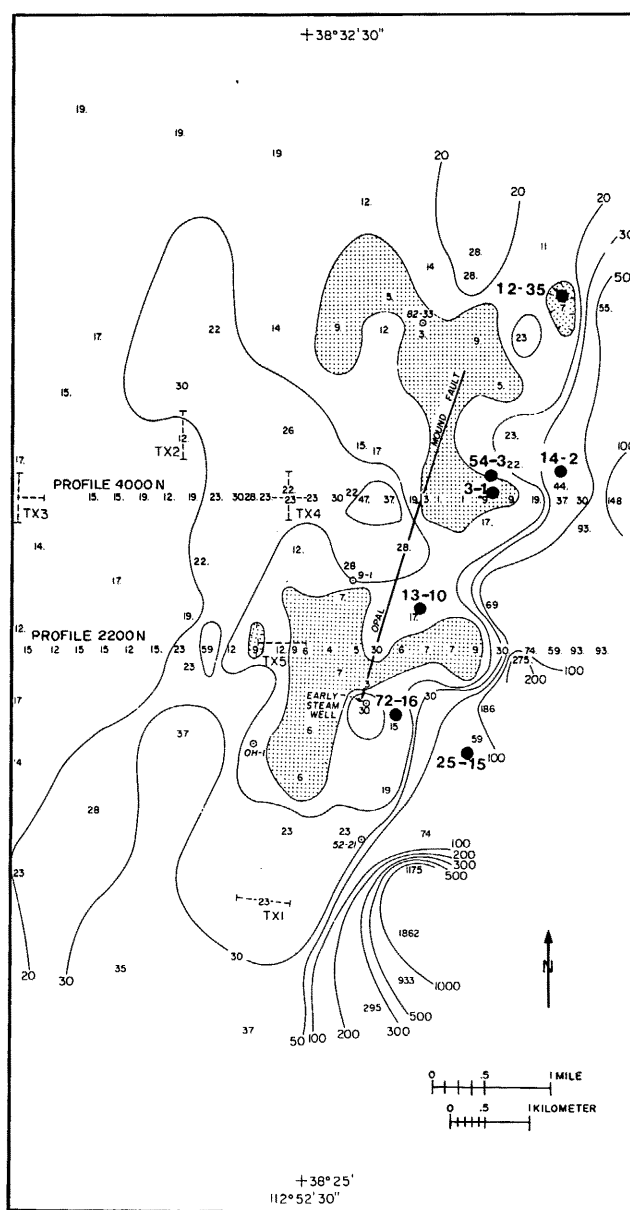


Figure 19. The CSAMT 32 Hz apparent resistivity map of Roosevelt Hot Springs KGRA. (from Sandberg and Hohmann, 1982).

noise that could be related to the geothermal reservoir. Robinson and Iyer (1981) make a modest claim that there is teleseismic evidence for a partial melt at about 15 km depth. Wechsler and Smith (1979) cast doubt on this interpretation and I am inclined to agree with them. Reflection and refraction surveys have yielded information on range front faulting and on depth of valley alluvium, but they have not given any useful information on the reservoir (Ross et al., 1982).

Gravity and magnetic methods have aided in

understanding the structure in the vicinity of Roosevelt Hot Springs, but their contribution to understanding the reservoir is nil.

The magnetotelluric method detected the shallow resistivity structure but at great expense. It also detected a partial melt in the upper mantle. Research is continuing to delineate the lateral extent of the partial melt and to determine whether or not it has any significance to the occurrence of the geothermal reservoir. No deep geothermal system and no magma body were

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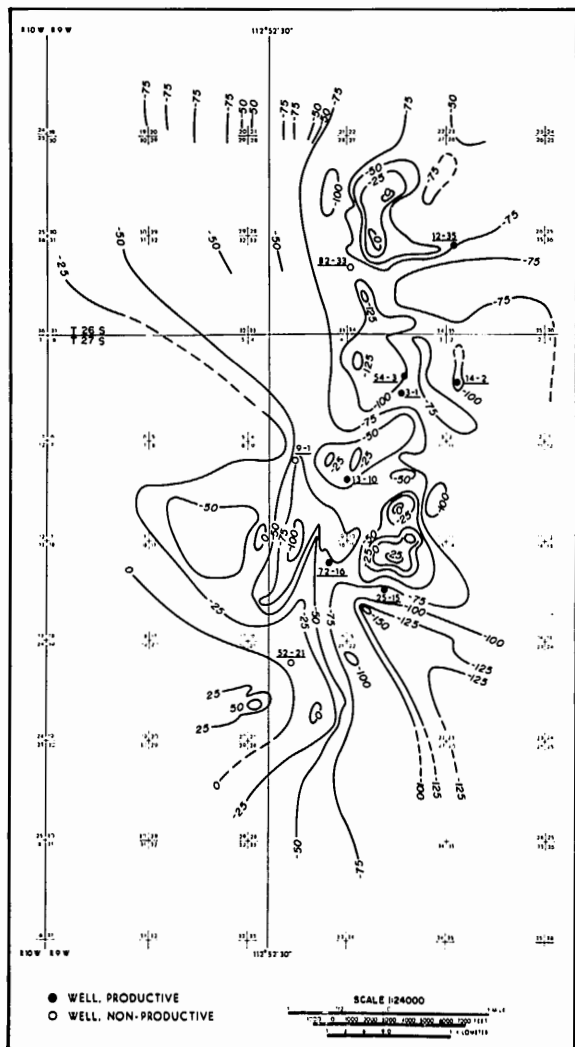


Figure 20. The self-potential map of Roosevelt Hot Springs KGRA (courtesy W. R. Sill).

detected by the MT survey. Detectability of magma chambers by MT is a questionable enterprise as Newman et al. (1983) have demonstrated.

