GEOPHYSICS, VOL. 50, NO. 12 (DECEMBER 1985); P. 2666-2699, 10 FIGS., 4 TABLES.

# State-of-the-art geophysical exploration for geothermal resources

Phillip. M. Wright*, Stanley H. Ward*, Howard P. Ross*, and Richard C. West*

## ABSTRACT

At the present stage of development, use of geothermal energy saves about 77 million barrels of oil per year worldwide that would otherwise be required for electrical power generation and direct heat applications. More than a dozen countries are involved in development of geothermal resources. Currently, only the moderate- and high-temperature hydrothermal convective type of geothermal system can be economically used for generating electric power. Lower-temperature resources of several types are being tapped for space heating and industrial processing. Geophysics plays important roles both in exploration for geothermal systems and in delineating, evaluating, and monitoring production from them. The thermal methods, which detect anomalous temperatures directly, and the electrical methods are probably the most useful and widely used in terms of siting drilling targets, but gravity, magnetics, seismic methods, and geophysical well logging all have important application.

Advances in geophysical methods are needed to improve cost effectiveness and to enhance solutions of geologic problems. There is no wholly satisfactory electrical system from the standpoint of resolution of subsurface resistivity configuration at the required scale, depth of penetration, portability of equipment, and survey cost. The resolution of microseismic and microearthquake techniques needs improvement, and the reflection seismic technique needs substantial improvement to be cost effective in many hard-rock environments. Well-logging tools need to be developed and calibrated for use in corrosive wells at temperatures exceeding 200°C. Well-log interpretation techniques need to be developed for the hard-rock environment. Borehole geophysical techniques and geotomography are just beginning to be applied and show promise with future development.

## INTRODUCTION

Development of geothermal resources is being aggressively pursued on a worldwide basis. Approximately 3 800 MW of electricity are currently being generated from geothermal energy, and about 10 000 thermal MW are being used for direct heat applications. While this level of energy may seem small compared to the estimated $8.4 \times 10^6$ MW of human use of fossil energy (Williams and Von Herzen, 1974), it nevertheless represents a savings in the consumption of about 77 million barrels of oil per year worldwide. Estimating the ultimate potential contribution of geothermal energy to mankind's needs is very difficult for at least three reasons: (1) long-range future energy costs, although generally predicted to be higher than today's levels, are uncertain, and a large number of lower-grade geothermal resources could become economic at higher energy prices; (2) only preliminary estimates of the worldwide resource base have been made; and (3) technology for using energy in magma, hot rock, and normal thermal-gradient resources, whose potential contributions are very large, is not yet available. Nevertheless, White (1965) estimated that the total heat stored above surface temperature in the Earth to a depth of 10 km is about $1.3 \times 10^{27}$ J, equivalent to the burning of about $2.3 \times 10^{17}$ barrels of oil. If even a small part of this heat could be made available, the contribution would be significant.

In the United States, commercial development of geothermal energy is pursued by private industry, and many of the data generated are not available for public inspection and use. In substantially all of the rest of the world, geothermal development is sponsored by federal governments, and data are reasonably accessible, although not all data appear in readily available journals. Active programs in geothermal exploration and development are being carried out in China, El Salvador, Ethiopia, France, Iceland, Indonesia, Italy, Japan, Kenya, Mexico, New Zealand, the Philippines, the Soviet Union, the

Manuscript reviewed by C. M. Swift, Jr. and J. N. Moore. Manuscript received by the Editor May 13, 1985; revised manuscript received June 3, 1985.
*Earth Science Laboratory, University of Utah Research Institute, 391 Chipeta Way, Suite C, Salt Lake City, Utah 84108

United States, and to a lesser extent in other countries. Expertise arising from first-hand experience in Iceland, Italy, New Zealand, Mexico, Japan, and the U.S. (primarily) is being used by the less-developed countries to assist their geothermal efforts. The United Nations (U.N.) sponsors both scientific work and education in underdeveloped countries. Exploration projects using U.N. funding have been carried out in El Salvador, Chile, Nicaragua, Turkey, Ethiopia, and Kenya, and the U.N. sponsors geothermal training programs at the United Nations University locations in Iceland, Italy, and New Zealand. La Organización Latinoamericana de Energía (OLADE), headquartered in Quito, Ecuador, provides support for geothermal development in Central and South America. In short, an infrastructure for geothermal development is being built throughout the world, and although it is small compared to the corresponding petroleum or minerals infrastructures, it is making important contributions.

We review here the application of geophysical methods to geothermal exploration and development and assess the current state-of-the-art. Previous reviews of geophysical applications were given by Pálmason (1976), McNitt (1976), Meidav and Tonani (1976), Ward (1983b), and Rapolla and Keller (1984), among others. There are more published accounts of geophysical work in the geothermal environment than can be discussed or referenced here. We cite a few typical references for each application discussed. We apologize for the emphasis on U.S. literature, with which we are most familiar, and for omissions in recognizing contributions of many authors.

## NATURE OF GEOTHERMAL RESOURCES

Geothermal resources have three common components: (1) a heat source, (2) a reservoir with porosity and permeability, and (3) a fluid to transfer the heat to the surface. In some exploitation schemes, the permeability must be created artificially. One useful classification of geothermal resource types is shown in Table 1. Hydrothermal resources, as the term implies, are characterized by natural thermal waters, and are divided into resources with and without significant large-scale convection. Hot-rock resources have no natural fluid to transport heat to the surface, and are the subject of current research to develop means of extracting their energy. Only hydrothermal resources have been developed to any extent, because the other resources are presently uneconomic.

### Classification of geothermal systems

*Convective hydrothermal resources* are geothermal resources in which the Earth's heat is actively carried upward by circulation of naturally occurring hot water or steam. Underlying some of the higher-temperature resources is presumably a body of molten or recently solidified rock with temperature possibly in the range 400°C to 1 100°C. Other convective resources result simply from circulation of water along faults and fractures or within permeable aquifers to depths where the rock temperature is elevated, with heating of the water and subsequent buoyant transport to the surface or near-surface.

Thermal waters can be produced from some *basins* or from

**Table 1. Geothermal resource classification. (Modified from White and Williams, 1975).**

| Resource Type | Temperature Characteristics |
|---|---|
| Convective Hydrothermal Resources | |
| Vapor dominated | about 240°C |
| Hot-water dominated | about 30°C to 350°C+ |
| Other Hydrothermal Resources | |
| Sedimentary basins/Regional aquifers (hot fluid in sedimentary rocks) | 30°C to about 150°C |
| Geopressured (hot fluid under pressure that is greater than hydrostatic) | 90°C to about 200°C |
| Radiogenic (heat generated by radioactive decay) | 30°C to about 150°C |
| Hot Rock Resources | |
| Part still molten | higher than 600°C |
| Solidified (hot, dry rock) | 90°C to 650°C |

*regional aquifers*. In an area of 25 000 $km^2$ in north-central U.S., the Madison and other formations contain thermally anomalous waters, whose origin is not fully understood. Substantial benefit is being realized in France for space heating by production of warm water contained in the Paris basin (Varet, 1982). Many other areas of occurrence of this resource type are known worldwide.

*Geopressured* resources consist of deeply buried fluids contained in permeable, sedimentary rocks warmed in a normal or anomalous geothermal gradient by their great depth of burial. These fluids are tightly confined by surrounding impermeable rock and thus bear pressure that is much greater than hydrostatic, that is, the fluid pressure supports a portion of the lithostatic load (Wallace et al., 1979). In the U.S. Gulf Coast area, these geopressured fluids have temperatures up to 150 C and also contain dissolved methane. Therefore, three sources of energy may be available: (1) heat, (2) mechanical energy derived from the high wellhead pressure, and (3) recoverable methane. The effects of producing these resources is being assessed by current research coordinated by the Idaho National Engineering Laboratory for the U.S. Department of Energy.

*Radiogenic* geothermal resources are postulated to occur, for example, in the eastern U.S. (Costain et al., 1980), where the coastal plain is blanketed by a layer of thermally insulating sediments. Granitic intrusions having enhanced heat production from radioactive decay occur in places beneath these sediments. Gravity and aeromagnetic surveys to locate covered intrusions, followed by heat-flow studies to distinguish heat sources, have been carried out largely under U.S. Department of Energy sponsorship. Only one attempt has been made to drill an area believed to contain a radiogenic geothermal resource, and this test was moderately encouraging in that temperatures of 80 C were encountered at a depth of about 1 520 m.

*Hot rock* resources comprise those which have little or no natural hydrothermal convection, and the resource may be molten, partly molten, or solidified. The feasibility and economics of extraction of heat from hot, dry rock is presently the subject of a cooperative research effort among the U.S., the Federal Republic of Germany, and Japan. The research is centered at the U.S. Department of Energy's Los Alamos National Laboratory in New Mexico (Smith and Ponder, 1982) and similar experiments have been carried out in England (Batchelor, 1982). This work indicates the technological feasibility of inducing permeability in hot, tight crystalline rocks at depths of about 3 km through hydraulic fracturing from a deep well. Sophisticated seismic techniques have been developed to map the fractures during their formation. A second borehole is used to intersect the fracture system. Water can then be circulated to transport the energy to the surface. Fluids at temperatures of 150 C to 200 C have been produced in this way from boreholes at Fenton Hill near the Valles Caldera, New Mexico.

Experiments are underway at the U.S. Department of Energy's Sandia National Laboratories in Albuquerque, New Mexico to learn how to extract heat energy directly from *molten rock* (Carson and Allen, 1984). Techniques for locating a shallow, crustal magma body, drilling into it, and implanting heat exchangers or possibly direct electrical converters remain to be developed. Neither these experiments nor those of the hot, dry rock type described above are expected to result in economic energy production in the near future. In Iceland, however, where geothermal energy was first tapped for space heating in 1928, economic technology has been demonstrated for extraction of thermal energy from young lava flows (Björnsson, 1980). A heat exchanger constructed on the surface of the 1973 lava flow on Heimaey in the Westman Island group recovers steam which results from downward percolation of water applied at the surface above hot portions of the flow. The space heating system which uses this energy has been operating successfully for over eight years.

It has been customary to speak of *high-temperature* resources as those having temperatures above 150°C, of *intermediate-temperature* resources as those with temperatures ranging from 90°C to 150°C, and of *low-temperature* resources as those with temperatures below 90°C. High-temperature and some moderate-temperature resources are partially amenable to development for electrical power generation, whereas those of lower temperature are usually considered for some direct-heat use such as space conditioning or industrial process heat.

Because preponderant use of geophysics has been in the exploration for and delineation of moderate- and high-temperature hydrothermal resources, we will emphasize such applications. A variety of factors has made the other resource types less attractive for exploration or development. The economics of development of low-temperature resources usually preclude anything beyond a simple, low-cost exploration effort. Discovery of new geopressured resources beyond those known through petroleum exploration has not received attention because the problems of their development center around economic producibility of known resources, not discovery or delineation of new resources. Hot, dry rock resource areas have not been widely sought because of lack of economic interest, although their exploration would present some interesting problems. Efforts by the U.S. Department of Energy to locate and drill into a shallow magma body are just getting started, and only a small overall effort can be expected until and unless these resources someday prove economic.

## Models for high-temperature hydrothermal systems

High-temperature hydrothermal systems are found in many different geologic environments. Because geophysical models cannot be separated from geologic, geochemical, or hydrological models of hydrothermal systems, it is appropriate to comment briefly on general characteristics of hydrothermal systems that are more or less universally accepted. This will help form a context in which to think about the various geophysical targets that a hydrothermal system may present.

Hydrothermal convection systems consist of hot, briny circulating fluids that are highly reactive chemically. Models for such systems have been discussed in White et al. (1971), Mahon et al. (1980), and Henley and Ellis (1983), among others. When a pluton intrudes the shallow crust, it begins to cool by conductive heat loss. If permeability is present, hydrothermal convection develops and dominates the cooling history (Cathles, 1977; Norton, 1984). Meteoric water penetrating to deep levels ($\sim 5$ km) is heated by the intrusive body. The heated water rises toward the surface as a result of its lower density and the hydraulic gradient resulting from cold water exterior to the hot column. The water looses heat as it approaches the surface, and the resulting cooler, denser water flows down the side of the hot column. A convection cell, or a series of cells, is developed.

The bulk of the water and steam in hydrothermal systems is derived from meteoric fluid, with the exception of those few systems where the fluids are derived from seawater or connate brines (Craig, 1963). As the fluids move through the reservoir rocks, their compositions are modified through dissolution of primary minerals and precipitation of secondary mineral assemblages. The waters generally become enriched in NaCl and depleted in Mg. Salinities may range from less than 10 000 ppm total dissolved solids in some volcanic systems to over 250 000 ppm total dissolved solids in basin environments such as the Salton Sea, California (Helgeson, 1968; Muffler and White, 1969; Ellis and Mahon, 1977; Bird and Norton, 1981).

The vertical pressure and temperature gradients in most hydrothermal convection systems lie near the curve of boiling point versus depth for saline water, and sporadic boiling occurs in many systems. Because boiling concentrates such acidic gases as $CO_2$ and $H_2S$ in the steam, the oxygenated meteoric fluids overlying a boiling reservoir are heated and acidified. This process may lead to deposition of clays and the formation of fluids having a distinct $NaHCO_3(-SO_4)$ chemical character.

The general structure of high-temperature systems associated with andesitic stratovolcanoes (e.g., the Cascade Range, U.S.; Ahuachapan, El Salvador), silicic or bimodal volcanic regimes (e.g., Coso, California; Steamboat Hot Springs, Nevada; the Taupo volcanic zone, New Zealand), and sedimentary basins (e.g., the Imperial Valley, California, and Mexicali Valley, Mexico) are shown in Figures 1a, 1b, and 1c. The mineral assemblages produced by the thermal fluids significantly alter the physical properties of the reservoir rocks. The six factors, temperature, fluid composition, permeability, and (to a lesser extent) pressure, rock type, and time, each control the distribution and type of hydrothermal alteration (Browne, 1978). The alteration minerals are strongly zoned in most systems. Beneath the water table, clay minerals, quartz, and carbonate are the dominant secondary minerals below temperatures of about 225°C. Chlorite, illite, epidote, quartz, and potassium feldspar are important at higher temperatures. In the highest-temperature fields (above 250°C), metamorphism to the greenschist or higher facies may occur, resulting in significant densification of the reservoir rocks. Precipitation of silica may occur through cooling of the hot brine. The porosity and permeability of the silicified rocks are thereby considerably reduced, which can effectively seal the sodium chloride reservoir and prevent its expansion or appearance at the surface. However, steam and gas may be able to move through the boundary and interact with the meteoric water above. The product of this interaction is usually a near-neutral pH sodium bicarbonate-sulfate water that forms a hot, secondary geothermal reservoir. Although the bicarbonate-sulfate waters may constitute an exploitable resource, the deep chloride water is the prime hydrothermal resource. In certain areas such as the Geysers, California, the fluid phase in the upper-level rocks is steam and the geothermal system is termed a vapor-dominated system. Beneath such a steam reservoir is presumably a sodium-chloride water resource.

Fumaroles may vent $CO_2$ and $H_2S$ at the surface, which interact with meteoric water to produce highly acidic waters that cause advanced argillic alteration of near-surface rocks. Intense alteration of this type may extend to depths of hundreds of meters below the surface in areas such as Cove Fort-Sulphurdale, Utah, where the water table is deep (Ross and Moore, 1985).

Outflow of the deep NaCl fluid may occur at a considerable distance from the hottest portion of a hydrothermal system. These chloride brines may emerge as boiling springs, frequently surrounded by silica deposits, or as a nonboiling mixture of meteoric $CaHCO_3$ and hydrothermal NaCl fluid. Because of the retrograde solubility of calcite with temperature, the mixed springs frequently precipitate travertine.

Regarding successful exploitation of hydrothermal systems, the key problem appears to be more in locating permeable zones than in locating high temperatures. Grindly and Browne (1976) noted that of 11 hydrothermal fields investigated in New Zealand, all of which have high temperatures (230°C to 300°C), five are nonproductive chiefly because of low permeability. Three of the eleven fields are in production (Wairakei, Kawerau, and Broadlands), and in each of these permeability limits production more than temperature does.

Permeability can be primary or secondary. Primary permeability in clastic rocks originates from intergranular porosity, and it decreases with depth due to compaction and cementation. In volcanic sequences, primary intergranular porosity and permeability exist, but open spaces also exist at flow contacts and within the flows themselves. Secondary permeability occurs in open fault zones, fractures and fracture intersections, along dikes, and in breccia zones produced by hydraulic fracturing (Brace, 1968; Wodzicki and Weisburg, 1970; Moore et al., 1985). Changes in permeability come about through mineral deposition and by leaching by the thermal fluids. Although none of the geophysical methods maps permeability directly, any geologic, geochemical, or hydrological understanding of the factors that control the permeability in a geothermal reservoir can be used to help determine geophysical methods potentially useful for detecting the boundaries and more permeable parts of the hydrothermal system.

## GEOPHYSICAL METHODS FOR GEOTHERMAL EXPLORATION

The discussion in this section covers the application and principal problems encountered in using geophysical methods in hydrothermal exploration. Table 2 is a classification of geophysical methods which also shows the common geothermal targets for each method. In what follows, the emphasis is on application of geophysics to exploration for hydrothermal resources.

### Thermal methods

A variety of thermal methods respond directly to high rock or fluid temperature, the most direct indication of a geothermal resource. Among these methods are measurements of thermal gradient and heat flow, shallow-temperature surveys, snow-melt photography, and thermal-infrared imagery.