Tuscarora

Pilkington et al. (1980) and Sibbett (1982) have described the geology of the Tuscarora geothermal prospect in Elko County, Nevada. Figure 23 portrays the simplified representation of the geology of the prospect according to Sibbett, who states, "The Tuscarora geothermal prospect is located at the north end of Independence Valley in northern Nevada. Thermal springs issue from Oligocene tuffaceous sediments near the center of an area of high thermal gradient. The springs are associated with a large siliceous sinter mound and are currently depositing silica and calcium carbonate. Measured

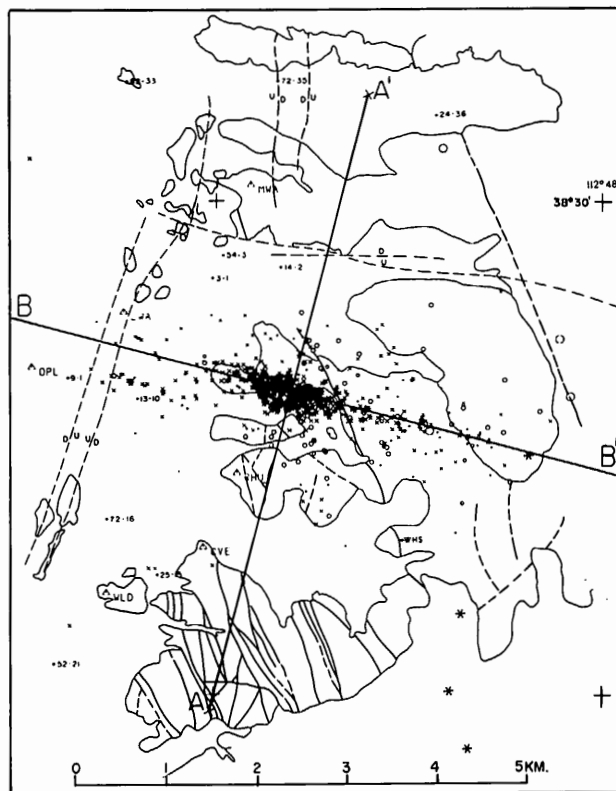


Figure 21. Microearthquakes occurring during swarm, July 1981. (after Zandt et al., 1982).

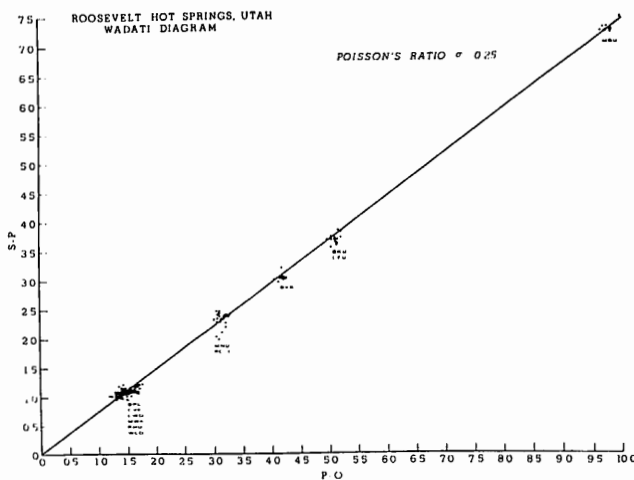


Figure 22. Wadati diagram derived from earthquakes occurring at Roosevelt Hot Springs KGRA during July, 1981.

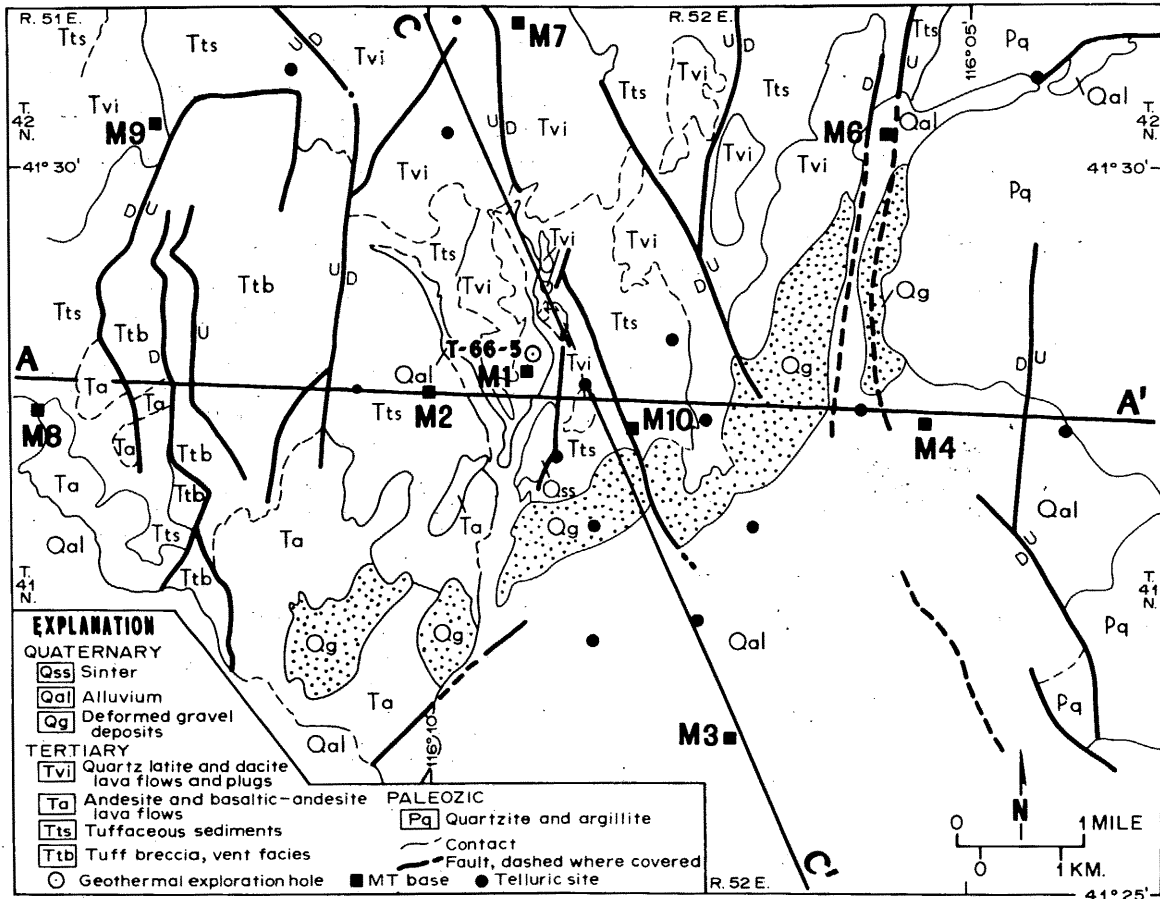


Figure 23. Generalized geology, Tuscarora. (from Sibbett, 1982).

fluid temperatures range up to 95°C, and chemical geothermometers indicate a reservoir temperature of 216°C. The Independence Valley contains 35- to 39-m.y.-old tuffs and tuffaceous sediments which overlie Paleozoic clastic and volcanic rocks and are overlain by Miocene lava and pyroclastic flows. The rocks have been deformed by normal faults trending north-south and northwest and by folds trending north-south which have been active in the Pleistocene."

The heat flow anomaly at the Tuscarora prospect is centered on Hot Sulphur Springs as Figure 24 illustrates (from Pilkington et al., 1980). The north-south faulting is evident in the gravity map of Figure 25 from Pilkington et al., 1980). Meidav and Tonani (1975) observed "that both microearthquake activity and thermal spring occurrences are more commonly associated with that side of the basin which has the steeper gravity gradient". If this is an observation upon which we can rely, let us test it with the available data. From Figure 25 it is evident that the east side of Independence Valley has the steepest gradient of gravity. The range lines in Figures 23 and 25 permit ready correlation of the gravity

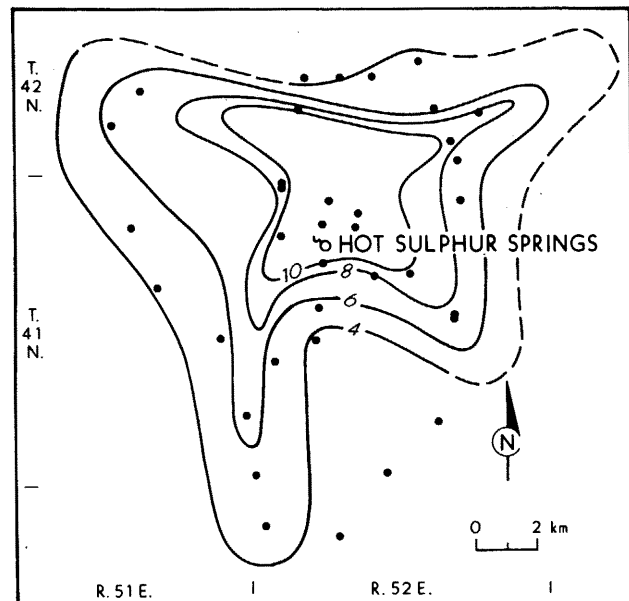


Figure 24. Contours of heat flow at Tuscarora (in HFU). (from Pilkington et al., 1980).

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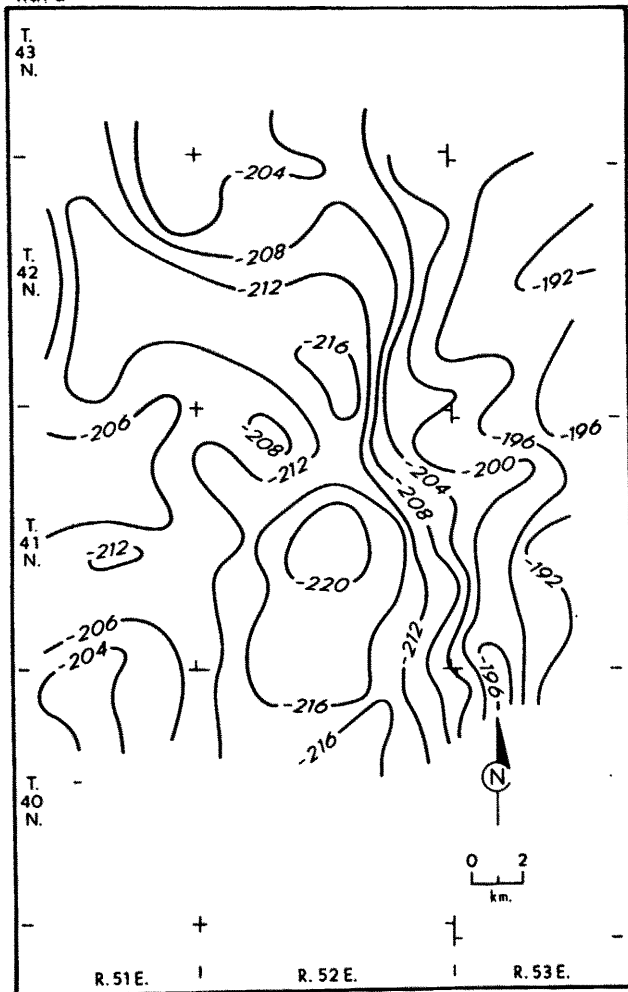