**Conventional thermal gradient and heat flow**—Thermal gradient and heat-flow surveys provide basic data about subsurface temperatures, and some program of thermal-gradient drilling is applied in most systematic geothermal exploration throughout the world. Drill holes must be deep enough to penetrate the near-surface hydrologic regime, which may be

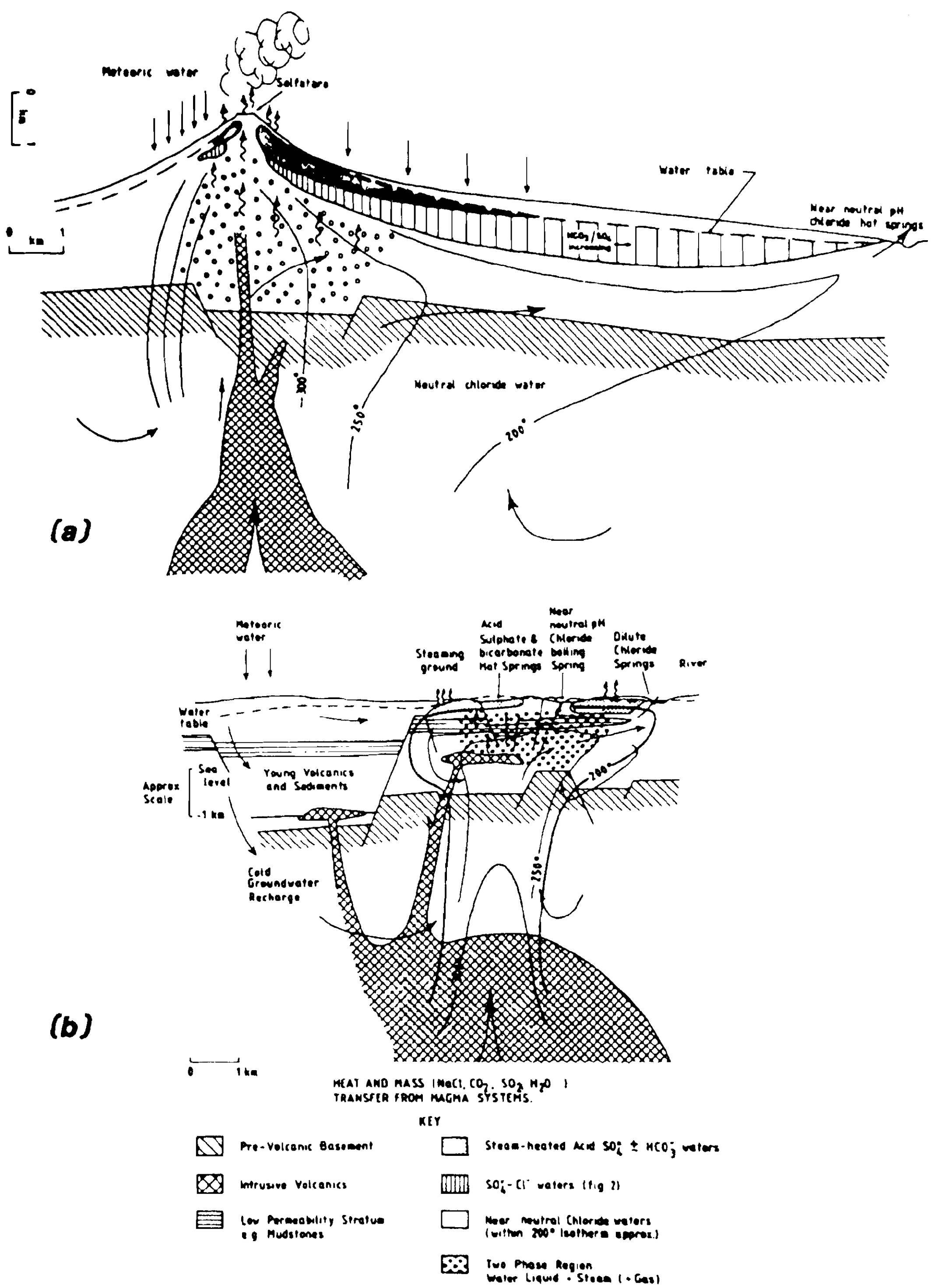
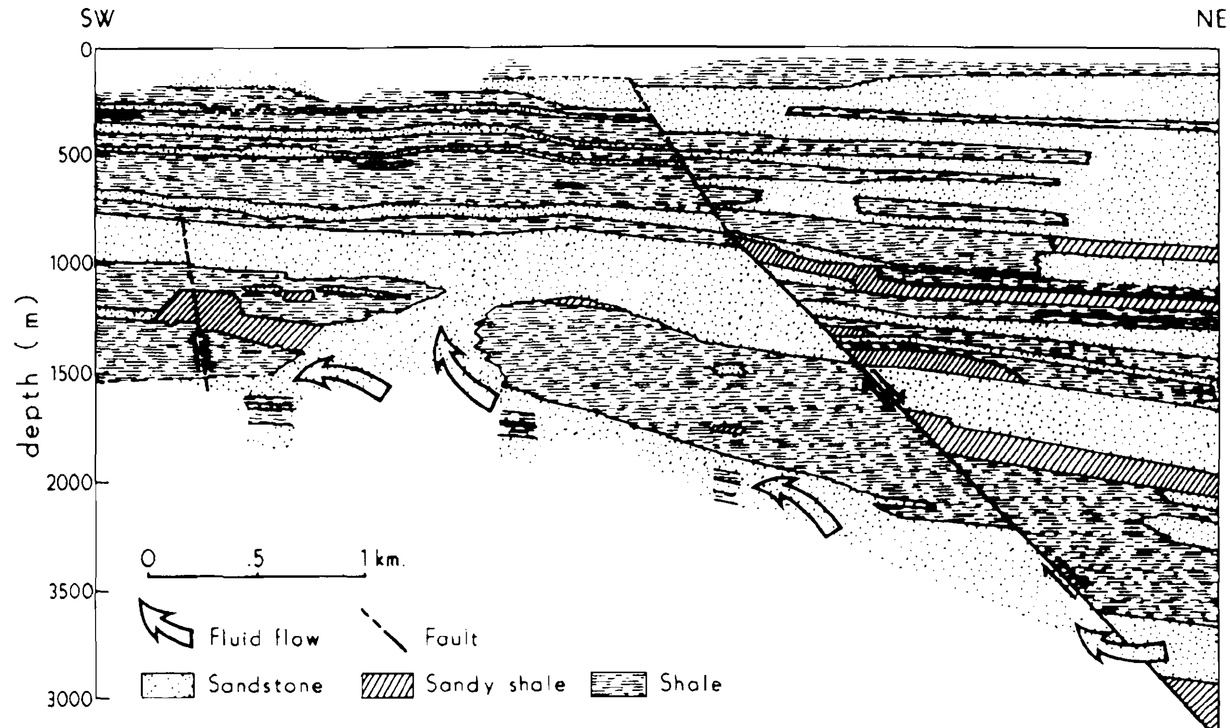



FIG. 1. Schematic geologic models for high-temperature systems. (a) Andesitic stratovolcano with high relief and substantial recharge. Chloride water discharges generally occur some distance from the upflow center and may be revealed by fumaroles, intense rock alteration, and steam-heated aquifers. Near-surface condensation of volcanic gases and oxidation result in acid sulfate waters in the core of the volcano (from Henley and Ellis, 1983). (b) Silic or bimodal volcanic terrains. The geothermal system is supplied by groundwater derived from meteoric water. Heat, gases, chloride, and other solutes and water are supplied by a deeply buried magmatic source which drives the convection system. Mixing between deeper chloride waters, steam-heated waters, and fresh groundwater may result in a variety of hybrid waters (from Henley and Ellis, 1983).

dominated by meteoric recharge with vertical and lateral flow of cold water. In permeable, high-rainfall areas, this flushed zone may exceed 1 km in thickness. The major limitation on the acquisition of thermal-gradient data is imposed by the drilling program. The main factor is drilling cost, but environmental restrictions, land control, permitting, and time involved are other considerations.

Interpretation of temperature, thermal-gradient, and heat-flow data and evaluation of resource potential from these measurements can be quite complex, as discussed in detail in Sass et al., 1971; Blackwell and Morgan, 1976; Chapman and Pollack, 1977; Lachenbruch, 1978; Sass et al., 1981. Heat flow over hydrothermal systems is often 5 to 500 times the regional average. Smith (1983) showed that the Beowawe, Nevada geothermal area is characterized by a wide range of temperature gradient and thermal conductivity values (65–144°C/km, 1.59–5.95 W/m°K) that combine to produce a nearly constant heat flow of 235 mW/m$^2$ above a depth of 1 600 m. At Roosevelt Hot Springs, Utah, the boundary of anomalous heat flow is considered to be the 100 mW/m$^2$ contour, which encompasses an area of more than 175 km$^2$, while the 1 000 mW/m$^2$ contour encompasses an area of about 15 km$^2$ (Ward et al., 1978). A maximum value of about 9 000 mW/m$^2$ occurs over a band 2 km wide parallel to and including the Opal Mound fault. Integration of the heat flux indicates that an estimated 60 MW is being continuously supplied by the source at depth. This is obviously a fairly large geothermal system. For comparison, the small, noncommercial resource at Marysvale, Montana has a maximum surface heat flow of about 800 mW/m$^2$ and the 100 mW/m$^2$ contour encloses an area of about 30 km$^2$ (Blackwell and Morgan, 1976). At East Mesa, California, the 125 mW/m$^2$ contour encloses an area of about 120 km$^2$ and the maximum heat flow value is somewhat over 300 mW/m$^2$ (Swanberg, 1976). At Coso, California, Combs (1980) found that geothermal gradients in 25 holes from 23 to 400 m in depth ranged from 25.3 C/km to 906 C/km with heat flow values ranging from 67 mW/m$^2$ to 964 mW/m$^2$. The area encompassed within the 418 mW/m$^2$ contour was about 150 km$^2$.

Regarding interpretation of heat flow data, early authors recognized that terrain effects and effects of active geologic processes such as uplift or deposition must be compensated; several authors provided methods to do so (Birch, 1950; Blackwell et al., 1980). Continuation of heat flow data to determine subsurface isotherms in geothermal areas was discussed in Brott et al. (1981), who concluded that the depth and shape of the boundaries of the hydrothermal system can be determined by this analysis. A particularly important topic for geothermal exploration is the relationship of measured thermal gradient and heat flow with the local and regional hydrologic regime. Smith and Chapman (1983) gave a review of previous work in this topic and reported on numerical solu-




**(c)**

FIG. (1c). Sedimentary basins (such as Mexicali Valley and Imperial Valley). Hot saline fluids rise along faults and in permeable horizons as shown in this simplified cross-section across the Cerro Prieto field (modified from Halfman et al., 1984). Fluid flow directions are indicated by arrows.

tions of the equations of fluid flow and heat transport used to quantify the effects of groundwater flow on the subsurface hydrothermal regime. In similar fashion, there is a significant effort to understand the mechanisms and effects of heat and mass transfer within hydrothermal systems (Bodvarsson, 1982). Much more work remains to be done at this interface between geophysics on the one hand and hydrology and reservoir engineering on the other.

**Shallow temperature surveys.**—One relatively low-cost method to determine near-surface temperatures is a shallow temperature survey. Temperatures are measured at depths of 2 to 5 m in holes typically drilled by truck-mounted or hand-portable drills or augers, at a low cost per hole. The use of such surveys has been limited because of the uncertainty that the results are related to the temperature distribution at depth. Lovering and Goode (1963), Poley and Van Steveninck (1970), and Kappelmeyer and Haenel (1974) discussed the perturbing effects. These effects are due to (1) diurnal solar heating variations; (2) annual solar heating variations; (3) aperiodic solar heating variations; (4) variations in surface albedo, which affect the amount of energy absorbed; (5) variations in surface roughness, which affect the amount of heat convected away due to turbulent flow of the wind; (6) variations of soil thermal diffusivity; (7) slope and exposure of the terrain; (8) variations in elevation; and (9) variations in level of groundwater and groundwater movement. Temperature variations from these effects are generally negligible below a depth of 20–30 m, with the exception of groundwater movement.

An early example of shallow temperature surveying at a geothermal area was presented by Kintzinger (1956) in his survey of hot ground near Lordsburg, New Mexico. Using thermistors emplaced at a depth of 1 m, he observed a temperature anomaly of some 10°C surrounding a hydrothermal area. Noble and Ojiambo (1976), emplacing thermistors at 1 m depth, helped delineate a geothermal area in Kenya. Lee and Cohen (1979) measured shallow, geothermal gradients at various sites at the Salton Sea, California, which ranged from 0.02°C/m to 4.3°C/m. Lachenbruch et al. (1976) provided a temperature map of the Long Valley area at a depth of 10 m. They concluded that as long as synoptic observations were used at the measuring sites, essentially the same temperature pattern would emerge for contours at a depth of 6 m, and much of it persisted at 3 m. Olmstead (1977) compared 1 m temperature data with 30 m temperature values for the Upsal Hogback area, Nevada, and concluded that the shallow temperature anomaly, without proper corrections, had little correlation with the 30 m temperature anomaly. LeSchack and Lewis (1983) gave a summary of applications of the technique to geothermal exploration along with case histories for Coso Hot Springs, California, Upsal Hogback, Nevada, and Animus

**Table 2. Geophysical targets in geothermal exploration.**

| METHOD | TARGETS |
|---|---|
| HEAT FLOW | Thermally anomalous rocks or fluids. |
| ELECTRICAL | |
| Resistivity | Hot brines, alteration, faults |
| Induced Polarization | Alteration, mineralization |
| CSEM & Scalar AMT | Hot brines, alteration, faults |
| MT/AMT | Hot brines, magma chamber, partial melt, structure |
| Self Potential | Flow of fluid and heat |
| Tellurics | Hot brines, alteration, faults |
| GRAVITY | Structure, alteration, densification, intrusions |
| MAGNETICS | Structure, alteration, rock type |
| SEISMIC | |
| Microseisms | Active hydrothermal processes |
| Microearthquakes | Active faulting and fracturing, distribution of velocity and attenuation |
| Teleseisms | Deep magma chamber |
| Refraction | Structure, distribution of velocity and attenuation |
| Reflection | Structure, distribution of velocity and attenuation |
| RADIOMETRIC | Alteration, $^{226}$Radon, $^{222}$Radium. |
| WELL LOGGING | Anomalous temperature, porosity, permeability, rock type |
| BOREHOLE GEOPHYSICS | |
| VSP | Velocity distribution, fractures |
| Electrical | Hot brines, alteration, faults |

Valley, New Mexico. In the absence of cold water flow in the near-surface, a shallow temperature survey could form the basis on which to plan a shallow or intermediate-depth thermal-gradient program.

**Snow-melt photography and thermal-infrared imagery.**— These temperature-sensitive methods have been used in reconnaissance geothermal exploration in some areas. Snow-melt photography has been used at Coso Hot Springs, California (Koenig et al., 1972) and Yellowstone National Park (White, 1969) to indicate surface areas of slightly elevated temperatures at low survey costs. Color aerial photographs of these areas, taken from hours to days after light to moderate snowfall, made the thermally anomalous areas visible because the snow melted faster over these areas than it did over nonthermal areas.

Airborne thermal infrared (IR) surveys have been used to map the occurrence of warm ground and hot springs in Kenya (Noble and Ojiambo, 1976) and hot springs along the coastline of volcanic islands such as Hawaii (Fischer et al., 1966; Furumoto, 1976). In Kenya the IR survey confirmed several known hot springs and located other, previously unknown areas of hot ground. Later ground-truth surveys determined that more than 90 percent of the areas indicated as anomalous on IR imagery had actual ground temperatures above ambient.

Dickinson (1976) gave an evaluation of the utility of the method at the Taubora geothermal field near Wairakei in New Zealand. Surveys were flown in the late afternoon and at dusk over areas of surface discharge features as well as over urbanization in the town of Taupo. The instrumentation was sensitive in the band 4 to 5.5 μm. Thermograms were interpreted into three temperature ranges: $<1°C$ above ambient temperature; $1°$ to $3°C$ above ambient temperature; and $>3°C$ above ambient. Inspection of aerial photographs and field checking helped eliminate the response of cultural features. Field checks consisted of a series of measurements of soil temperature at depths ranging from 0.05 m to 1 m. Because vegetation over thermally anomalous areas exhibited elevated temperatures, the presence of trees and scrub did not appear to disturb the survey results. The resulting temperature anomaly map was used to indicate areas of thermal discharge and to estimate a total surface heat flow of 111 MW for the area surveyed.

### Electrical methods

Perhaps the most important physical property change due to the presence of a hydrothermal system, other than elevated temperature and heat flow, is the change in electrical resistivity of the rock-fluid volume (Moskowitz and Norton, 1977). Higher temperature increases ionic mobility up to about 300 C, and hence increases conductivity. Ionic conduction in rocks also increases with increased porosity, increased salinity, and increased amounts of certain minerals such as clays and zeolites. Most hydrothermal systems have an associated zone of anomalously low resistivity due to one or more of these factors. At depths exceeding 5 to 15 km, mineral semiconduction dominates aqueous electrolytic conduction, and partial melts and magma may become very conductive (Lebedev and Khitarov, 1964; Shankland and Waff, 1977; Rai and Manghnani, 1978). Although magma is conductive due to mineral semiconduction, the amount of contained water substantially affects the conductivity, dry magmas being much less conductive than wet ones. In geothermal exploration wet magmas may be sought because they have a sufficiently high content of volatile elements to produce the fracturing needed for hydrothermal convection to develop.

Thermal brine and alteration occur predominantly along faults, so electrical 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, these methods can also map faults, stratigraphy, intrusions, and geologic structure in general, independent of the presence of brine or alteration. Hohmann and Ward (1981) reviewed applications of electrical methods in mining exploration, and many points made in their article are also applicable to geothermal exploration.

**Galvanic resistivity.**—The uses of the Schlumberger and Wenner arrays were described in Hatherton et al. (1966), Zohdy et al. (1973), Arnorsson et al. (1976), Stanley et al. (1976), Tripp et al. (1978), and Razo et al. (1980), and others. The Schlumberger array is the most convenient for depth sounding, i.e., estimation of the thicknesses and resistivities of the layers of a horizontally layered earth (Pàlmason, 1976). Successful use of the head-on Schlumberger method was reported in Lezama (1984), and by others working in Iceland. A significant problem with the Schlumberger array, and with galvanic resistivity sounding techniques in general, is the effect of lateral resistivity variations on the measurements. Many, if not most, geothermal areas are characterized by three-dimensional (3-D) resistivity structure at the scale of the electrode separations required for soundings to 1 to 2 km. Although lateral resistivity variations can sometimes be recognized on sounding curves and correctly interpreted by using 2-D or 3-D modeling techniques, there is often not enough data to do this. Using sounding techniques in areas of complex structure or lithologic variations requires great care.