Figure 25. Contours of Bouguer gravity at Tuscarora. (from Pilkington et al., 1980).

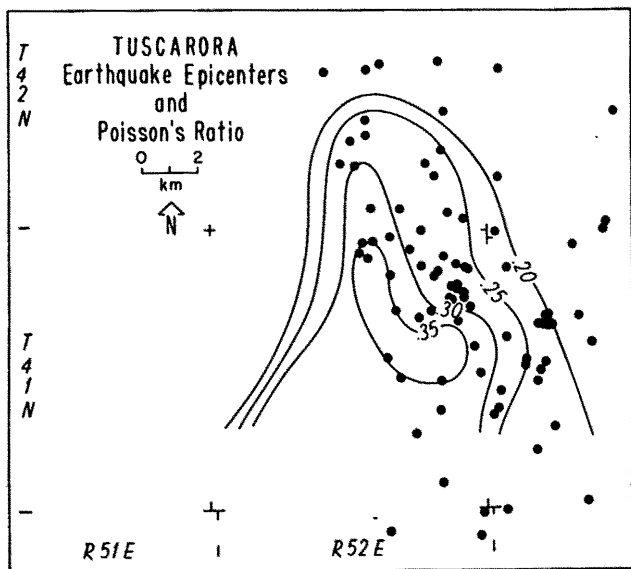


Figure 26. Distribution of microearthquakes and contours of Poisson's ratio. (from Pilkington et al., 1980).

data with the geology. The microearthquakes of Figure 26, as documented by Pilkington et al. (1980), are scattered but clearly follow a range front fault suggested by the steep gravity contours of Figure 25. Contrariwise, Hot Sulphur Springs occur well west of this presumed fault line.

As concluded by Sibbett (1982),

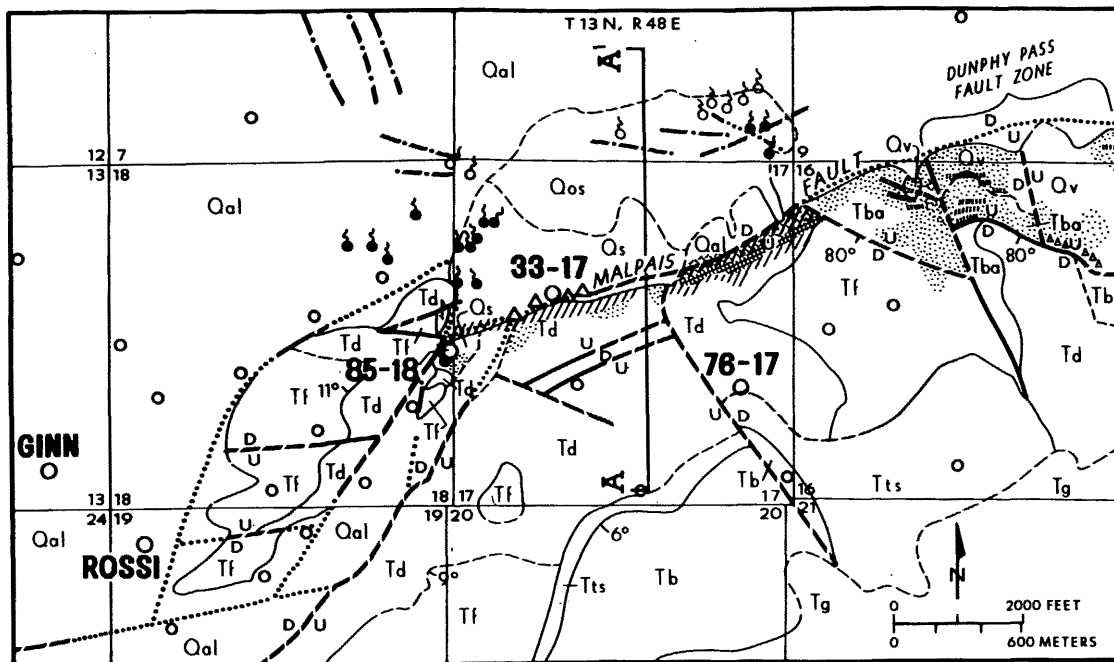
"The surface expression of the Hot Sulphur Springs thermal system is controlled by a fault zone trending N20°E. Exposed argillic alteration produced by the thermal system is limited to the spring area. Quartz-sericite alteration which predates the present thermal system is present along the fault zone."

"The subsurface character of the geothermal system is not known, but the geophysical and geological data are consistent with an interpretation that the reservoir is 3 to 5 km southeast of the hot springs. In this model, meteoric water circulates down along the range-front fault system and is heated at depth. The thermal waters rise along major fractures, perhaps the intersection of the N10°E and N30°W fault zones, into either a solution reservoir in the lower-plate carbonates or a fracture reservoir in the overlying Valmy Group quartzite. The fracture reservoir and feeder channelways may have been formed by brecciation along the thrust fault and by formation of the deep graben. The reservoir cap consists of the incompetent and less permeable Tertiary tuffs and tuffaceous sedimentary rocks, the base being 1,200 m or more below the surface. Some of the thermal fluids migrate up major fractures within the Paleozoic shale, chert and greenstone unit which overlies the Valmy Group quartzite. The fluids probably move updip to the northwest along gravel aquifers either at the base of or within the tuffaceous sedimentary rocks, ultimately reaching the surface along the faults at the hot springs. Cold water aquifers in the thick quartzite gravel overlying the tuffaceous sedimentary rocks apparently mask the thermal anomaly directly above the reservoir."

Drilling on this prospect has been confined to the general vicinity of the heat flow high surrounding Hot Sulphur Springs. If Sibbett's model is correct, and resistivity data (Mackelprang, 1982) would tend to support it, then drilling might recommence along the zone of the eastern range front depicted by the gravity and microearthquake data. In this region, the contoured values of Poisson's Ratio shown in Figure 26 are probably not indicative of the characteristics of the reservoir but rather of the fluid saturated alluvium of Independence Valley (Berkman and Lange, 1980).

Beowawe

Figure 27 depicts the generalized geologic map at Beowawe (from Sibbett, 1983). The Beowawe Geysers have formed a 850 m long sinter terrace at the base of the Malpais scarp.



EXPLANATION		MAP SYMBOLS	
Qs	SILICEOUS SINTER	—	CONTACT
Qal	ALLUVIUM, FINE GRAVEL AND SAND	—	FAULT, WITH BRECCIA
Tg	GRAVEL DEPOSITS AND TAN MUDSTONE	—	PHOTO LINEAR
Tb	OLIVINE-PYROXENE BASALT FLOWS	—	QUARTZ AND CALCITE VEINS
Tts	TUFFACEOUS SEDIMENTARY ROCKS	—	MODERATE HEMATITE
Tf	FELSITE LAVA FLOWS	—	STRONG HEMATITE STAIN
Td	PORPHYRITIC DACITE LAVA FLOWS	—	ARGILLIC
Tbf	BASALT LAVA FLOWS	—	SILICIFICATION AND MINOR SINTER
Tt	TUFFACEOUS SEDIMENTARY ROCKS	—	
Tba	BASALTIC-ANDESITE LAVA FLOWS	—	
Tc	CONGLOMERATE, CLASTS OF CHERT, SILTSTONE AND QUARTZITE	—	
Ov	VALMY FORMATION, SILTSTONE, QUARTZITE, CHERT AND ARGILLITE	—	
		—	THERMAL SPRING
		—	STEAM VENT
		—	GINN GEOTHERMAL WELL
		—	THERMAL GRADIENT HOLE

Figure 27. Generalized geology, Beowawe. (from Sibbett, 1983).

"The Beowawe geothermal system in northern Nevada is a structurally controlled, water-dominated resource with a measured temperature of 212°C (414°F). Surface expression of the system consists of a large, active opaline sinter terrace that is present along a Tertiary to Quaternary normal fault escarpment. The thermal system appears to be controlled by the subsurface intersection of the east-northeast trending, north dipping Malpais fault with a pre-existing northwest trending fault which dips south and has 884 m of vertical displacement.

Surface alteration associated with the geothermal system is vertically zoned along the Malpais escarpment with, from base to top: hematite stained, argillized rock along the fault trace; silicification and quartz veining; and argillic, acid leach zone at the top. Subsurface alteration generally increases with depth in the volcanic rocks and is most intense in basaltic-andesite lava flows which are capped by tuffaceous sedimentary rock." (Sibbett, 1983)

Figure 28 presents the contoured heat flow values, which provide a focus for attention to the

flexure in the Malpais Fault near well 85-18. Smith (1983) discusses the thermal hydrology and heat flow at Beowawe. Swift (1979) outlined the area of low resistivity, as in Figure 29, and this outline passes through the center of the heat flow high. A dipolar self-potential anomaly, shown in Figure 30, best delineates the convective system at Beowawe (Swift, 1979).

According to Swift (1979), gravity and magnetic data delineate the Malpais Fault and the important north-northwest structures which may control the reservoir at depth. Seismic reflection data are too noisy, because of the interbedded volcanic section, to provide definitive structural information. In contrast, Swift (1979) reports that the Geysers area at Beowawe appears as an earth noise source. Magnetotelluric data seem to have yielded only regional information at Beowawe, based on Swift's remarks.