Use of the bipole-dipole array in geothermal exploration was reported in Risk et al. (1970), Beyer et al. (1976a), Keller et al. (1975), Williams et al. (1976), Jiracek and Smith (1976), Stanley et al. (1976), and Souto (1978). Keller et al. (1977) used this method effectively in the reconnaissance exploration for geothermal resources on the East Rift Zone of Kilauea Volcano, Hawaii Island. The bipole-dipole array has been used successfully over broad areas of resistivity lows caused by hydrothermal alteration, but is no longer used because it failed to produce distinctive anomalies over some geothermal systems (Dey and Morrison, 1977). Also, the reduced resistivity values are strongly dependent upon the local resistivity distribution in the vicinity of the transmitting dipole (Frangos and Ward, 1980).

Dipole-dipole arrays were used in surveys reported in Beyer (1977), Fox (1978a), Ward et al. (1978), Baudu et al. (1980), Patella et al. (1980), Wilt et al. (1980) and others. This array is widely used in geothermal, mineral, and petroleum exploration because it permits efficient collection of a large number of data points which when interpreted can separate lateral variations in resistivity from depth variations in resistivity. Numerical modeling programs are widely available to determine the resistivity distribution and intrinsic resistivity values in the subsurface (Dey and Morrison, 1976; Rijo, 1977; Killpack and

Hohmann, 1979). McNitt (1976) recognized the advantage of the dipole-dipole technique in discriminating between vertical and horizontal resistivity boundaries and commented that resistivity surveys were generally by far the most effective of all the geophysical surveys used in the United Nations exploration programs between 1965 and 1975.

Repetitive high-precision dipole-dipole surveys have also been used to monitor changes in the reservoir due to production of the Cerro Prieto geothermal field (Wilt and Goldstein, 1981). A zone in which resistivity increased with time was related to the reservoir and was presumed to be caused by decreasing temperature and salinity from the inflow of fresher, cooler water. Resistivities above and flanking this region showed a systematic decrease with time which was difficult to explain, but which appeared to be related to ascending hot, more saline, fluids at the eastern edge of the producing zone.

**Induced polarization.**—Use of the induced polarization (IP) method theoretically makes possible mapping of the distribution of pyrite and clays, common alteration products in hydrothermal systems. Ward and Sill (1983) recently reviewed application of this method to geothermal exploration. Few IP measurements are reported for hydrothermal areas, and those examined show low-amplitude anomalies with no definite relationship to the hydrothermal system (Zohdy et al., 1973).

**Controlled-source electromagnetics (CSEM).**—Keller (1970) made a baseline review of the applications of active and passive electromagnetic (EM) methods in geothermal exploration. Subsequent illustrations of the success and failure of these methods have been included in Lumb and MacDonald (1970), Keller and Rapolla (1974), Goldstein et al. (1982), and Keller et al. (1981). These methods have been used as an alternative to resistivity methods in some geothermal environments. Time-domain EM methods (TDEM) have been used in volcanic areas of high surface impedance such as Hawaii (Kauahikaua. 1981) where grounded resistivity surveys are slow and costly. Morrison et al. (1978) and Wilt et al. (1981) described a high-power system developed at Lawrence Berkeley Laboratory primarily for geothermal exploration. The primary limitation of these EM methods is that interpretation techniques until recently were available only for the layered-earth, 1-D case. If the subsurface has a 2-D or 3-D resistivity distribution, as it usually does in hydrothermal environments, interpretation using 1-D techniques can produce misleading results.

CSAMT is a subset of CSEM and a subset of AMT, in which the transmitter is a grounded bipole. Two orthogonal, horizontal components of the electric and magnetic fields are measured (as in magnetotellurics). Sandberg and Hohmann (1982) evaluated its use in the Roosevelt Hot Springs, Utah geothermal system. CSAMT's advantages over galvanic resistivity methods are it is sometimes faster and is less affected by lateral resistivity variations when providing sounding information (Ward, 1983a).

**Scalar audiomagnetotellurics (AMT).**—The AMT method uses either natural or artificial EM fields in the 10 Hz to 20 kHz band. Keller (1970), Hoover and Long (1976), Hoover et al. (1978), and Jackson and O'Donnell (1980), among others, reported its use in geothermal exploration. The method has two problems: first, the natural fields are occasionally too weak to obtain useful information, and second (and far more important) the scalar data are totally inadequate for interpretation in 2-D and 3-D terrains in which the tensor AMT method should be used. The CSAMT method is a substantial improvement over scalar AMT because the direction of the inducing fields can be controlled, thus simplifying interpretation in 2-D and 3-D environments. In spite of the interpretational difficulties with scalar AMT data, the technique has been used to produce anomalies that apparently reflect low subsurface resistivity due to hydrothermal systems.

**Tensor magnetotellurics and audiofrequency magnetotellurics (MT, AMT).**—Papers describing application of the tensor MT/AMT method in geothermal areas include Hermance et al. (1976), Stanley et al. (1977), Dupis et al. (1980), Gamble et al. (1980), Musmann et al. (1980), Wannamaker et al. (1980), Berktold (1982), Martinez et al. (1982), and Stanley (1982). Ward and Wannamaker (1983) prepared a comprehensive review of data acquisition, processing, and interpretation for the method, plus a full discussion of the problems it encounters in geothermal exploration.

The tensor MT/AMT method is usually too expensive to be used for mapping the resistivity distribution in the shallow parts of a geothermal system. The method is more logically used to map regional structure, the deeper parts of convective hydrothermal systems, and magma chambers, and to detect and delineate zones of partial melt in the deep crust and upper mantle. Some workers combine telluric current data with a reduced number of MT stations to reduce cost (e.g., Goldstein and Mozley, 1978). In the telluric-magnetotelluric method, a tensor MT station (base) is operated simultaneously with several distant telluric stations (remotes). If the magnetic field is uniform over the area of the stations, magnetic data from the base station, combined in calculations with electric data from the remote stations, yield impedances equal to those resulting from remote tensor MT measurements. The limitation to this method is that while the incident magnetic field may be uniform over large distances, secondary magnetic fields vary considerably, particularly in geologically complex areas. Thus impedances calculated for remote sites may only approximate the true impedances.

MT has been used in most high-temperature resource exploration programs in the western U.S. This use is attributed to the method's advertised depths of exploration and to the assumption that the method can detect the molten or partially molten source of heat. Neither attribute is necessarily correct. The conductivity of magma at elevated temperatures is strongly dependent upon the partial pressure of water (Lebedev and Khitarov, 1964); therefore a dry partial melt is more difficult to detect using MT than a wet partial melt. Depth of exploration depends to a certain extent on the near-surface resistivity structure. The size and other characteristics of the magma body are also important. Newman et al. (1985) explored conditions under which crustal magma bodies could be detected, and concluded that if the body is isolated, i.e., has broken off from conductive magma at depth, it is more easily detected than if it maintains connective roots to the mantle. Also, a carefully performed 2-D or 3-D modeling of the data is required to predict accurately the distribution of resistivities in the subsurface. We attribute the rather limited success of MT in geothermal exploration to inadequate interpretation, poor data quality in some instances, and misapplication of the method.

**Self-potential.**—Spontaneous-potential (SP) anomalies over convective hydrothermal systems arise from electrokinetic and thermoelectric effects, which couple the generation of natural voltages with the flow of fluids and the flow of heat, respectively (Sill, 1983). SP measurements in geothermal areas have shown anomalous regions associated with near-surface thermal zones and faults thought to be fluid conduits (Zohdy et al., 1973; Corwin, 1976; Anderson and Johnson, 1976; Zablocki, 1976; Mabey et al., 1978; Corwin and Hoover, 1979). The signs of SP anomalies can be either positive or negative and the anomalies are often dipolar. Noise in SP surveys arises in telluric currents, electrode drift, topographic effects, variations in soil moisture, cultural noise, vegetation potentials, and electrokinetic potentials due to flowing surface and subsurface water. Although SP surveys are relatively easy to perform, they are difficult to interpret in terms of the nature and location of the source. Sill (1983) developed interpretation techniques that have considerable potential for solving some of these problems.

**Telluric currents.**—The telluric method is most 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 caused, for example, by variations in geologic structure or by hydrothermal systems. The method requires 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, we can map resistivity variations qualitatively (Pálmason, 1976).

The method has been used in geothermal exploration as reported in Beyer (1977), Isherwood and Mabey (1978), Jackson and O'Donnell (1980), and others. Use of the telluric method for regional surveys is convenient for detection of areas worthy of more detailed exploration by resistivity methods. Problems with use of the method include random noise, geologic noise due to overburden, lack of resolution, and the effects of topography; however, the main problem is that it is semiquantitative at best.

## Gravity method

Density contrasts among rock units permit use of the gravity method to map intrusions, faulting, deep valley fill, and geologic structure in general. Gravity surveys are used in the Basin and Range and similar settings as a relatively inexpensive means of obtaining structure and thickness of alluvium. Geothermally related anomalies in the basins are most commonly residual gravity highs interpreted to reflect densification of porous sediments, structural highs, or anomalous geometry of fault zones (Isherwood, 1976; Isherwood and Mabey, 1978).

Gravity has proven useful in locating positive anomalies associated with densification of sediments due to metamorphism and silica deposition in the Imperial Valley of California (Muffler and White, 1969; Biehler, 1971; Elders et al., 1978). At the Broadlands field in New Zealand, the major cause of a positive gravity anomaly is attributed to an increase in density of rocks through alteration and deposition by ascending hot waters (Hochstein and Hunt, 1970). In other areas, gravity highs are expected due to rhyolite domes and hydrothermal alteration (MacDonald and Muffler, 1972). A common association of negative gravity anomalies with granitic intrusion is well-known to mining geophysicists (Wright, 1981). Isherwood (1976) concluded that a large gravity low over the Mt. Hannah area at The Geysers field in California is most likely due to a hot, silicic magma under this area. This interpretation was supported by teleseismic studies (Iyer et al., 1979). Goldstein and Paulsson (1976), Berkman and Lange (1980), and Edquist (1981) found gravity particularly useful in mapping range-front normal faults in the Basin and Range province. Detailed gravity data have delineated major faults that probably control the geothermal fluid flow at Cove Fort-Sulphurdale, Utah (Ross and Moore, 1985). High-precision gravity surveys were used to monitor temporal reservoir changes due to production (Grannell, 1980).

Regional gravity studies and their interpretation may play a major role in understanding the tectonic framework of geothermal systems in the Cascade Range and in other, similar volcanic environments. Bacon (1981) reported a contiguous zone of gravity lows west of the High Cascades in central Oregon and noted that these define major structural trends and delineate fault zones which may localize the movement of geothermal fluids. The zone of gravity lows coincides with (1) an abrupt east-to-west decrease in heat flow from High Cascades values of 100 mW/m$^2$ to values around 40 mW/m$^2$ to the west, and (2) a substantial east-to-west increase in depth to the lower crustal conductor defined by magnetotelluric soundings. Couch et al. (1982) reported similar interpretations. Williams and Finn (1982) reported that large silicic volcanoes with calderas exceeding 10 km diameter produce gravity lows when proper densities of 2 150 to 2 350 kg/m$^3$ are used for the Bouguer reduction, whereas other volcanoes produce gravity highs as a result of higher-density subvolcanic intrusive complexes.

## Magnetic method

Magnetic surveys, either airborne or ground, have been conducted at many geothermal prospects. They can be used for structural or lithologic mapping or for mapping decreases in the magnetization of rocks caused by hydrothermal alteration. Magnetic anomalies in New Zealand geothermal fields have been interpreted as resulting from a conversion of magnetite to pyrite (Studt, 1964). A magnetic low occurs over a part of the hot spring area at Long Valley, and was interpreted in Kane et al. (1976) as the result of magnetite destruction. Such an effect would, of course, remain in extinct hydrothermal systems.

The locations of faults, fracture zones, intrusives, silicic domes, and major alteration areas are apparent on data we examined from Coso Hot Springs, California; from Baltazor, Tuscarora, McCoy, and Beowawe in Nevada; and from Cove Fort-Sulphurdale and Roosevelt Hot Springs in Utah. Magnetics are routinely used in Iceland to delineate dikes, some of which are bordered by zones of high permeability (Pálmason, 1976; Flovenz and Georgeson, 1982).

Magnetic data may also yield regional information of value in exploration. The Monroe Hot Springs, Chief Joseph, Cove Fort-Sulphurdale, and Roosevelt Hot Springs KGRAs are all located in close proximity to a major magnetic discontinuity which trends east-west for a distance exceeding 150 km. This trend reflects the northern margin of the Pioche-Beaver-Tushar mineral belt with many intrusive and volcanic rocks to the south, and thin volcanics overlying thick Paleozoic through Tertiary sediments and few intrusions to the north. The magnetic trend clearly indicates a major tectonic-geologic feature important to geothermal resource localization. Bacon (1981) interpreted major structural trends and fault zones from aeromagnetic data in the Cascades.

Magnetic data can also be used to determine the depth to the Curie isotherm (Bhattacharyya and Leu, 1975; Shuey et al., 1977; Okubo et al., 1985; and many others). These interpretations depend upon many assumptions and therefore have limitations. It is assumed that long-wavelength negative anomalies due to lithologic changes do not significantly perturb the interpretation and that the decreased magnetization of crustal rocks at depth is due to temperatures above the Curie point rather than to deep-seated lithologic changes. In addition, because the bottom of a magnetized prism is not accurately determined, accuracy of individual Curie-point depths can be poor. Nevertheless, the Curie point analysis can be useful in regional exploration.

### Seismic methods

**Microseisms.**—Two methods have been proposed to utilize microseisms for delineating geothermal reservoirs. The first is based on the speculation that hydrothermal processes radiate seismic energy in the frequency band 1 to 100 Hz. If this phenomenon exists, the exploration method becomes a rather straightforward "listening" survey, using stations on a 0.5 to 2 km grid. Contours of noise power on the surface should delineate noise sources. This is the standard noise survey sometimes used in geothermal exploration. Noise in the 1 to 10 Hz band sometimes arises from nearby cultural sources such as traffic, trains, rivers, wind, etc. It is also known that seismic noise amplitudes are usually higher over alluvium and soft sedimentary basins than over hard rock. Thus, noise power anomalies may merely reflect a local increase in sediment cover. Ground noise surveys have yielded high levels of noise over Taupo, New Zealand (Clacy, 1968), The Geysers (Lange and Westphal, 1969), and in the Imperial Valley (Douze and Sorrells, 1972).

A second approach interprets the noise field as propagating elastic waves of appropriate type and uses the propagation characteristics to make inferences about the source. Iyer and Hitchcock (1976) postulated that seismic waves radiating from a hydrothermal source a few kilometers deep may propagate as body waves and thus can, in principle, be distinguished from cultural microseisms which generally propagate as surface waves. Seismic arrays can determine the phase velocity of microseisms and thereby distinguish body waves emanating from deep sources and exhibiting high phase velocities (typically exceeding 4 km/s) from surface waves.

There is limited evidence that body waves do exist in association with geothermal occurrences. Liaw and Suyenaga (1982) detected high-velocity body waves in data recorded at Beowawe, but they did not detect body waves at Roosevelt Hot Springs. Liaw and McEvilly (1979) failed to find body waves at Leach Hot Springs, Nevada, but they did find microseismic energy propagating as fundamental-mode Rayleigh waves from the vicinity of the thermal manifestations. Their paper additionally presented the foundations for proper survey design and data analysis. Oppenheimer and Iyer (1980) found microseisms at two recording sites near Norris Geyser basin, Yellowstone National Park that were propagating from near-surface sources in the geyser basin as both surface and body waves in the frequency range 1.4 to 6.3 Hz. The low phase velocities, 1 to 4 km/s, appear to preclude body waves originating from deep hydrothermal sources in the basin. It is apparent that careful data collection and analysis are required to produce valid results using microseismic techniques.

**Microearthquakes.**—Microearthquakes are frequently related to major hydrothermal convection systems. Accurate location of these earthquakes can provide data helpful in locating active faults that may channel hot water toward the surface (Ward and Björnsson, 1971; Lange and Westphal, 1969; Hamilton and Muffler, 1972; Ward et al., 1979). Microearthquake (MEQ) surveys have been completed in several geothermal areas including Iceland (Ward and Björnsson, 1971), East Mesa (Combs and Hadly, 1977), Coso (Combs and Rotstein, 1976), and Wairakei (Hunt and Lattan, 1982).

*P*- and *S*-wave velocities may be retrievable from microearthquake data. Gupta et al. (1982) used microearthquake data to obtain regional *P*- and *S*-wave velocities for The Geysers. Ideally, 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 $Q$ 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 McEvilly (1979) found a high $Q$ value in the production zone at The Geysers from microearthquake and refraction surveys, whereas they found a lower $Q$ deeper in the crust from a refraction survey. Majer (1978) reported that a refraction survey yielded high $Q$ values at Leach Hot Springs due to silica densification of sediments. Gertson and Smith (1979) found high $Q$ values over the geothermal system at Roosevelt Hot Springs, using refraction data.

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 ($\sigma < 0.20$). The ratio 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. Thus, determination of Poisson's ratio from microearthquake (MEQ) surveys can conceivably result in determining whether a hydrothermal reservoir is vapor or water dominated. Majer and McEvilly (1979) and Gupta et al. (1982) noted Poisson's ratios of 0.13 to 0.16 over the production zone at The Geysers, California, and values 0.25 and higher outside of it. The low Poisson's ratio in part corresponds to a decrease in *P*-wave velocity.

For any of the above analyses of MEQ data, a good model of the subsurface velocity distribution is required. Lack of velocity control is a principal problem in analysis of MEQ data. Some geothermal systems, such as Roosevelt Hot Springs, have a generally low, episodic occurrence of microearthquakes (Zandt et al., 1982). Swarms of earthquakes occur, but in the intervals between them, insufficient activity may preclude any of the foregoing analyses. Indeed, passive seismic data can be recorded for a two- or three-week period or longer and the erroneous conclusion may result that the geothermal system is unimportant since it is not seismically active during the time of recording.