6.0 COMPARATIVE CASE HISTORIES

An evaluation has been made, in Table 3, of the contribution made by each of 14 geophysical

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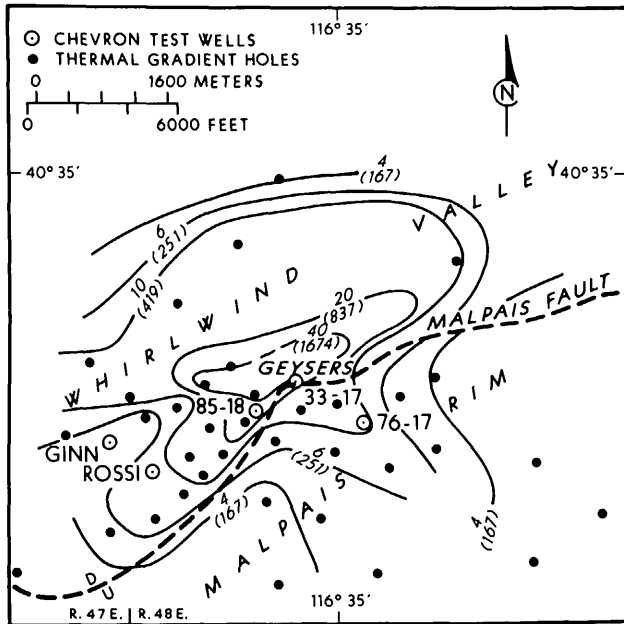


Figure 28. Contours of heat flow at Beowawe (in HFU). (courtesy B. S. Sibbett).

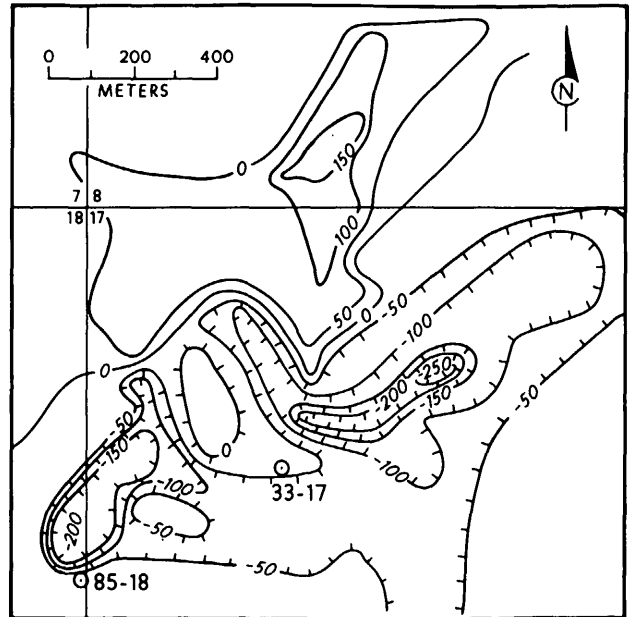


Figure 30. Contours of self-potential at Beowawe. (from DeMouley and Corwin, 1980).

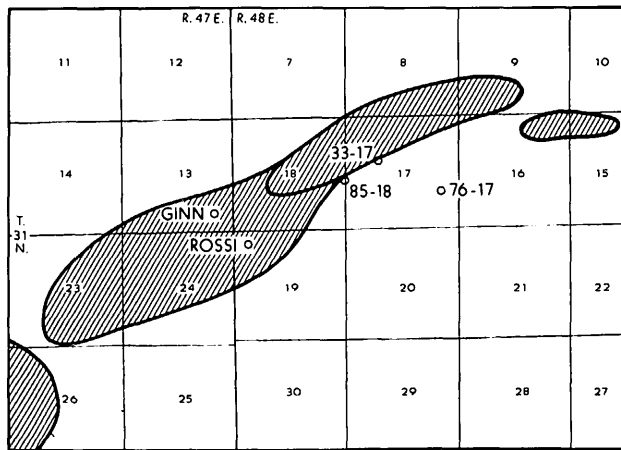


Figure 29. Outline of zone of low resistivity at Beowawe. (after Swift, 1979).

methods to understanding the known or postulated reservoir at each of 13 geothermal prospects. Each method has been rated from 1 (good) through 4 (poor). The rating is subjective, but a serious attempt has been made to apply the rating uniformly throughout the matrix of Table 3. The surprising results of this analysis are that: a) geophysical methods are uniformly inconsistent in performance, b) no geophysical method ranks one, five methods rank about two, eight rank between two and one half and three, while none ranks four, c) no combination of any four methods has been

successful at more than one site where "successful" means a ranking of one or two, d) the most useful of the methods, judging by their mean rankings, are heat flow, microearthquakes, gravity, resistivity, and self-potential, e) the least effective methods are earth noise, magnetics, and magnetotellurics. The CSAMT method has been included with CSEM for purposes of the comparative study. There are 23 entries of good (1), 29 of fair (2), 48 of questionable (3), and 10 entries of poor (4) in Table 3. Thus 53% of the time the results of geophysical surveys in geothermal exploration are either questionable or poor.

Overall, this study provides a somewhat discouraging picture. I would tend to use heat flow, microearthquakes, gravity, resistivity (or CSAMT), and self-potential methods at all prospects. Once these data were interpreted and correlated, I would then decide whether or not additional geophysical surveying was justified. All geophysical surveys should be designed with one or more conceptual geological models in mind and the density and extent of the geophysical coverage should be compatible with the range of conceptual geologic models.