**Teleseisms.**—If a sufficiently distant microearthquake is observed with a closely spaced array of seismographs, changes in $P$-wave traveltime from station to station can be due to velocity variations near the array. A magma chamber beneath the geothermal system would give rise to low $P$-wave velocities and hence to late observed traveltimes. Steeples and Iyer (1976a, b) found relative $P$-wave delays of 0.3 s at stations in the west-central part of the Long Valley caldera. Reasenberg et al. (1980) recorded relative $P$-wave delays of 0.2 s at Coso. Iyer et al. (1979) found relative $P$-wave delays as large as 0.9 s at The Geysers. Robinson and Iyer (1981) reported relative $P$-wave delays up to 0.3 s at Roosevelt Hot Springs. One may speculate that relative $P$-wave delays are caused by partial melts or magmas, as at Coso, Long Valley, and The Geysers, but 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) suggested 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.

**Refraction.**—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 ratios. Detection of a characteristic attenuation or a "bright spot," as found over reservoirs in petroleum exploration, would be a useful feature (Ward et al., 1979; Applegate et al., 1981), but this has not been reported.

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., 1976; Hill, 1976; Majer, 1978; Ackermann, 1979; Gertson and Smith, 1979). Hill et al. (1981) reported a 270 km profile from Mount Hood to Crater Lake in the Cascades and presented results in terms of crustal velocity structure. These data contribute to a better understanding of regional geology and are indirectly used in geothermal exploration.

The seismic refraction method does not resolve structure as well as 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., 1978). Majer and McEvilly (1979) reported locally high $P$-wave velocities in the production zone at The Geysers as determined from refraction surveys. Beyer et al. (1976b), Majer (1978), and Gertson and Smith (1979) found anomalous velocities and amplitudes of refracted waves passing through the reservoir regions at Grass Valley, Leach Hot Springs, and Roosevelt Hot Springs, respectively. Majer (1978) interpreted the high $Q$ determined at Leach Hot Springs as resulting from silica densification of sediments.

**Reflection.**—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 and Mexicali Valley types. However, where the structure becomes highly faulted or folded, diffraction of seismic waves occurs at discontinuities and makes the task of interpreting structure difficult.

Conventional reflection seismic surveys give good definition of Basin and Range border faulting and depths to the base of alluvial fill at Roosevelt Hot Springs, UT and Soda Lake, San Emidio, Dixie Valley, 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 yields substantial structural information along the range front (Ross et al., 1982). At Beowawe, extensive and varied digital processing was ineffective in eliminating the ringing due to a complex near-surface intercalated volcanic-sediment section (Swift, 1979). Majer (1978) found reflection data useful in delineating structure in Grass Valley, NV. At Soda Lake, in 1977, Chevron obtained 1 200 percent CDP seismic reflection coverage. The seismic data delineated a complex northeast-southwest trending graben from the shore of Soda Lake passing south of Upsal Hogback. The reflectors dip to the southwest, consistent with the presence of a small basin over the gravity low. The maximum depths of reliable seismic data are governed by a thin basalt unit and vary from 730 to 1 220 m (Hill et al., 1979).

Zoback and Anderson (1983) demonstrated use of seismic reflection data in mapping the style of initial faulting, infill, and subsequent slumping and faulting in some basins in the Basin and Range province. Denlinger and Kovach (1981) showed that seismic-reflection techniques applied to the steam system at Castle Rock Springs (The Geysers area) were potentially useful for detecting fracture systems within the steam reservoir, as well as for obtaining other structural-stratigraphic information. Beyer et al. (1976b) reported on the value of seismic-reflection profiling for mapping concealed normal faults associated with the Leach Hot Springs geothermal system, Grass Valley, NV. Blakeslee (1984) processed seismic-reflection data obtained by the Comisión Federal de Electricidad over the Cerro Prieto, Mexico geothermal field, and was able to define subtle fault features and other important velocity features related to hydrothermal effects.

### Radiometric methods

Gamma-ray spectrometry may be used to map the areal distributions of $^{40}K$, $^{238}U$, and $^{232}Th$. If $^{226}Rn$ or $^{222}Ra$ is present in a geothermal system, it will be detected in the $^{214}Bi$ peak, since they also are daughter products of $^{238}U$ decay. An examination of hot-spring waters in Nevada indicates varying abundances of $^{226}Rn$ and $^{222}Ra$ in spring systems where $CaCO_3$ is the predominant material being deposited. Systems where silica predominates are relatively low in radioactivity (Wollenburg, 1976). The use of alpha-cup detectors for radon emanating from hydrothermal systems was reported in Wollenburg (1976) and Nielson (1978). Surface radon emission surveys are capable of detecting open channels that may conduct geothermal fluids. However, the method has had little use in geothermal exploration.

### Geophysical well logging

Much research still must be done to increase the understanding of the responses of various well logs in goethermal reservoirs and their typically fractured, altered, commonly igneous, and metamorphic host rocks. In spite of the relative lack of knowledge of well-log response in geothermal reservoirs, several logs or log combinations have been used successfully to investigate such properties as lithology, alteration, fracturing, density, porosity, fluid flow, and sulfide content, all of which may be critical in deciding how and in what intervals to complete, case, cement, or stimulate the well.

Many logging techniques used by petroleum and mining industries have been adopted or modified for use in geothermal exploration and development programs. The major differences in usage are the requirements of high-temperature tools and the different interpretation required for hard-rock (volcanic, igneous) lithologies. Other differences include a strong emphasis on fracture identification and the effects of hydrothermal alteration upon certain log responses. Discussions of these items and interpretation of well-log suites from various geothermal areas are numerous (Glenn and Hulen, 1979; Keys and Sullivan, 1979; Sanyal et al., 1980; Glenn and Ross, 1982; Halfman et al., 1982).

**Borehole geophysics**

The class of techniques called borehole geophysics requires a combination of surface and in-hole sources and/or receivers or sources and receivers in separate boreholes.

**VSP.**—The least experimental of the borehole geophysical techniques is vertical seismic profiling (VSP) using both *P*- and *S*-wave surface sources (usually mechanical vibrators) arranged circumferentially around the well. Direct and reflected waves are detected by means of strings of downhole geophones clamped to the well wall, or by hydrophones. VSP has been used mainly to trace seismic events observed at the surface to their point of origin in the earth and to obtain better estimates for the acoustic properties of a stratigraphic sequence (Balch et al., 1982). Oristaglio (1985, this issue) presents a guide to the current uses of VSP. Gal'perin (1973) presented a review of VSP research in the USSR including results of three-component VSP (*P*- and *S*-wave sources with three-component detectors) to estimate compressional-shear velocity ratios and Poisson's ratio. An *S*-wave shadow zone was detected following one hydrofracturing operation at 700 m (Fehler et al., 1982). On the basis of data from three shotpoints, a finite-difference model showed that the shadow data fitted other information about the hydrofracture. However, due to the low-frequency *S*-wave source and the long wavelength of the *S*-wave (60 m) in the medium, it was apparent that the fractured region was required to have large dimensions (a few wavelengths) for this shadow effect to occur.

There has been some interest in developing methodologies to derive fracture permeability information from tube waves (Paillet, 1980). Crampin (1978, 1984) and others have argued that VSP conducted with three-component geophones might prove extremely useful for mapping the fractured conditions of rocks if seismic anisotropy information were extracted from the shearwave splitting effect.

**Electrical techniques.**—Borehole-to-borehole and borehole-to-surface resistivity methods may also be applicable to geothermal exploration. Daniels (1983) illustrated the utility of hole-to-surface resistivity measurements with a detailed study of an area of volcanic tuff near Yucca Mountain, Nevada. He obtained total-field resistivity data for a grid of points on the surface with current sources in three drill holes, completed a layered-earth reduction of the data, and interpreted the residual resistivity anomalies with 3-D ellipsoidal modeling techniques. Yang and Ward (1985) presented theoretical results relating to detection of thin oblate spheroids and ellipsoids of arbitrary attitude. The theoretical model results indicate that cross-borehole resistivity measurements are a more effective technique than single-borehole measurements for delineating resistivity anomalies in the vicinity of a borehole.

Beasley and Ward (1986) obtained interesting results in their numerical mise-à-la-masse studies. The dip of the body and the location of the energizing electrode within it were both varied. The maximum depth at which a body could be located and still produce a detectable surface anomaly was dependent upon the position of the buried electrode and upon the contrast in resistivity between the body and the host. They found that locating the buried electrode just outside the body did not significantly alter the results from those when the electrode was embedded in the inhomogeneity.

From the above studies we tentatively conclude that (1) the cross-borehole method produces larger anomalies than does a single-borehole method; (2) cross-borehole anomalies using a pole-pole array are smaller than those for a cross-borehole dipole-dipole array; (3) the cross-borehole mise-à-la-masse method produces larger anomalies than for the other cross-borehole methods; and (4) anomalies due to a thin sheet were generally much smaller than those for a sphere, as expected (e.g., Dobecki, 1980).

Surface-to-borehole EM in which a large transmitter is coaxial with the well and a downhole detector is run in the well may provide useful information on the location of conductive fractures intersecting the wellbore. Whether this technique will work in cased wells and whether a "crack" anomaly can be distinguished from a stratigraphic conductor are topics under study. Dyck and Young (1985, this issue) provide a more complete review of the various borehole methods.

**Geotomography.**—Geotomography is a term applied to any of several geophysical methods which use multiple transmitter and receiver positions in a borehole-to-borehole or borehole-to-surface array to effect a detailed imaging of subsurface physical properties. Bois et al. (1972) reviewed a number of early well-to-surface, surface-to-well, and well-to-well measurements and refined the implementation and interpretation of well-to-well seismic measurements. A current example of seismic imaging by surface and borehole techniques was given in Rundle et al. (1985). A consortium of institutions collaborated to perform three experiments an expanding spread profile about a fixed common midpoint, a conventional CPD reflection profile, and a VSP profile of one drill hole—all at the Long Valley caldera, CA. Integrated interpretation of the data yielded a cross-section of the caldera showing the ring fracture system near Minaret Summit, the configurations of postcaldera volcanics and welded Bishop Tuff, and the interface with basement rock of the Sierra Nevada. Such detailed studies show great promise of providing structural details in the geothermal environment.

Lager and Lytle (1977) adopted the technique for high-frequency EM measurements between boreholes. Daily et al. (1982) and Daily (1984) described the application of borehole-to-borehole measurements using radio frequency signals between 1 MHz and 40 MHz to map the EM attenuation of oil-shale. The method appears successful for mapping rubbled zones resulting from explosions and retorted zones within the oil-shale.

Geotomography using electrical methods may contribute to fracture delineation in geothermal environments. Nabighian et al. (1984) described cross-hole magnetometric resistivity (MMR) measurements in which massive sulfide mineralization was mapped at a depth exceeding 500 m. The advantages of mapping current flow in a plate-like body by locating the magnetic detector in a borehole were illustrated by numerical models. Yang and Ward (1985) presented numerical modeling results which illustrated that cross-borehole resistivity measurements are much more effective than single-borehole measurements for detecting deeply buried fractures and ore deposits. Their model results suggested that the depth, dip, and strike of conductive fracture zones could possibly be determined by tomography in suitably placed boreholes.

## COSO GEOTHERMAL AREA—GEOPHYSICAL STUDIES

It seems appropriate in this review to illustrate a few selected geophysical data sets in the geothermal environment. Because of space limitations relative to the large amount of data available, we chose just one area whose geology is well-known and where drilling has established the presence of a significant high-temperature convection system. The Coso geothermal area, Inyo County, southeastern California (Figure 2) provides an instructive example where both regional and detailed geophysical data contribute to an understanding of the geothermal resource.

### Geologic setting

The Coso geothermal area is located in the Coso Range of the western Basin and Range province immediately east of the southern Sierra Nevada. Regional geologic mapping of the area was completed by Duffield and Bacon (1977), who expanded the results of several earlier workers. Northerly trending fault-block mountains are formed of diverse lithologies which vary in age from Precambrian through Holocene. The oldest rocks are complexly folded Precambrian through Early Mesozoic marine sedimentary and volcanic rocks, many of which are regionally metamorposed (Hulen, 1978). This older sequence is intruded by Jurassic-Late Cretaceous granitic stocks and plugs which appear to be portions of the southern Sierra Nevada batholith. Late Cenozoic volcanic rocks were erupted in two periods, 4.0–2.5 m.y. and $\leq$ 1.1 m.y. (Duffield et al., 1980), and formed domes, flows, and pyroclastic deposits which covered many crystalline rocks in the Coso geothermal area. Hulen (1978) completed detailed geologic mapping and alteration studies of approximately 40 km$^2$ of the immediate Coso geothermal area in support of the U.S. Department of Energy drilling program at well CGEH-1. A generalization of his map, Figure 3, provides a useful reference base for our evaluation of the geophysical data. Hulen (1978) and Duffield et al. (1980) described hydrothermal alteration and active thermal phenomena (fumaroles, steaming boreholes, and "warm ground") which occur throughout an irregular 20 km$^2$ area along the eastern margin of the Coso rhyolite dome field. Drill hole CGEH-1 was drilled to a depth of 1 470 m in 1977, primarily in a mafic metamorphic sequence and a leucogranite which intruded the metamorphic rocks. This hole indicated temperatures in excess of 177°C and convective heat flow which appeared to be limited to an open fracture system between depths of 564 and 846 m (Galbraith, 1978). Subsequently, several successful drill holes completed by California Energy Corporation have established the presence of a hydrothermal system.

### Thermal studies

The Coso geothermal area is well expressed in quantitative thermal data. Combs (1980) completed a comprehensive study of the heat flow as determined in 24 shallow (35–110 m) and 2 deeper boreholes in an area of approximately 240 km$^2$ centered about the rhyolite dome field. He measured thermal gradients ranging from 25.3°C/km to 906°C/km, which he attributed to convecting hot water and former convective transport of heat by dikes that fed the domes and flows. Terrain-corrected heat-flow values ranged from 67 to 960 mW/m$^2$. Figure 4, from Combs' (1980) study, presents the heat flow in the upper zone (15 m $\leq$ depth $\leq$ 65 m). A map of heat flow for a deeper interval (35 m $\leq$ depth $\leq$ 300 m) shows a similar pattern with somewhat reduced heat-flow values. The heat-flow anomaly is principally confined to the east-central portion of the rhyolite dome field and trends northeast to include Coso Hot Springs. The anomalous heat flow terminates abruptly along the north-trending range front fault which passes through Coso Hot Springs, and along a northwest trend 3 km north of Devil's Kitchen.

LeSchack et al. (1977) and LeSchack and Lewis (1983) described shallow temperature surveys completed at Coso. The

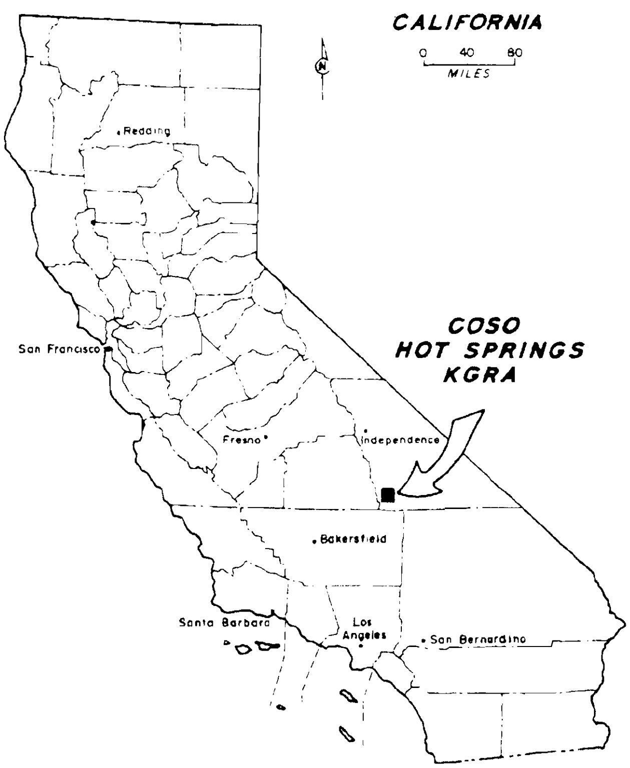



FIG. 2. Location map, Coso Hot Springs geothermal area, California.

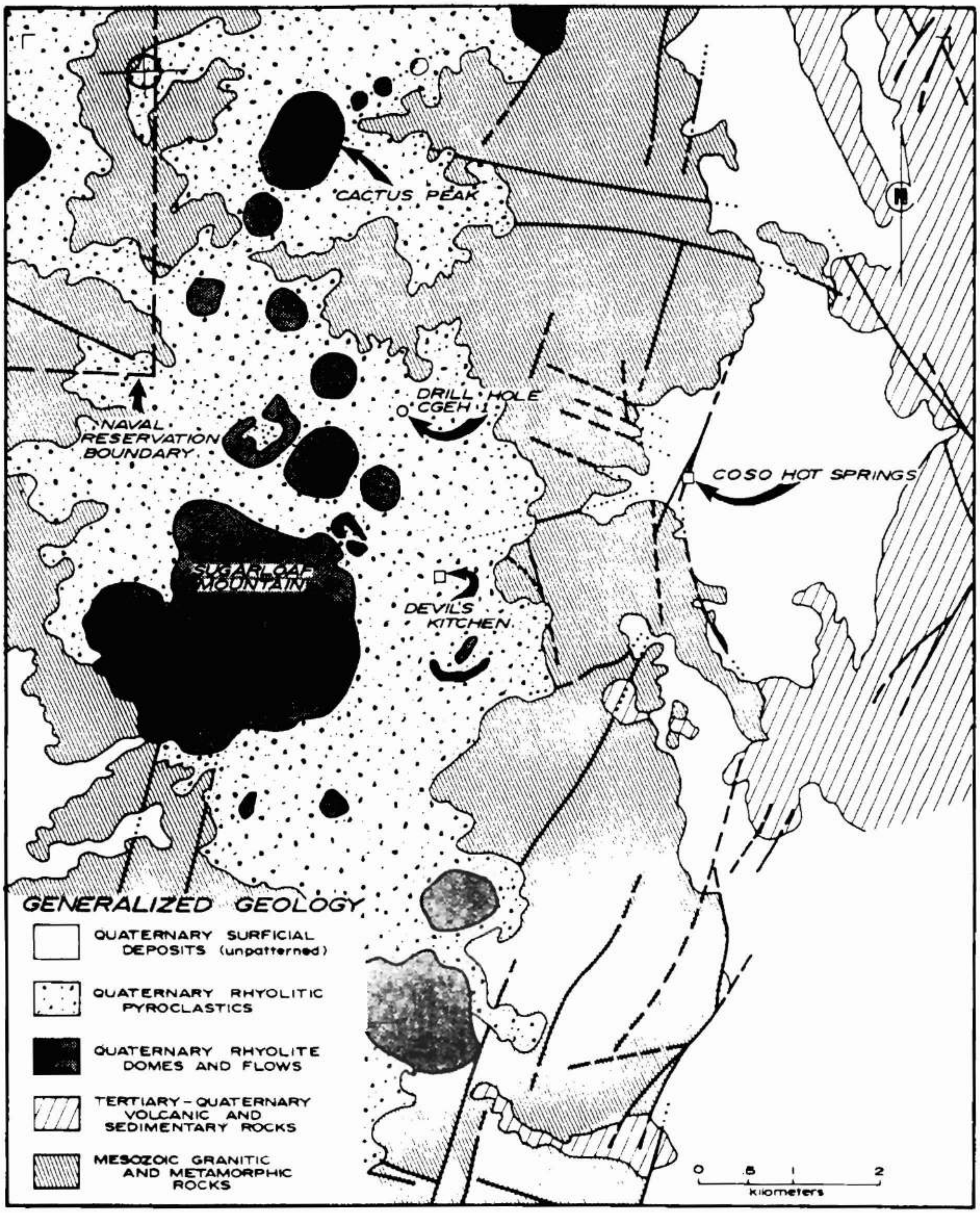



FIG. 3. Geology of the Coso geothermal area, California (after Hulen, 1978).

shallow (2 m) temperature measurements were made with a thermistor probe backfilled in a 2 m deep augered hole after the thermistor equilibrated with surrounding earth temperatures. Figure 5 shows the 2 m temperatures, corrected for elevation, for this survey. Temperatures of approximately 27.4 to 31.7 C form an anomaly pattern quite similar to the 400 mW/m$^2$ heat flow contour of Combs (1980) shown in Figure 4. LeSchack and Lewis (1983) described a more complete data reduction which yields residual anomalies exceeding 8 C for the 2 m temperatures at Coso.