Clearly the reader will want to know why each method has performed poorly or questionably under some circumstances. To attempt to satisfy the reader's curiosity, I have prepared Table 4 in which I have indicated my assessment of why a method rated a 3 or a 4 in Table 3. Six categories of answers seem to be sufficient. These are:

TABLE 3 CONTRIBUTION TO UNDERSTANDING RESERVOIR

METHOD	IGNEOUS RELATED													MEAN
	COSO*	LONG VALLEY*	ROOSEVELT HOT SPRINGS*	BALTAZOR	BEOVAWE*	COVE FORT*	DESERT PEAK*	GRASS VALLEY	MCCOY	SAN EMIDIO	SODA LAKE +	TUSCARORA	RAFT RIVER	
EARTH NOISE	3	3	4		1			4	3	3				3.0
MICROEARTHQUAKES	1	3	3	3		2		1	1			1		1.9
TELESEISMS	2	1	3						3			4		2.6
REFRACTION	2	2	3					3					2	2.4
REFLECTION			3		3	4		1		2	2			2.5
GRAVITY	4	2	4	1	2	3	3	1	3	1	2	1	2	2.2
MAGNETICS	1	3	4	3	3	2		4	3		4	3	3	3.0
RESISTIVITY	2	3	1	1	2	1	4	2	3	2	3	2	3	2.2
CSEM & CSAMT		3	1					3	3		2			2.4
SCALAR AMT	2	3		2									3	2.5
MT/AMT			3		3		3	3	3		3	3	3	3.0
SELF POTENTIAL		1	2	3	1			2	2			3	3	2.1
TELLURICS	3			3				2	2					2.5
HEAT FLOW	1	3	1	2	1	2	1	2	3	3	1	3	2	1.9

* EXPECTED TO BE COMMERCIAL IN NEAR TERM
 + EXPECTED TO BE COMMERCIAL IN LONG TERM

CODE: 1 GOOD
 2 FAIR
 3 QUESTIONABLE
 4 POOR

TABLE 4 REASONS FOR POOR PERFORMANCE OF GEOPHYSICAL METHODS

METHOD	IGNEOUS RELATED													SUMMARY
	COSO*	LONG VALLEY*	ROOSEVELT HOT SPRINGS*	BALTAZOR	BEOVAWE*	COVE FORT*	DESERT PEAK*	GRASS VALLEY	MCCOY	SAN EMIDIO	SODA LAKE	TUSCARORA	RAFT RIVER	
EARTH NOISE	T ₁	T ₁	N					N	R	T ₁				3T ₁ ,2N,1R
MICROEARTHQUAKES		R	R	R										3R
TELESEISMS	I		I						R			R		2R,2I
REFRACTION			N					N						2N
REFLECTION			N		T ₂	T ₁								1T ₁ ,1T ₂ ,1N
GRAVITY	N		N			N	N		R					4N,1R
MAGNETICS		R	N	R	R			N	R		R	D	R	6R,2N,1D
RESISTIVITY		R					T ₁		D		R		R	3R,1T ₁ ,1D
CSEM & CSAMT		R						I	I					2I,1R
SCALAR AMT		R												2R
MT/AMT			N		N		I	I	I		N	I	R	4I,3N,1R
SELF POTENTIAL				T ₁								R	R	2R,1T ₁
TELLURICS	R			D										1R,1D
HEAT FLOW		R							R	R		R		4R

T₁ - TECHNOLOGY DEFICIENT, HAS BEEN IMPROVED
 T₂ - TECHNOLOGY DEFICIENT, HAS NOT BEEN IMPROVED
 I - INTERPRETATION PROCEDURE QUESTIONABLE

D - DATA SET INCOMPLETE
 N - NO RECOGNIZABLE SIGNATURE OVER RESERVOIR
 R - RELATIONSHIP TO RESERVOIR UNCERTAIN

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- T₁ - technology deficient, but has since been improved,
 - T₂ - technology deficient, and has not yet been improved,
 - I - interpretation procedure questionable,
 - D - data set incomplete,
 - N - no recognizable signature over reservoir,
- and
- R - relationship to reservoir uncertain.

Table 4 contains 6 T₁, 1 T₂, 12 N, 28 R, 8 I, and 3 D entries. The T₁, I and D entries total 17 (29%), indicating that improvement in equipment or interpretation has or can be made. Unfortunately, we can do nothing about the poor ratings associated with the 41 entries (71%) reporting T₂, N, and R.

The following comments follow on each method:

earth noise

The technology has improved so that the method should perform better when advantage is taken of f-k processing.

microearthquakes

Some geothermal reservoirs do not yield sufficient numbers of microearthquakes to permit the method to be successful.

telescisms

The method does not always produce useful results.

refraction seismology

The method is not always applicable to reservoir delineation. It should be used late in the exploration sequence.

reflection seismology

Data processing advances in reflection seismology may ultimately permit us to work in volcanic environments and efforts should be directed toward that end. However, the method will not always be applicable to reservoir delineation and should be used late in the exploration sequence.

gravity and magnetics

The main use of the gravity and magnetic methods will continue to be as aids in geological mapping of structure. Gravity seems to be quite useful in mapping range front faults with which some geothermal reservoirs are associated. The magnetic method occasionally will prove useful in mapping zones of magnetite destruction.

resistivity

The technology of data acquisition, processing, and interpretation has been much improved for Schlumberger and dipole-dipole

resistivity surveys. Algorithms for 2D modeling of resistivity data are now used routinely. Algorithms for 3D modeling are available. The bipole-dipole method has not proven itself to be satisfactory for reservoir delineation in most cases.

CSEM/CSAMT

These methods have not been sufficiently tested in geothermal environments to be certain of their future. Interpretation procedures for CSAMT are capable of encompassing 2-D and 3-D earth models, although there has been a reluctance on the part of industry to use them. Reliable computer algorithms to permit 2D modeling of CSEM data are only now emerging. Algorithms for 3D modeling are still on the drawing boards.

scalar AMT

Scalar AMT doesn't always produce results that are as good as we prefer and its use should be restricted to reconnaissance surveys.