### Electrical surveys

Schlumberger vertical electric soundings (VES), telluric mapping, and AMT soundings have been completed in an area of approximately 900 km$^2$ which included the Coso geothermal area (Jackson et al., 1977). Jackson and O'Donnell (1980) presented an interpretation of these and other data and noted a close correlation between the 7.5 Hz AMT low, the VES data, and the 400 mW/m$^2$ anomaly of Combs (1980).

The University of Utah Research Institute completed more detailed dipole-dipole resistivity surveys in September, 1977 as part of the U.S. Department of Energy resource assessment program which included the drilling of CGEH-1 (Fox, 1978a; Fox et al., 1978). A grid of three north-south lines and six east-west lines was surveyed to map the resistivity structure of a 41 km$^2$ area. An electrode spacing of 300 m was used for 41 line-km of survey, and a 150 m spacing for an additional 13 line-km. Figure 6 shows survey line locations and a map of apparent electrical resistivity obtained by contouring the $n = 3$ (third separation) values of the 300 m dipole data. The data represent an average apparent resistivity distribution for the surface to about 150 m depth, rather than a more specific intrinsic resistivity distribution that could be obtained by numerical modeling. The $\leq 15\ \Omega \cdot$m low-resistivity zone includes Coso Hot Springs, Devil's Kitchen, and much of the $\leq 16\ \Omega \cdot$m resistivity low defined in Jackson et al. (1977) using AMT and VES, although some additional detail is indicated in Figure 6.

Figure 7 presents the observed apparent resistivity along the central portion of line 1, an east-west profile, which crosses the rhyolite domes on the west, the northern edge of Devil's Kitchen, and approximately 1 000 m south of Coso Hot Springs. The corresponding geologic section is shown as presented in Hulen (1978). Extremely high resistivities (100–3 540 $\Omega \cdot$m) were mapped along the north side of Sugarloaf Mountain, a rhyolite dome. Rather uniform low apparent resistivities (5–10 $\Omega \cdot$m) to the east occur where alluvium overlies granitic rocks. The central portion of the line is dominated by low to moderate (10–30 $\Omega \cdot$m) resistivities, and resistivity increases with in-

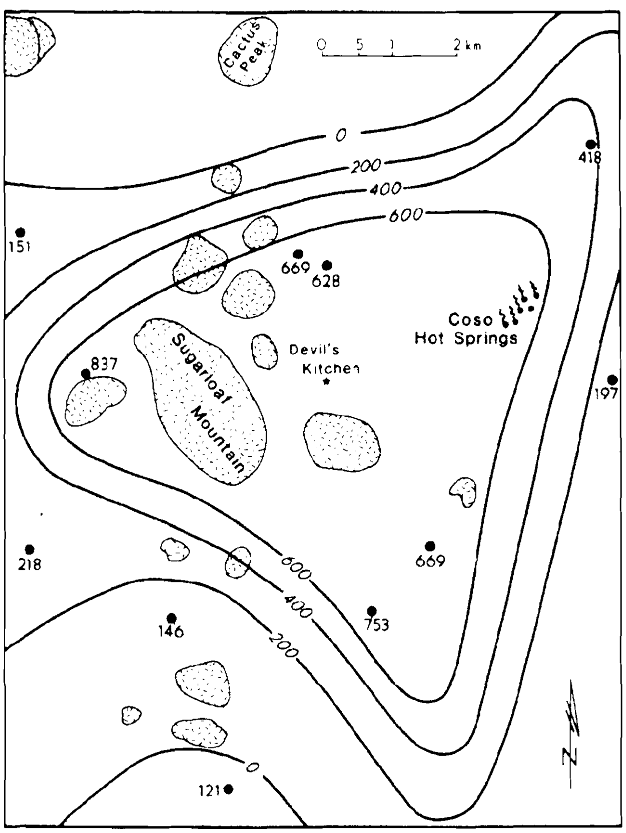



FIG. 4. Heat flow map of the Coso geothermal area, California (after Combs, 1980). Heat flow values and contours in milliwatts per square meter for the upper zone (15 m $\leq$ depth $\leq$ 65 m).

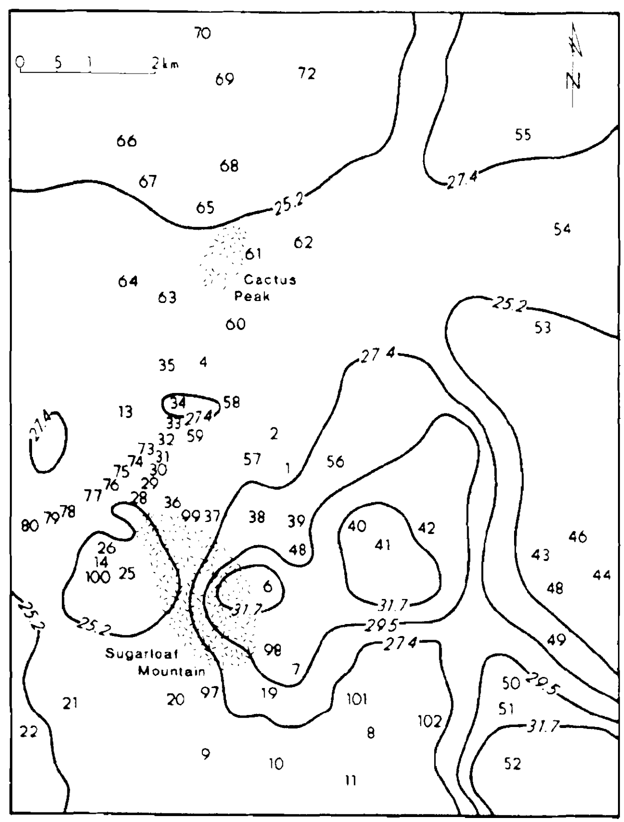



FIG. 5. Shallow-temperature measurements at the Coso geothermal area, California (after LeSchack and Lewis, 1983). Temperatures in degrees Centigrade at 2 m depth.

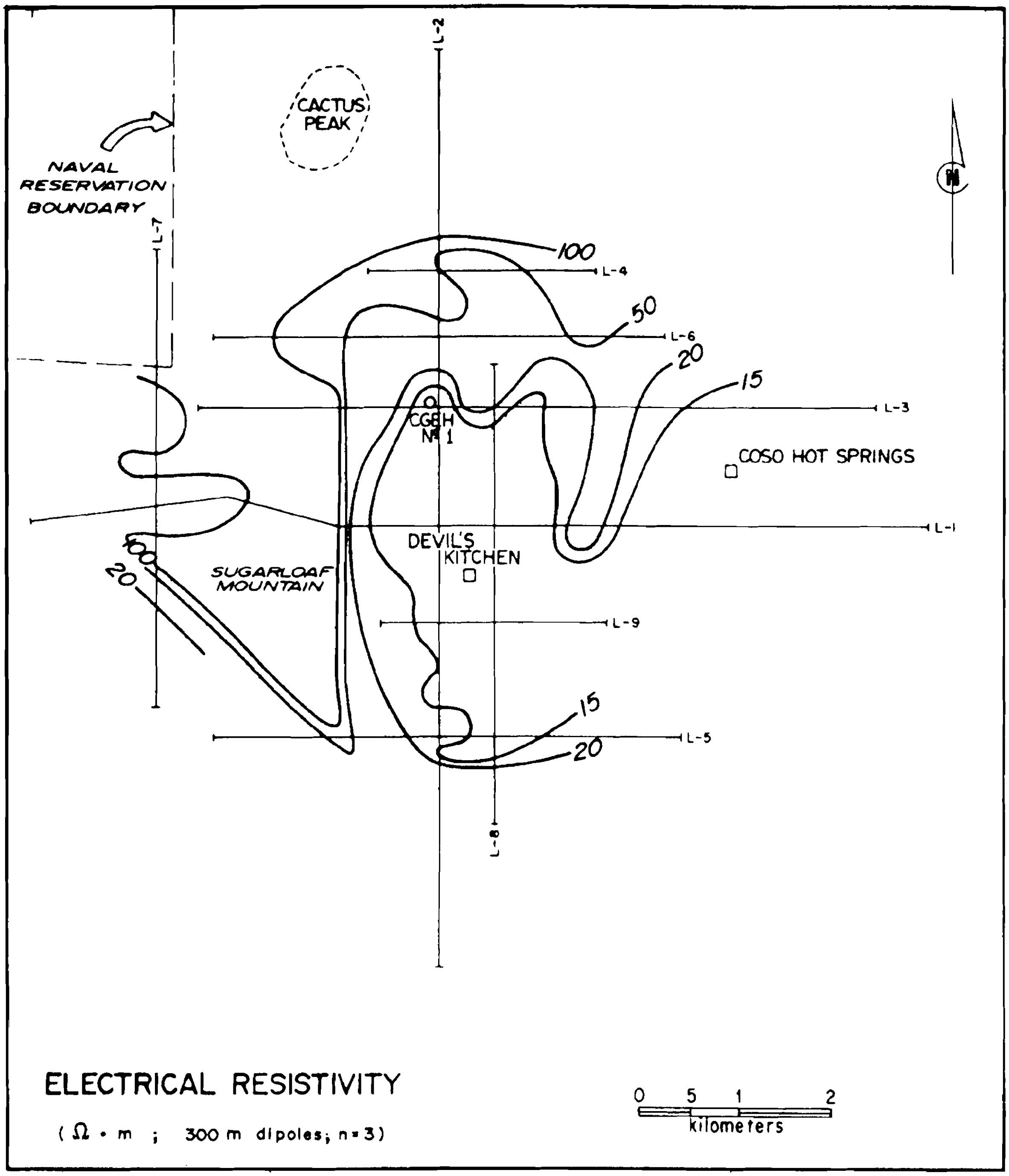



FIG. 6. Dipole-dipole electrical resistivity survey of the Coso geothermal area, California. Dipole lengths of 300 and 150 m were used. The contoured apparent resistivity in ohm · meters is shown for third separation ($n = 3$) values of the 300 m dipole lines (after Fox, 1978a).

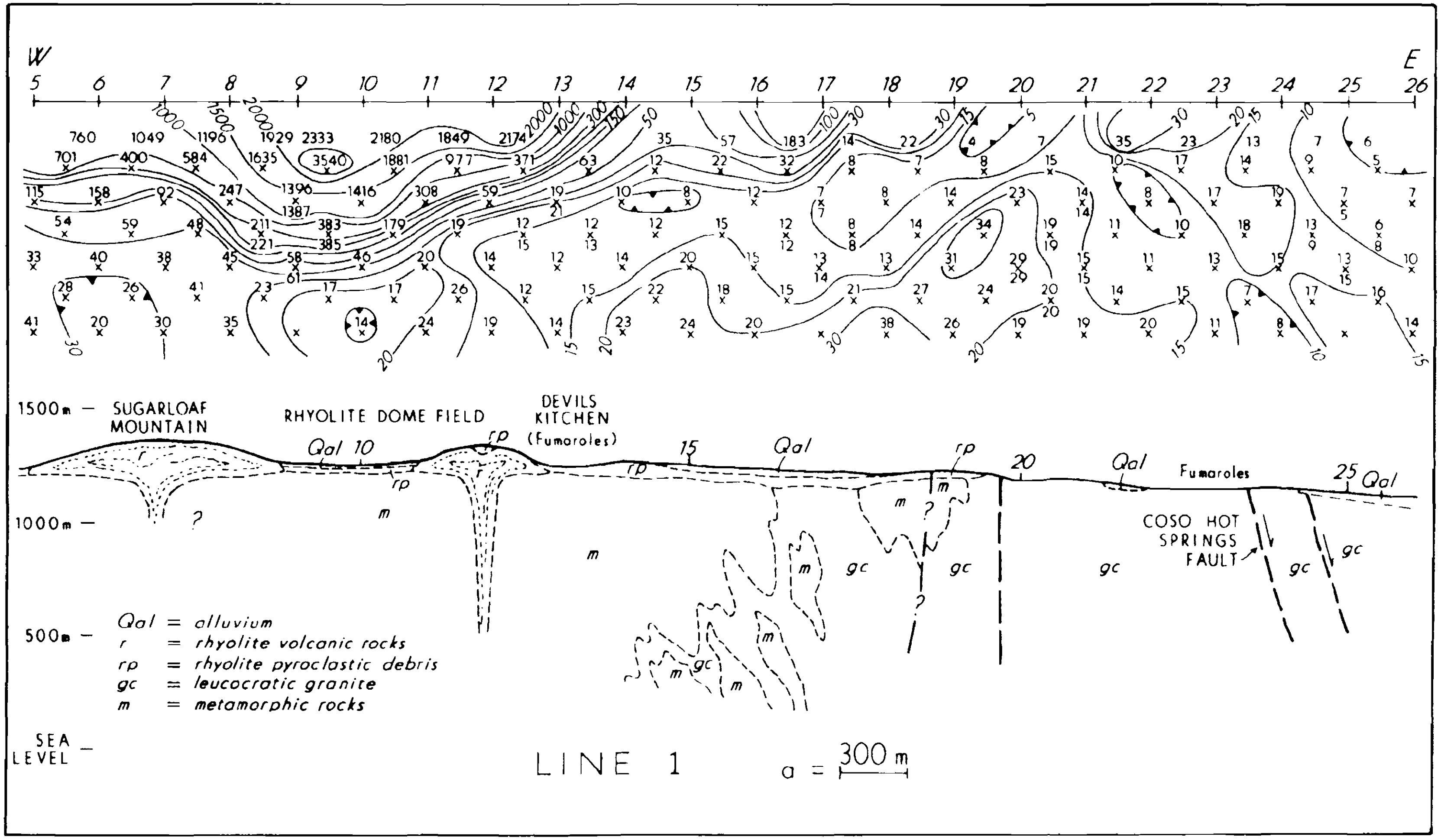



FIG. 7. Observed apparent resistivity ($\Omega \cdot m$) and geologic cross-section for line 1 (300 m dipoles) at Coso geothermal area. Line location shown in Figure 6. Geologic cross-section is modified from Hulen (1978); resistivity data from Fox (1978a).

creasing separation. Occasional lower resistivities (4–10 Ω · m) occur near Devil's Kitchen and along an east-northeast trending fracture zone between stations 18 and 21. Although these are not very low resistivities compared with many geothermal areas, the values are abnormally low for metamorphic and granitic rocks, and probably indicate alteration and fluids which are largely confined to irregularly spaced fracture zones (Fox et al., 1978).

### Detailed aeromagnetic survey

A detailed low-altitude aeromagnetic survey of 927 line-km was completed over the Coso area by the University of Utah Research Institute for the U.S. Department of Energy in September 1977 (Fox, 1978b). The data were recorded on north-south flight lines with a 400 m line spacing at a mean terrain clearance of approximately 230 m (Figure 7). In his interpretation of the detailed survey data Fox (1978b) identified more than 40 specific magnetic sources, most of which could be related to reduced terrain clearance and/or mapped rock type changes. Most of the rhyolite domes are expressed as positive magnetic anomalies in part as a result of the reduced terrain clearnace.

Basement lithologic and structural information is apparent even when the data are presented at a 200 nT contour interval, as shown in Figure 8. Fox (1978b) identified northeast- and northwest-trending basement magnetic discontinuities which correspond in part to mapped faults and structural trends proposed by other authors. Most significant, however, is a broad magnetic low up to 500 nT below background which covers about 26 km$^2$ in the southeast intersection of the two major trends. Rock magnetization measurements, geologic mapping, and alteration studies indicate that the magnetic low was caused partially by magnetite destruction resulting from hydrothermal alteration by the geothermal system as well as by primary lithologic changes at depth.

### Seismicity

The region that includes the Coso Range and the southern Sierra Nevada is one of the more active seismic areas in southern California, as summarized in Walter and Weaver (1980). The seismicity of the Coso geothermal area was reported in some detail in Combs and Rotstein (1976) and in Walter and Weaver (1980) who also provided an in-depth summary of many previous studies. Combs and Rotstein (1976) recorded several hundred earthquakes on an array of three-component seismographs during an operating period of three weeks. Most of the 78 events which were located occurred at depths of 5 to 10 km just north of Sugarloaf Mountain and at depths of 1 to 3 km in the vicinity of Coso Hot Springs. In 1975, Walter and Weaver (1980) established a 16-station seismographic network for an area approximately 40 km north-south by 30 km east-west in the Coso range as part of the U.S.G.S. studies to evaluate the geothermal resource potential. They recorded 4 216 local earthquakes ($0.5 < m < 3.9$) during the first two years of operation. Many of these events occurred in a 520 km$^2$ area which included Coso Hot Springs (CHS), Devil's Kitchen, and the rhyolite domes as shown in Figure 9. Included in this seismicity were six earthquake swarms, four of which were spatially related to the rhyolite field, and two swarms which occurred along the Coso Basin fault system. Walter and Weaver (1980) concluded that the Coso Hot Springs area itself was not characterized by any unusual seismic activity, in contrast to the earlier study reported in Combs and Rotstein (1976). The Walter and Weaver (1980) very detailed study characterized the seismicity and faulting within the rhyolite field and identified the fault system between the rhyolite field and the adjacent Coso Basin as an important tectonic boundary. but it was considered insufficient to determine the geothermal production capability of the fault system.

In an accompanying paper, Young and Ward (1980) presented a 3-D attenuation model for the Coso Hot Springs area as determined from teleseismic data. They determined that a shallow zone of high attenuation exists with the upper 5 km in the Coso Hot Springs-Devil's Kitchen-Sugarloaf Mountain area which they believed corresponds to a shallow vapor-liquid mixture, or "lossy," near-surface lithology. No zone of significantly high attenuaton was interpreted for the 5 to 12 km depth interval, but high attenuation was noted below 12 km. Reasenberg et al. (1980) analyzed teleseismic $P$-wave residuals and mapped an area of approximately 0.2 s excess traveltime which they attributed to a low-velocity body between 5- and 20-km depth in the area of high heat flow and hydrothermal activity. They hypothesized that the low-velocity body could be caused by the presence of a partial melt in the middle crust.

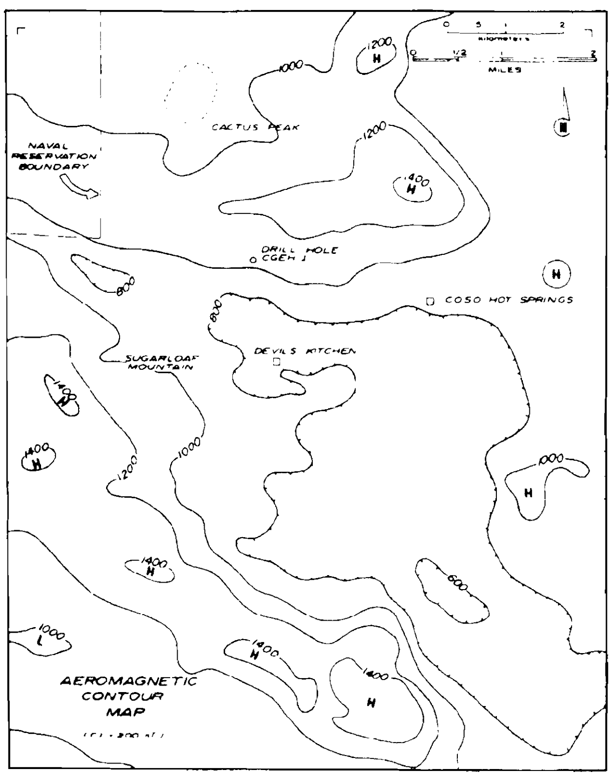



FIG. 8. Low-altitude aeromagnetic survey of the Coso geothermal area, California. Contour interval is 200 nT. Modified from Fox (1978b).

**Integrated anomaly summary**

Figure 10 summarizes the spatial overlap of the magnetic and resistivity lows, the 400 mW/m$^2$ heat flow anomaly, and the anomalous ($\geq 26$ C) ground temperatures at 2 m depth. The data are superposed on alteration and thermal features mapped in Hulen (1978). The prospect areas as indicated by the various data sets are generally in good agreement, except perhaps for the extension of the heat flow high north of the belt of active thermal phenomena. Locations of several successful wells drilled by California Energy Corporation are also shown. This drilling confirmed the presence of a high-temperature convective hydrothermal system in which the fluids are confined to major fracture zones within the crystalline rocks. Active exploration continues in the Coso area, and future geoscientific studies and drilling will continue to improve our model of the hydrothermal system.

## ANALYSIS OF WORLDWIDE APPLICATION

In conjunction with ongoing research, we recently conducted a computer-aided bibliographic search to determine the worldwide applications of geophysics in geothermal exploration and development using the GEOREF data base. The 554 references selected were supplemented by approximately 200 additional references obtained through specific literature search. A total of 47 countries or geographic regions and 88 geothermal resource areas are represented. Geothermal exploration in the US comprised 59 percent of the reference list, introducing a significant bias in the data set. Italy is next with 5 percent, and five other countries, the USSR, Japan, Mexico, Iceland, and New Zealand, each provide somewhat less than 5 percent. References for these seven countries comprise 80 percent of the list.

Considering all resource areas and temperatures, 11 methods, including three different galvanic resistivity arrays, have had significant utilization: galvanic resistivity sounding or VES (59 percent), gravimetric (52 percent), temperature gradient (50 percent), heat flow (48 percent), magnetic (39 percent), MT (35 percent), dipole-dipole resistivity (33 percent), reflection seismology (33 percent), MEQ (32 percent), remote sensing (28 percent), and bipole-dipole resistivity (26 percent). The least-used methods included CSAMT, IP, pole-dipole resistivity, earth noise, and geomagnetic soundings. The greatest utili-

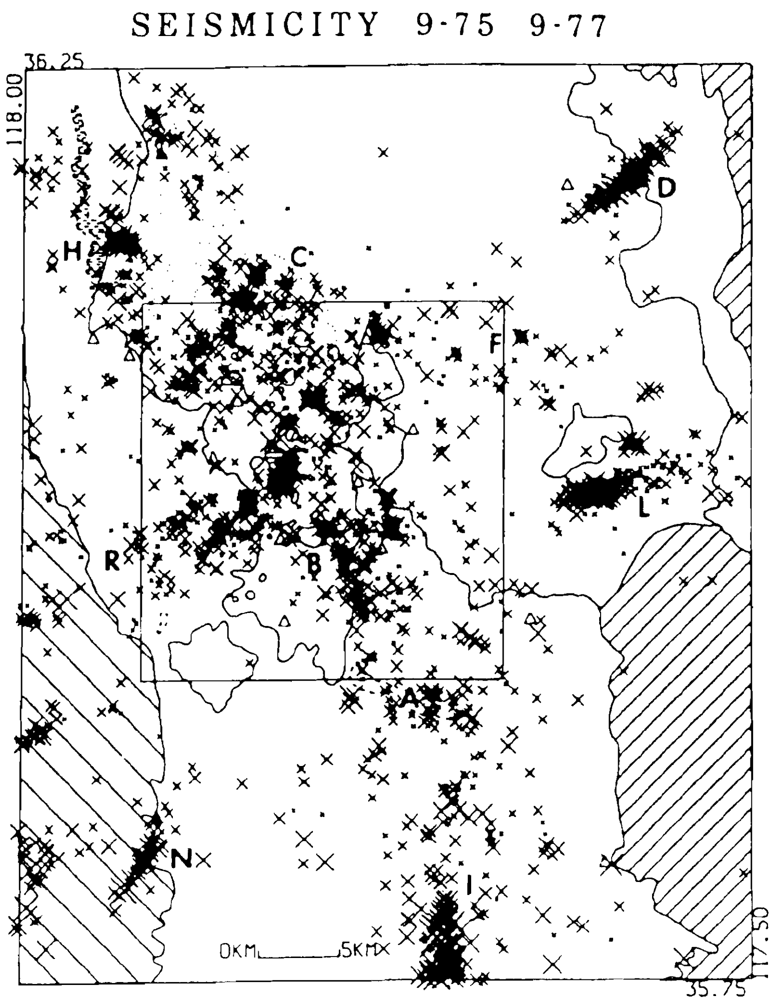



FIG. 9. Located earthquakes in the Coso Range, September 27, 1975 to September 30, 1977. Box includes geothermal subarea shown in other figures. R indicates Red Hill seismic zone, B locates seismic zone on west side of Coso Basin. Cross-hatching designates Sierra Nevada on west and Argus range on the east. From Walter and Weaver (1980).

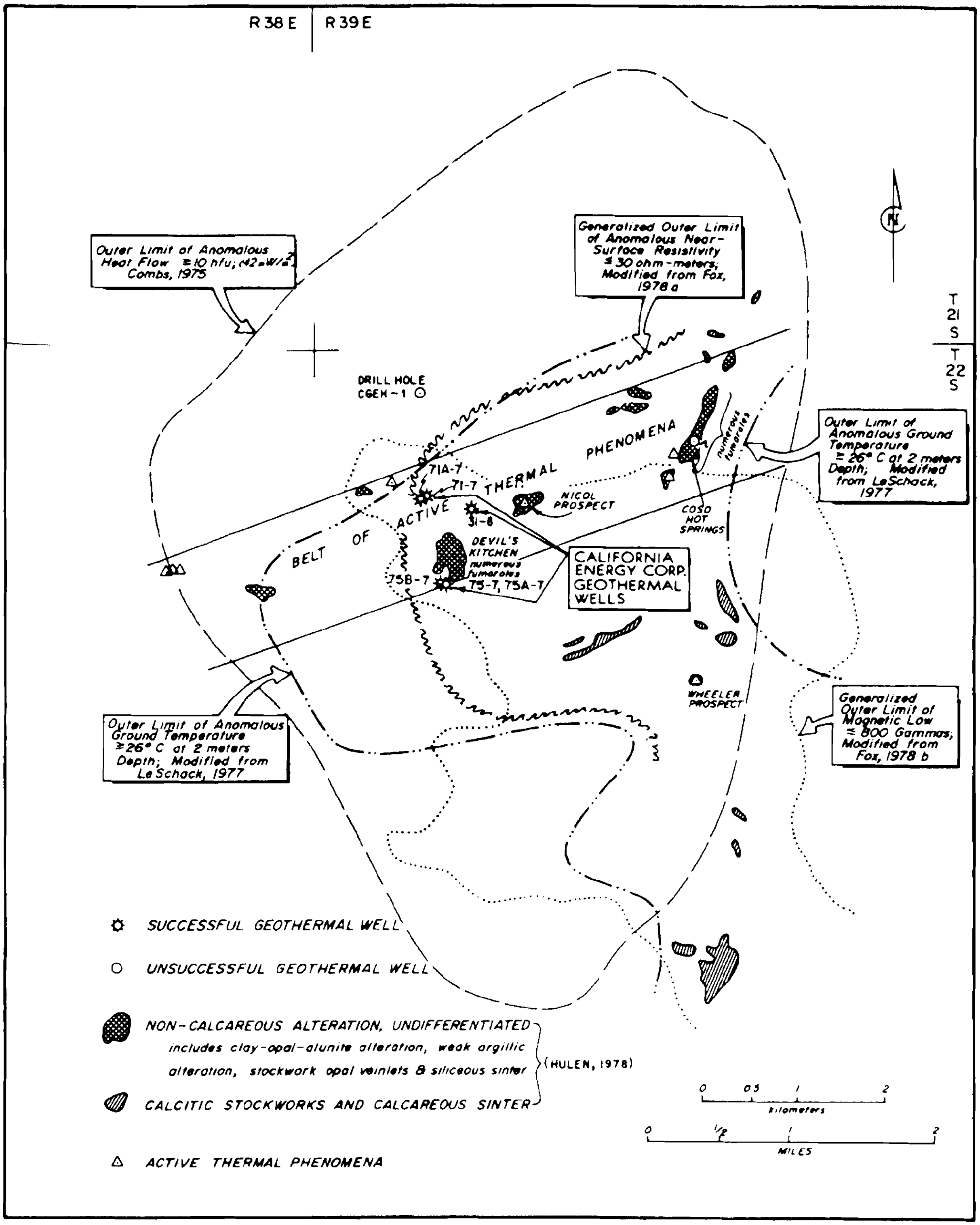



FIG. 10. Geophysical anomaly summary for the Coso geothermal area. Thermal manifestations, alteration areas, and drill holes are shown for reference (after Hulen, 1978).

zation of geophysics was in the exploration for moderate- and high-temperature resources where the potential value of the resource justifies application of costly exploration methods. The exploration for lower-temperature resources, in addition, was often limited to the immediate vicinity of potential users, as in the Paris basin. An average of six different geophysical methods were utilized in the 88 resource areas.

The results are summarized on the basis of geologic setting in Table 3. Some elaboration of this tabulation is warranted. Vertical electric soundings (VES) had significant usage in all five geologic environments, probably due to the familiarity of the method and its general suitability for depth sounding in reconnaissance exploration programs. It is less suited to detailed surveys in complex geologic environments, however, than other galvanic resistivity arrays or controlled-source EM methods.

The gravimetric method was widely used in all geologic environments largely because of its low unit area cost for reconnaissance surveys and its utility in defining geologic structure. We suspect that its usage for the detection of geothermal densification is a small portion of the total usage.

Thermal-gradient or heat-flow data were reported in most of the more comprehensive studies, as would be expected. An indication of only moderate usage in extrusive and rift valley environments results in part from our separate tabulation of the two specific thermal methods, although in some studies both were identified. The tabulation may also indicate the problems with these methods in volcanic areas such as the Cascades and the Snake River plain, where high recharge rates or overlying cold water aquifers reduce the effectiveness of the thermal methods or greatly increase the costs of acquiring meaningful data because deep holes are required.

The dipole-dipole method received moderate usage in all but the basin environment, and the bipole-dipole method received moderate usage in all except the basin and intrusive environments. In many resource areas these methods supplemented earlier VES surveys.

Magnetotelluric methods were employed in all environments at a moderate level, perhaps replacing the use of dipole-dipole or bipole-dipole surveys in some basin or igneous environments. Reflection seismology was also reported at a moderate utilization in these two environments.

One obvious criticism of this compilation is that the level of use of a method does not necessarily indicate its value as an exploration tool. Too often a technique that was successfully employed in one environment is tried in other geologic settings where it is not appropriate. Ward (1983b) provided an evaluation of the geophysical methods used in the exploration of geothermal resources in the Basin and Range Province of the western U. S. He evaluated 14 methods in 13 high-temperature sites (including Long Valley, Coso Hot Springs, Roosevelt Hot Springs, and Raft River) and concluded that: (a) none of the various geophysical methods was uniformly consistent in performance; (b) none of the methods could be ranked in the good category and only five methods were ranked in the good-to-fair category (MEQ, gravimetric, electrical resistivity, SP, and heat flow/TG); (c) the least effective methods were earth noise, magnetic, and MT; and (d) no combination of any four methods was ranked as good-to-fair in success at more than one site.

Table 3 indicates only a moderate selectivity in applying the geophysical methods to different geologic environments. Certainly, understanding the geology and the probable physical property contrasts, perhaps by forming a preliminary conceptual model of the resource type as illustrated in Figures 1a, 1b, and 1c, is the key to the cost-effective use of geophysics in geothermal exploration.

Other considerations became apparent during our study which suggest a greater selectivity in the application of geophysics than may be apparent in a general tabulation such as Table 3. The literature reports geophysical studies on three quite different scales: reconnaissance or regional; detailed or prospect scale; and finally, the reservoir or near-borehole scale.

The field of exploration and assessment of geothermal resources is a relatively young one which has evolved rapidly by drawing on preexisting petroleum and mining technology. Exploration to date has focused on high-temperature systems, many of which had some surface expression. The field has already progressed to much deeper and totally blind systems. These are factors worthy of consideration as we try to evaluate the past usage of geophysical methods in geothermal exploration and as we evaluate the state-of-the-art.

## STATE-OF-THE-ART

We now briefly evaluate the state-of-the-art in application of geophysics to exploration for and within moderate- and high-temperature geothermal systems. Our comments are summarized in Table 4.

### Thermal methods

Instrumentation for measuring thermal conductivity seems to be adequate. A better understanding of the variations of this parameter with temperature, pressure, porosity, and the effects of hydrothermal alteration is needed, however. Instrumentation for precise measurement of temperature downhole is adequate for temperatures below about 250°C, but above this temperature several components of usual borehole systems begin to fail in the corrosive hydrothermal environment. Logging equipment rated for higher temperature is needed for study of the higher-temperature parts of hydrothermal systems as well as for hot-rock environments such as those encountered in deep continental scientific drilling. Referring to interpretation techniques, it seems to us that continued work on the understanding of regional and local hydrologic effects on temperature measurements is needed to understand observed thermal-gradient and heat-flow patterns. Continued development of 2-D and 3-D algorithms to model jointly hydrology and heat transport in complex geologic situations including uplift, deposition, erosion, faulting, extension, and intrusion is needed.

We believe that available equipment and interpretation techniques for shallow-temperature surveys are adequate and that experience reported in the literature is sufficient to facilitate decisions on whether or not to apply this technique in specific exploration problems. The thermal IR technique has seen only limited use in hydrothermal exploration, and this will continue to be the case as exploration emphasizes the search for concealed resources.

### Electrical methods

There is no wholly satisfactory electrical method for exploring concealed resources in rugged volcanic terrains. Galvanic

Table 3. Utilization of geophysical methods in geothermal exploration.