MT/AMT

There is a tendency to use 1D inversion, only, in interpretation of MT/AMT data. Algorithms for 2D and 3D interpretation are available but industry seems reluctant to use them. This is the weakest link in application of the method. Data acquisition and processing have improved greatly in recent years, to the extent that there is no excuse for data of poor quality.

There is also a tendency by industry to use MT/AMT when it is not warranted. It should be reserved for special situations, usually late in the exploration sequence.

self-potential

The self-potential method shows great promise, if the new quantitative interpretation schemes work as expected. The method doesn't always produce meaningful signatures over geothermal systems.

tellurics

Since it lacks a truly quantitative scheme for interpretation, the method should be used only in reconnaissance.

heat flow

As Blackwell (1983) has indicated, heat flow as measured at surface is not always a reliable indicator of a high quality geothermal resource. Reference should be made to Blackwell's paper for cognizance of the advantages and limitations of heat flow in geothermal applications.

7.0 CONCLUSIONS

The performance of 14 different geophysical methods used at 13 high temperature geothermal

sites in the Northern Basin and Range Province has been somewhat disappointing. Heat flow, microearthquakes, gravity, resistivity and self-potential methods appear to be the most consistently useful, although none of them performs up to expectations all of the time. Recent improvements in interpretation procedures are expected to benefit the resistivity and self-potential methods. The least effective methods at the present time are earth noise, reflection seismology, magnetics, magnetotellurics, and tellurics. Recent improvements in survey design will make the earth noise method much more effective for those reservoirs which emit earth noise. Accomplished improvements in interpretation of magnetotelluric data, if adopted by industry, will improve that method's performance. However, the magnetotelluric method ought to be reserved for those special prospects to which it is applicable. Improvements in processing reflection seismic data in volcanic covered areas may make that method more generally applicable in the late stages of an exploration sequence. The teleseismic method, developed to a reasonable degree of sophistication, may occasionally contribute, in a broad sense, to understanding the reservoir. The refraction seismic method probably ought to be considered a subset of combined reflection-refraction surveys for optimum resolution of subsurface velocity structures. However, refraction seismic surveys should be considered also as an adjunct to microearthquake surveys since they can provide 3D velocity distributions necessary for interpretation of microearthquake data. Controlled source electromagnetic and controlled source audiomagnetotelluric methods require much more exposure in geothermal exploration before they can be evaluated properly. However, the controlled source audiofrequency magnetotelluric method looks very promising, and it may replace resistivity surveying if its cost-effectiveness can be established as is expected. Scalar AMT and tellurics seem destined for use as inexpensive reconnaissance techniques which will find occasional application.

There is evidence of need for improvement in survey design, data acquisition, data processing, and data interpretation in geothermal exploration in the Northern Basin and Range Province. Yet the predominant problems with application of geophysical methods in this environment are related to fundamental limitations of the methods as have been described in the text.

Of the five case histories presented for comparison in this article, we find the following results.

Long Valley

While no high temperature reservoir has been defined, there is a chance that ultimately the teleseismic, microearthquake and self-potential methods may prove to be valuable in delineating a postulated fracture-controlled reservoir beneath the medial graben in the resurgent dome. The

existence of such a reservoir probably requires regional faulting through the resurgent dome following the graben, and subsequent extension produced by recent bulging of the dome. Refraction seismology and gravity have been exceptionally useful in quantifying the potential reservoir, advocated by the USGS, to be the Bishop Tuff within the Long Valley caldera. The heat flow anomaly at this site may be misleading.

Coso

The combination of resistivity, scalar AMT, shallow heat flow, and airborne magnetic data, when added to the knowledge of hydrothermal alteration provided by careful geological mapping, have clearly defined the drilling targets. Teleseismic P-wave delays and refraction seismology are methods which have provided information of secondary importance.

Roosevelt Hot Springs

The heat flow pattern at Roosevelt Hot Springs has been fundamental in outlining the reservoir. However, careful application of the resistivity and CSAMT techniques has provided the detail necessary to site wells. Self-potential data at this site have not yet been interpreted quantitatively but ultimately it may prove to be useful in understanding and delineating the reservoir.

Tuscarora

Microearthquakes and gravity draw attention to the eastern margin of Independence Valley, while heat flow and surface manifestations draw attention to a region around Hot Sulphur Springs about 4 km to the west. Drilling has been concentrated in the vicinity of Hot Sulphur Springs whereas our speculation dictates that the reservoir might be along the range front fault bounding the eastern margin of Independence Valley. Dipole-dipole resistivity interpretations would tend to confirm this notion. The heat flow data at this site may be misleading.

Beowawe

The self-potential method has been the best confirmation of the heat flow anomaly which seems to be centered over the shallow reservoir. Dipole-dipole resistivity has provided supporting evidence for the location of the drilling target. Gravity and magnetics have helped to define the Malpais Fault zone while application of MT/AMT was necessary to attempt to delineate the deep reservoir.

From the above comments it should be evident that there is no common denominator of geophysical methods which might lead us to a better record of discovery and delineation of geothermal reservoirs. Exploration for igneous-related geothermal systems is no different from those geothermal systems of no obvious igneous relationship. Exploration for geothermal

resources of any type, those with surface manifestations or those without, is difficult.

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