UTILIZATION OF GEOPHYSICAL METHODS IN GEOTHERMAL EXPLORATION

| | |
|---|---|
| Rift Valley: | significant - VES method |
| | moderate - MEQ, gravimetric, magnetic, MT, dipole-dipole, bipole-dipole and heat flow and TG methods |
| Basin and Range: | significant and moderate - all methods except geomagnetic soundings, CSAMT, HEP, pole-dipole, IP, and BG methods |
| Intrusive: | significant - gravimetric, magnetic, VES, and temperature gradient methods |
| | moderate - reflection seismology, AMT, MT, dipole-dipole and heat flow methods |
| Extrusive: (volcanic) | significant - gravimetric and VES methods |
| | moderate - MEQ, reflection seismology, magnetic, MT, dipole-dipole, bipole-dipole, SP, heat flow, TG, and remote sensing methods |
| Basin: | significant - gravimetric, VES, heat flow and TG methods |
| | moderate - reflection seismology, MT and telluric methods |

Abbreviations used:

AMT - audiomagnetotelluric
CSAMT - controlled source audiomagnetotelluric
MT - magnetotelluric
BG - borehole geophysical
HEP - horizontal electrical profiling
IP - induced polarization
MEQ - microearthquake
SP - self-potential
TG - thermal gradient
VES - vertical electrical profiling

resistivity surveys, which are relatively easy to run and for which interpretation methods are reasonably well worked out, often lack adequate depth of penetration. Scalar AMT, which is easy to run and for which highly portable equipment is available, does not provide enough data to resolve the subsurface resistivity structure adequately and typically lacks depth penetration. The tensor MT AMT method is able to resolve complex structure better, but uses very sophisticated, marginally portable equipment and requires a highly trained crew and complex, sophisticated interpretation. Therefore MT AMT is generally too costly for simultaneously providing large area (reconnaissance) and detailed survey coverage. The CSEM methods are relatively easy to run but equipment is only marginally portable and adequate 2-D and 3-D interpretation is only now becoming available. SP surveys are easy and inexpensive but quantitative interpretation is difficult and often ambiguous. In view of the relevance of electrical methods to geothermal exploration, further development of electrical equipment and techniques specifically for the geothermal environment seems like a wise research investment. Edwards and West (1985, this issue) and Smith (1985, this issue) also address the state-of-the-art of electrical methods.

Resistivity methods are adequately developed. The 2-D and 3-D interpretation algorithms are available but are still not in universal use. CSEM methods show the potential to contribute much more to hydrothermal exploration than they have so far. There is need for development of state-of-the-art portable equipment and for the continued development of 2-D and 3-D interpretation techniques. These geophysical techniques also need further field evaluation. Much of the data collected in the past were interpreted using layered-earth models, and we consider these models to be inadequate in most geothermal environments. The scalar AMT technique is adequately developed and tested and we conclude it has only limited use in geothermal exploration. Tensor MT and AMT are still classed as untested techniques. Adequate equipment has not been available for very long, and many of the first geothermal applications suffered from poor data quality. In addition, the majority of the interpretation has been done using layered-earth models, which are, as we have stated before, totally inadequate for the geothermal environment, especially when the scale of the measurement is considered.

The spontaneous-potential method lacks interpretation techniques to facilitate the level of information needed to decide whether or not to drill test an anomaly. Substantial gains in quantitative interpretation theory have been made since 1980, but the algorithms are probably in limited use. If the technique can be understood better, more field evaluation might be warranted. We consider that the telluric current method has limited application because of its semiquantitative nature and is sufficiently developed and understood.

### Gravity and magnetic methods

Paterson and Reeves (1985, this issue) provide a current evaluation of the state-of-the-art of the gravity and magnetic methods. These methods both seem to be developed adequately for routine application to geothermal exploration problems. Advances in instrumentation and interpretation will continue and will be adapted as appropriate for geothermal use.

### Seismic methods

The *microseismic* methods lack adequate field testing, largely because of the poor level of understanding of survey design and data analyses prior to 1980. Continued work on data processing and interpretation as well as further testing in geothermal environments is warranted. Microearthquake surveys have potential to contribute to defining drill targets especially for deep or blind hydrothermal systems. Less expensive equipment is needed so field deployment time can be increased to mitigate to some extent the episodic nature of the phenomenon. Equipment, interpretation, and field testing of the *teleseismic* and *refraction* techniques are deemed adequate for routine application where appropriate, although we recognize that advances will continue.

The *seismic reflection* method has potential for greater contributions to geothermal work than it has made to date. Although the method will always be expensive per unit of coverage, if the information derived could be increased an adequate payout may result. Portable, high-resolution gear is now becoming available for shallow reflection work in hard-rock environments. Better techniques of data acquisition and processing are needed for use in volcanic terrains, many of which are considered bad recording areas even after years of research by the petroleum industry. Laster (1985, this issue) evaluates current seismic data acquisition capabilities and Schultz (1985, this issue) discusses seismic data processing in detail. Interpretation of steep structures and supression of diffraction effects (Stolt and Weglein, 1985, this issue) are needed in the geothermal environment. Much of the required research and development will probably be provided by the petroleum industry as it has been in the past.

### Radioactive methods

Conventional radioactive methods will not likely play a significant role in geothermal exploration and, therefore, we deem them adequately developed. Grasty et al. (1985, this issue) discuss current interpretation practices for multichannel gamma-ray data.

### Well logging

There are significant needs for both new equipment development and for new interpretation techniques in well logging, and these needs are summarized in Sanyal et al. (1980) and in Lawrence Berkeley Laboratory (1984). The main instrumentation problem is lack of downhole tools for logging in slim holes at geothermal temperatures. Most tools are limited to temperatures below 175 C to 200 C, although a few have capability to 260 C. Neither tools nor cable exists for temperatures above 300 C. This lack of high-temperature downhole instrumentation seriously compromises the quantity of data that can be obtained in many of the hydrothermal systems currently under production or development, and will also compromise information on some of the deep research drill holes currently being planned under the Continental Scientific Drilling Program. Regarding interpretation, few of the available tools are calibrated for the hard-rock environment, and quantitative interpretation techniques remain to be worked out for many of the measurements. In summary, relatively little of the well logging sophistication available to the petro-

Table 4. Research and technique development needs in geophysical methods.

| METHOD | INSTRUMENTATION | INTERPRETATION | EXPERIENCE |
|---|---|---|---|
| **Thermal Methods** | | | |
| Heat Flow/Gradient | Need temp logging for T > 250°C | Need models for hydrologic effects, heat transport in hydrothermal environment | Adequate |
| Shallow Temperature | Adequate | Adequate | Adequate |
| Thermal IR | Adequate | Adequate | Adequate |
| **Electrical Methods** | | | |
| Resistivity/IP | Adequate | Need broader use of 2-D, 3-D techniques | Adequate |
| CSEM | Need portable equipment, reduced costs | Need to develop 2-D, 3-D interpretation techniques | Need better evaluation |
| Scalar AMT | Adequate | Adequate | Adequate |
| Tensor MT/AMT | Need portable equipment, reduced costs | Need broader use of the 2-D, 3-D techniques available | Need better evaluation |
| SP | Adequate | Need better interpretation techniques | Need better evaluation |
| Telluric Current | Adequate | Adequate | Adequate |
| Gravity Method | Adequate | Adequate | Adequate |
| Magnetic Method | Adequate | Adequate | Adequate |
| **Seismic Methods** | | | |
| Microseisms | Adequate | Need further development of techniques | Need better evaluation |

TABLE 4 (cont.)

| METHOD | INSTRUMENTATION | INTERPRETATION | EXPERIENCE |
|---|---|---|---|
| Microearthquakes | Need reduced equipment/survey costs | | Need better evaluation |
| Teleseisms | Adequate | Adequate | Adequate |
| Refraction | Adequate | Adequate | Adequate |
| Reflection | Need portable high resolution equipment | Need continued development and application of petroleum technology at reduced costs | Need better evaluation |
| Radioactive Method | Adequate | Adequate | Adequate |
| Well Logging | Need logging tools for T > 225°C | Need interpretation in hard-rock environments | Need more evaluation |
| Borehole Geophysics | | | |
| VSP | Adequate | Adequate | Need evaluation |
| Electrical | Need equipment | Need to develop techniques | Need evaluation |
| Geotomography | Need equipment | Need to develop techniques | Need experience |

leum industry (Snyder and Fleming, 1985, this issue) is available to the high-temperature geothermal industry.

**Borehole geophysics**

Borehole geophysics has not undergone the development required even to assess its potential contribution to geothermal development. VSP techniques have emerged as being important in petroleum exploration, and development for these purposes will be important in geothermal application. Electrical borehole techniques have neither been developed nor seriously applied, although some numerical modeling capability exists to assess their contribution. Seismic geotomography is in the research and development stage, and its analog, electrical geotomography, has received virtually no effort. We believe that the borehole techniques are fertile ground for research and development.

## ACKNOWLEDGMENTS

We thank Charles M. Swift, Jr. and Joseph N. Moore for their timely review and helpful suggestions for improvement of the manuscript. This work was supported by the U.S. Department of Energy under contracts DE-AC07-80ID12079 and DE-AC03-84SF12196. Joan Pingree typed the many drafts of this manuscript, while Patrick Daubner coordinated the illustrations. We are grateful to both of them.

## REFERENCES

Ackermann, H. D., 1979, Seismic refraction study of the Raft River geothermal area, Idaho: Geophysics, **44**, 216–225.

Anderson, L. A., and Johnson, G. R., 1976, Application of the self-potential method to geothermal exploration in Long Valley, California: J. Geophys. Res., **81**, 1527–1532.

Applegate, J. K., Goebel, V. S., Kallenberger, P., and Rossow, J., 1981, The use of seismic reflection techniques in geothermal areas throughout the U.S.: Presented at 51st Ann. Internat. Mtg. and Expos., Soc. of Explor. Geophys., Los Angeles.

Arnorsson, S., Björnsson, A., Gislason, G., and Gudmundsson, G., 1976, Systematic exploration of the Krisuvik high-temperature area, Reykjanes Peninsula, Iceland: *in* Proc. Second U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 853–864.

Bacon, C. R., 1981, Geology and geophysics of the Cascade Range: Presented at 51st Ann. Internat. Mtg. and Expos., Soc. Explor. Geophys., Los Angeles.

Balch, A. H., Lee, M. W., Miller, J. J., and Ryder, R. T., 1982, The use of vertical seismic profiles in seismic investigation of the earth: Geophysics, **47**, 906–918.

Batchelor, A. S., 1982, The stimulation of a hot dry rock geothermal reservoir in the Cornubian Granite, England: *in* Proc. Eighth Workshop Geoth. Reservoir Eng., Stanford Univ., Stanford, California, SGP-TR-60.

Baudu, R., Bernhard, J., Georgel, J. M., Griveau, P., Rugo, R., 1980, Application of d.c. dipolar methods in the Upper Rhinegraben: *in* Advances in European and Geoth. Res., D. Reidel Co., 823–832.

Beasley, C. W., and Ward, S. H., 1986, Theoretical borehole-to-borehole and borehole-to-surface resistivity anomalies of geothermal fracture zones: Geophysics, **51**, January.

Berkman, F., and Lange, A. L., 1980, Tuscarora geophysics—preliminary report: Amax Exploration, Inc. internal rep., open filed by Univ. of Utah Res. Inst., Earth Sci. Lab.

Berktold, A., 1982, Electromagnetic studies in geothermal regions: Proc. 6th Workshop on Electromagnetic Induction in the Earth and Moon, Dept. Physics, Univ. of Victoria, Canada.

Beyer, J. H., 1977, Telluric and D.C. resistivity techniques applied to the geophysical investigation of Basin and Range geothermal systems: Univ. of California, Lawrence Berkeley Lab., Rep. LBL-6325.

Beyer, J. H., Morrison, H. F., and Dey, A., 1976a, Electrical exploration of geothermal systems in the Basin and Range valleys of Nevada: *in* Proc. Second U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 889–894.

Beyer, J. H., Dey, A., Liaw, A., Majer, E., McEvilly, T. V., Morrison, H. F., and Wollenberg, H., 1976b, Preliminary open file report, geological and geophysical studies in Grass Valley, Nevada: Univ. of California, Lawrence Berkeley Lab., Rep. LBL-5262.

Bhattacharyya, B. K., and Leu, L. K., 1975, Analysis of magnetic anomalies over Yellowstone National Park: mapping of Curie-point isothermal surface for geothermal reconnaissance: J. Geophys. Res., **80**, 4461–4465.

Biehler, S., 1971, Gravity studies in the Imperial Valley, *in* Cooperative geological-geophysical-geochemical investigations of geothermal resources in the Imperial Valley area of California: Univ. California, Riverside, Education Res. Service, 29–41.

Birch, F., 1950, Flow of heat in the Front Range, Colorado: Bull., Geol. Soc. Am., **61**, 567–630.

Bird, D. K., and Norton, D. L., 1981, Theoretical prediction of phase relations among aqueous solutions and minerals: Salton Sea Geothermal System: Geochim. Cosmochim. Acta, **45**, 1479–1493.

Björnsson, S., 1980, Natural heat saves millions of barrels of oil: Atlantica and Iceland Rev., **18**, 28–37.

Blackwell, D. D., and Morgan, P., 1976, Geological and geophysical exploration of the Marysville geothermal area, Montana, USA: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 895–902.

Blackwell, D. D., Steele, J. L., and Brott, C. A., 1980, The terrain effect on terrestrial heat flow: J. Geophys. Res., **85**, 4757–4772.

Blakeslee, S., 1984, Seismic discrimination of a geothermal field: Cerro Prieto: Lawrence Berkeley Lab., LBL-17859.

Bodvarsson, G. S., 1982, Mathematical modeling of the behavior of geothermal systems under exploitation: Ph.D. thesis, Univ. of California, Berkeley.

Bois, P., La Porte, M., Lavergne, M., and Thomas, G., 1972, Well-to-well seismic measurements: Geophysics, **37**, 471–480.

Brace, W. F., 1968, The mechanical effects of pore pressure on the fracturing of rocks: Geol. Survey Canada, paper 68-52.

Brott, C. A., Blackwell, D. D., and Morgan, P., 1981, Continuation of heat flow data: A method to construct isotherms in geothermal areas: Geophysics, **46**, 1732–1744.

Browne, P. R. L., 1978, Hydrothermal alteration in active geothermal fields: Annual Review Earth and Plan. Sci., **6**, 229–250.

Carson, C. C., and Allen, A. D., 1984, A program to investigate the engineering feasibility of extracting energy from shallow magma bodies: Trans., Geoth. Res. Coun., **8**, 3–5.

Cathles, L. M., 1977, An analysis of the cooling of intrusives by ground water convection which includes boiling: Econ. Geol., **72**, 804–826.

Chapman, D. S., and Pollack, H. N., 1977, Regional geotherms and lithospheric thickness: Geology, **5**, 265–268.

Clacy, G. R. T., 1968, Geothermal ground noise amplitude and frequency spectra in the New Zealand volcanic region: J. Geophys. Res., **73**, 5377–5383.

Combs, J., 1980, Heat flow in the Coso geothermal area, Inyo County, California: J. Geophys. Res., **85**, 2411–2424.

Combs, J., and Rotstein, Y., 1976, Microearthquake studies at the Coso geothermal area, China Lake, California: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 909–916.

Combs, J., and Hadly, D., 1977, Microearthquake investigation of the Mesa geothermal anomaly, Imperial Valley, California: Geophysics, **42**, 17–33.

Corwin, R. F., 1976, Self-potential exploration for geothermal reservoirs: *in* Proc. 2nd U.N. Sympos. on the Development and Use of Geothermal Res., San Francisco, **2**, 937–946.

Corwin, R. F., and Hoover, D. B., 1979, The self-potential method in geothermal exploration: Geophysics, **44**, 226–245.

Costain, J. K., Glover, L. III, and Sinha, A. K., 1980, Low temperature geothermal resources in the Eastern United States: EOS, **61**, 1–13.

Couch, R. W., Pitts, G. S., Gemperle, M., Braman, D. E., and Veen, C. A., 1982, Gravity anomalies in the Cascade Range in Oregon: Structural and thermal implications: Oregon Dept. Geol. Min. Ind. open-file rep. 0-82-9.

Craig, H., 1963, The isotopic geochemistry of water and carbon in geothermal areas, *in* Nuclear geology on geothermal areas, Spoleto: Pisa, Consiglio Nazionale delle Recerche, Laboratorie de Geologia Nucleare.

Crampin, S., 1978, Seismic wave propagation through a cracked solid: Polarization as a possible dilatancy diagnostic: Geophys. J., **53**, 467–496.

——— 1984, Anisotropy in exploration seismics: First Break, **2**(3), 19–21.

Daily, W., 1984, Underground oil-shale retort monitoring using geotomography: Geophysics, **49**, 1701–1707.

Daily, W. D., Lytle, R. J., Laine, E. F., Okada, J. T., and Deadrick, F. J., 1982, Geotomography in oil shale: J. Geophys. Res., **87**, B7, 5507–5515.

Daniels, J. J., 1983, Hole-to-surface resistivity measurements: Geophysics, **48**, 87–97.

Denlinger, R. P., and Kovach, R. L., 1981, Seismic-reflection investigations at Castle Rock Springs in The Geysers geothermal area: *in* McLaughlin, R. J., and Donnelly-Nolan, J. M., Eds., Research in The Geysers-Clear Lake Geothermal Area, Northern California: U.S.G.S. Prof. Paper 1141, 117–128.

Dey, A., and Morrison, H. F., 1976, Resistivity modeling for arbitrarily shaped two dimensional structures, Part I: Theoretical formulation: Lawrence Berkeley Lab. Rep. 5223.

——— 1977, An analysis of the bipole-dipole method of resistivity surveying: Geothermics, **6**, 47–81.

Dickinson, D. J., 1976, An airborne infrared survey of the Tauhara Geothermal Field, New Zealand: *in* Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 955–961.

Dobecki, T. L., 1980, Borehole resistivity curves near spheroidal masses: Geophysics, **45**, 1513–1521.

Douze, E. G., and Sorrells, G. G., 1972, Geothermal ground noise surveys: Geophysics, **37**, 813–824.

Duffield, W. A., and Bacon, C. R., 1977, Preliminary geologic map of the Coso volcanic field and adjacent areas, Inyo County, California, with a table of new K/Ar dates by G. B. Dalrymple: U.S.G.S., open-file rep. 77-311, scale 1:50,000.

Duffield, W. A., Bacon, C. R., and Dalrymple, G. B., 1980, Late Cenozoic volcanism, geochronology, and structure of the Coso Range, Inyo County, California: J. Geophys. Res., **85**, 2381–2404.

Dupis, A., Marie, P., and Petian, G., 1980, Magnetotelluric prospecting of the Mont Dore area, *in* Advances in European Geoth. Res.: D. Reidel, 935–943.

Dyck, A. M., and Young, R. P., 1985, Physical characterization of rock masses using borehole methods: Geophysics, **50**, 2530–2541.

Edquist, R. K., 1981, Geophysical investigations of the Baltazor Hot Springs Known Geothermal Resource Area and the Painted Hills thermal area, Humboldt County, Nevada: Univ. of Utah Res. Inst., Earth Science Lab. Rep. DOE/ID/12079-29.

Elders, W. A., Hoagland, J. R., McDowell, S. D., and Cobo, R. J. M., 1978, Hydrothermal mineral zones in the geothermal reservoir of Cerro Prieto: *in* Proc., First Sympos. on the Cerro Prieto geothermal field, Baja, California, Mexico, September, 1978: Lawrence Berkeley Lab., LBL-**7098**, 68–75.

Ellis, A. J., and Mahon, W. A. J., 1977, Chemistry and geothermal systems: Academic Press Inc.

Fehler, M., Turpening, R., Blackway, C., and Mellen, M., 1982, Detection of a hydrofrac with shear wave vertical seismic profiles: Presented at 52nd Ann. Internat. Mtg. and Expos., Soc. Expl. Geophys., Dallas.

Fischer, W. A., Davis, B. A., and Souza, T., 1966, Fresh water springs of Hawaii from infrared images: U.S. Geol. Surv. Hydrologic Atlas, HA-218.

Flovenz, O. G., and Georgeson, L. S., 1982, Prospecting for near vertical aquifers in low-temperature geothermal areas in Iceland: Geoth. Res. Counc., Trans., **6**, 19–22.

Fox, R. C., 1978a, Dipole-dipole resistivity survey of a portion of the Coso Hot Springs KGRA, Inyo County, California: Univ. of Utah Res. Inst., Earth Sci. Lab., Rep. ID0/77.5.6.

——— 1978b, Low-altitude aeromagnetic survey of a portion of the Coso Hot Springs KGRA, Inyo County, California: Univ. of Utah Res. Inst., Earth Sci. Lab., Rep. ID0/77.5.7.

Fox, R. C., Ross, H. P., and Wright, P. M., 1978, Dipole-dipole resistivity survey of a portion of the Coso Hot Springs KGRA, Inyo Country, California: Presented at the 48th Ann. Internat. Mtg. and Expos., Soc. Explor. Geophys., San Francisco.

Frangos, W., and Ward, S. H., 1980, Bipole-dipole survey at Roosevelt Hot Springs thermal area, Beaver County, Utah: Univ. of Utah Res. Inst., Earth Sci. Lab., Rep. DOE/ID/12079-15.

Furumoto, A. S., 1976, A coordinated exploration program for geothermal sources on the island of Hawaii: *in* Proc., Second United Nations Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 993–1003.

Galbraith, R. M., 1978, Geological and geophysical analysis of Coso Geothermal Exploration Hole No 1 (CGE-1), Coso Hot Springs KGRA, California: Univ. of Utah Res. Inst., Earth Sci. Lab. Rep. DOE/ID/78-1701.b.4.2.

Gal'perin, E. I., 1973, Vertical seismic profiling: Soc. Explor. Geophys.

Gamble, T. D., Goubau, W. M., Goldstein, N. E., and Clarke, J., 1980, Referenced magnetotellurics at Cerro Prieto: Geothermics, **9**, 49–63.

Gertson, R. C., and Smith, R. B., 1979, Interpretation of a seismic refraction profile across the Roosevelt Hot Springs, Utah and vicinity: Univ. of Utah, Dept. Geol. and Geophys., Rep. ID0/78-1701.a.3.

Glenn, W. E., and Hulen, J. B., 1979, A study of well logs from Roosevelt Hot Springs KGRA, Utah: *in* Soc. Prof. Well Log Analysts. 20th Ann. Logging Sympos. Trans., paper II.

Glenn, W. E., and Ross, H. P., 1982, A study of well logs from Cove Fort-Sulphurdale KGRA, Utah: Univ. of Utah, Res. Inst., Earth Sci. Lab., Rep. 75.

Goldstein, N. E., and Mozley, E., 1978, A telluric-magnetotelluric survey at Mount Hood, Oregon: Univ. California, Lawrence Berkeley Lab. Rep. LBL-7050.

Goldstein, N. E., Norris, R. A., and Wilt, M. J., 1978, Assessment of surface geophysical methods in geothermal exploration and recommendations for future research: Univ. California, Lawrence Berkeley Laboratory, Rep. LBL-6815.

Goldstein, N. E., and Paulsson, B., 1976, Interpretation of gravity surveys in Grass and Buena Vista Valleys, Nevada: Geothermics, **7**, 29–50.

Goldstein, N. E., Mozley, E., and Wilt, M., 1982, Interpretation of shallow electrical features from electromagnetic and magnetotelluric surveys at Mount Hood, Oregon: J. Geophys. Res., **87**, 2815–2828.

Grannell, R. B., 1980, The use of surface gravity methods in monitoring subsurface reservoir changes, with case studies at Cerro Prieto, Mexico, and Heber, California: Trans., Geoth. Res. Counc., **4**, 49–52.

Grasty, R. L., Glynn, J. E., and Grant, J. A., 1985, The analysis of multi-channel airborne gamma-ray spectra: Geophysics, **50**, this issue, 2611–2620.

Grindly, G. W., and Browne, P. R. L., 1976, Structural and hydrological factors controlling the permeabilities of some hot-water geothermal fields: *in* Proc. Second United Nations Sympos. on the Development and Use of Geoth. Res., San Francisco, **1**, 377–386.

Gupta, H. K., Ward, R. W., and Lin, T-L., 1982, Seismic wave velocity investigation at The Geysers-Clear Lake geothermal field, California: Geophysics, **47**, 819–824.

Halfman, S. E., Lippmann, M. J., and Zelwer, R., 1982, The movement of geothermal fluid in the Cerro Prieto field as determined from well log and reservoir engineering data, *in* Proc. 8th Workshop Geoth. Reservoir Eng., Stanford Univ., Stanford, California, SGP-TR-60.

Halfman, S. E., Lippmann, M. J., Zelwer, R., and Howard, J. H., 1984, Geologic interpretation of geothermal fluid movement in Cerro Prieto Field, Baja, California, Mexico: Bull., Am. Assn. Petr. Geol., **68**, 18–30.

Hamilton, R. M., and Muffler, L. J. P., 1972, Microearthquakes at The Geysers geothermal area, California: J. Geophys. Res., **77**, 2081–2086.

Hatherton, T., MacDonald, W. J. P., and Thomson, G. E. K., 1966, Geophysical methods in geothermal prospecting in New Zealand: Bull. Volcan., 485–497.

Helgeson, H. C., 1968, Geologic and thermodynamic characteristics of the Salton Sea geothermal system: Am. J. Sci., **266**, 129–166.

Henley, R. W., and Ellis, A. J., 1983, Geothermal systems ancient and modern, a geochemical review: Earth Sci. Rev., **19**, 1–50.

Hermance, J. F., Thayer, R. E., Björnsson, A., 1976, The telluric-magnetotelluric method in the regional assessment of geothermal potential, *in* Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1037–1048.

Hill, D. P., 1976, Structures of Long Valley Caldera, California, from a seismic refraction experiment: J. Geophys. Res., **81**, 745–753.

Hill, D. G., Layman, E. G., Swift, C. M., Jr., and Yungul, S. H., 1979, Soda Lake, Nevada, thermal anomaly: Trans., Geoth. Res. Counc., **3**, 305–308.

Hill, D. P., Mooney, W. D., Fuis, G. S., and Healy, J. H., 1981, Evidence on the structure and tectonic environment of the volcanoes in the Cascade Range, Oregon and Washington, from seismic refraction/reflection measurements: Presented at the 51st Ann. Internat. Mtg. and Expos., Soc. Explor. Geophys., Los Angeles.

Hochstein, M. P., and Hunt, T. M., 1970, Seismic, gravity, and magnetic studies, Broadlands geothermal field, New Zealand, *in* Proc. U.N. Sympos. on the Development and Utilization of Geoth. Res., Pisa, Geothermics Spec. Issue 2, **2**, 333–346.

Hohmann, G. W., and Ward, S. H., 1981, Electrical methods in mining geophysics: Econ. Geol., 75th Anniversary Volume, 806–828.

Hoover, D. B., and Long, C. L., 1976, Audiomagnetotelluric methods in reconnaissance geothermal exploration: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1059–1064.

Hoover, D. B., Long, C. L., and Senterfit, R. M., 1978, Some results from audiomagnetotelluric investigations in geothermal areas: Geophysics, **43**, 1501–1514.

Hulen, J. B., 1978, Geology and alteration of the Coso geothermal area, Inyo County, California: Univ. of Utah Res. Inst., Earth Sci. Lab. Rep. DOE/ID/28392-4.

Hunt, T. M., and Lattan, J. H., 1982, A survey of seismic activity near Wairakei geothermal field, New Zealand: J. Volcan. and Geoth. Res., **14**, 319–334.

Isherwood, W. F., 1976, Complete Bouguer gravity map of The Geysers area, California: U.S. Geol. Surv. open-file rep., 76-357.

Isherwood, W. F., and Mabey, D. R., 1978, Evaluation of Baltazor known geothermal resource area, Nevada: Geothermics, 7, 221–229.

Iyer, H. M., and Hitchcock, T., 1976, Seismic noise as a geothermal exploration tool: techniques and results, *in* Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1075–1083.

Iyer, H. M., and Stewart, R. M., 1977, Teleseismic technique to locate magma in the crust and upper mantle, *in* Dick, H. J. B., Ed., Magma genesis: Oregon Dept of Geol. and Min. Ind., Bull. **96**, 281–299.

Iyer, H. M., Oppenheimer, D. H., and Hitchcock, T., 1979, Abnormal *P*-wave delays in The Geysers-Clear Lake geothermal area, California: Science, **204**, 495.

Jackson, D. B., O'Donnell, J. E., and Gregory, D. I., 1977, Schlumberger soundings, audio-magnetotelluric soundings and telluric mapping in and around the Coso Range, California: U.S. Geol. Surv. open-file rep. 77-120.

Jackson, D. B., and O'Donnell, J. E., 1980, Reconnaissance electrical surveys in the Coso Range, California: J. Geophys. Res., **85**, 2502–2516.

Jiracek, G. R., and Smith, C., 1976, Deep resistivity investigations at two known geothermal resource areas (KGRAs) in New Mexico: Radium Springs and Lightning Dock: New Mexico Geol. Soc. Spec. Pub., **6**, 71–76.

Kane, M. F., Mabey, D. R., and Brace, R., 1976, A gravity and magnetic investigation of the Long Valley Caldera, Mono County, California: J. Geophys. Res., **81**, 754–762.

Kappelmeyer, O., and Haenel, R., 1974, Geothermics with special reference to application: Geoexploration Monograph Ser. 1, **4**, Gebruder-Borntraeger.

Kauahikaua, J., 1981, Interpretation of time-domain electromagnetic soundings in the East Rift geothermal area of Kilauea volcano, Hawaii: U.S. Geol. Surv. open-file rep. 81–979.

Keller, G. V., 1970, Induction methods in prospecting for hot water, *in* Proc. U.N. Sympos. on the Development and Utilization of Geoth. Res., Pisa, Geothermics Spec. Issue 2, **2**, 318–332.

Keller, G. V., and Rapolla, A., 1974, Electrical prospecting methods in volcanic areas: *in* Civetta, L., Gasparini, P., Luongo, G., and Rapolla, A., Eds., Physical volcanology: Elsevier Science Publishers.

Keller, G. V., Furgerson, R. Lee, C. Y., Harthill, N., and Jacobson, J. J., 1975, The dipole mapping method: Geophysics, **40**, 451–472.

Keller, G. V., Skokan, C. V., Skokan, J. J., Daniels, J., Kauahikaua, J. P., Klein, D. P., and Zablocki, C. J., 1977, Geo-electric studies on the east rift, Kilauea Volcano, Hawaii Island: Hawaii Inst. Geophys. Rep. HIG-77-15.

Keller, G. V., Taylor, K., and Santo, J. M., 1981, Megasource EM method for detecting deeply buried conductive zones in geothermal exploration: Presented at the 51st Ann. Internat. Mtg. and Expos., Soc. Explor. Geophys., Los Angeles.

Keys, W. S., and Sullivan, J. K., 1979, Role of borehole geophysics in defining the physical characteristics of the Raft River geothermal reservoir, Idaho: Geophysics, **44**, 1116–1141.

Killpack, T. J., and Hohmann, G. W., 1979, Interactive dipole-dipole resistivity and IP modeling of arbitrary two-dimensional structures (IP2D user's guide and documentation): Univ. Utah Res. Inst., Earth Sci. Lab. Rep., 15.

Kintzinger, P. R., 1956, Geothermal survey of hot ground near Lordsburg, New Mexico: Science, **124**, 629–630.

Koenig, J. B., Gawarecki, S. J., and Austin C. G., 1972, Remote sensing survey of the Coso geothermal area, Inyo County, California: Naval Weapons Center Tech. Publ. 5233, China Lake, CA.

Lachenbruch, A. H., 1978, Heat flow in the Basin and Range Province and thermal effects of tectonic extension: Pure and Appl. Geophys., **117**, 34–50.

Lachenbruch, A. H., Sorey, M. L., Lewis, R. E., and Sass, J. H., 1976, The near-surface hydrothermal regime of Long Valley caldera: J. Geophys. Res., **81**, 763–768.

Lager, D. L., and Lytle, R. J., 1977, Determining a subsurface electromagnetic profile from high-frequency measurements by applying reconstruction technique algorithms: Radio Sci., **12**, 249–260.

Lange, A. L., and Westphal, W. H., 1969, Microearthquakes near The Geysers, Sonoma County, California: J. Geophys. Res., **74**, 4377–4382.

Laster, S. J., 1985, The present state of seismic acquisition: one view: Geophysics, **50**, this issue, 2443–2451.

Lawrence Berkeley Laboratory, 1984, Proposed scientific activities for the Salton Sea Scientific Drilling Project: A report of the experiments panel: Lawrence Berkeley Lab., Univ. of California, Rep. LBL-17716.

Lebedev, E. B., and Khitarov, N. I., 1964, Dependence on the beginning of melting of granite and the electrical conductivity of its melt on high water vapor pressure: Geokhimija, **3**, 195–201.

Lee, T.-C., and Cohen, L. H., 1979, Onshore and offshore measurements of temperature gradients in the Salton Sea Geothermal Area, California: Geophysics, **44**, 206–215.

LeSchack, L. A., Lewis, J. E., and Chang, D. C., 1977, Rapid reconnaissance of geothermal prospects using shallow temperature surveys: LeSchack Associates, Ltd. Tech. Rep. to Dept. of Energy.

LeSchack, L. A., and Lewis, J. E., 1983, Geothermal prospecting with Shallo-Temp surveys: Geophysics, **48**, 975–996.

Lezama, G. F., 1984, On Schlumberger soundings and head-on measurements: U.N. Univ. Geothermal Training Programme, Iceland, Rep. 1984-9.

Liaw, A. L., and McEvilly, T. V., 1979, Microseisms in geothermal exploration—Studies in Grass Valley, Nevada: Geophysics, **44**, 1097–1115.

Liaw, A., and Suyenaga, W., 1982, Detection of geothermal microtremors using seismic arrays: Presented at 52nd Ann. Internat. Mtg. and Expos., Soc. of Explor. Geophys., Dallas.

Lovering, T. S., and Goode, H. D., 1963, Measuring geothermal gradients in drill holes less than 60 feet deep, East Tintic District, Utah: U.S. Geol. Surv. Bull. 1172.

Lumb, F. T., and MacDonald, W. J. P., 1970, Near-surface resistivity surveys of geothermal areas using the electromagnetic method, *in* U.N. Sympos. on Development and Utilization of Geoth. Res., Pisa, Geothermics Spec. Issue **2**, 311–317.

Mabey, D. R., Hoover, D. B., O'Donnell, J. E., and Wilson, C. W., 1978, Reconnaissance geophysical studies of the geothermal system in southern Raft River Valley, Idaho: Geophysics, **43**, 1470–1484.

MacDonald, W. J. P., and Muffler, L. J. P., 1972, Recent geophysical exploration of the Kawerau geothermal field, North Island, New Zealand: New Zealand J. Geol. and Geophys., **18**.

Mahon, W. A. J., Klyen, L. E., and Rhode, M., 1980, Neutral sodium bicarbonate/sulfate hot waters in geothermal systems: Chinetsu (J. of the Japan Geothermal Energy Assn.), **17**, 11–24.

Majer, E. L., 1978, Seismological investigations in geothermal regions: Univ. of California, Lawrence Berkeley Lab., Rep. LBL-7054.

Majer, E. L., and McEvilly, T. V., 1979, Seismological investigations at The Geysers geothermal field: Geophysics, **44**, 246–249.

Martinez, M., Fabrial, H., and Romo, J. M., 1982, Magnetotelluric studies in the geothermal area of Culiacan, Mexico: Sixth Workshop on Electromagnetic Induction in the Earth and Moon, Internat. Ass'n. Geomag. and Aeron., Victoria, British Columbia, Dept. of Physics, Univ. of Victoria (abstr).

McNitt, J. R., 1976, Summary of United Nations geothermal exploration experience, 1965 to 1975: *in* Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1127–1134.

Meidav, T., and Tonani, F., 1976, A critique of geothermal exploration techniques: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1143–1154.

Moore, J. N., Adams, M. C., and Stauder, J. J., 1985, Geologic and geochemical investigations of the Meager Creek geothermal system, British Columbia, Canada: Proc. Tenth Workshop on Geoth. Res. Eng., Stanford Univ., Stanford, CA.

Morrison, H. F., Goldstein, N. E., Hoversten, M., Oppliger, G., and Riveros, C., 1978, Description, field test, and data analysis of a controlled-source EM system (EM-60): Univ. of California, Lawrence Berkeley Laboratory, Rep. LBL-7088.

Moskowitz, B., and Norton, D., 1977, A preliminary analysis of intrinsic fluid and rock resistivity in active hydrothermal systems: J. Geophys. Res., **82**, 5787–5795.

Muffler, L. J. P., and White, D. E., 1969, Active metamorphism of Upper Cenozoic sediments in the Salton Sea geothermal field and the Salton Trough, southeastern California: Bull., Geol. Soc. Am., **80**, 157–182.

Musmann, G., Gramkow, B., Lohr, V., and Kertz, W., 1980, Magnetotelluric survey of the Lake Laach (Eifel) volcanic area: Advances in European Geoth. Res.: D. Reidel Co., 904–910.

Nabighian, M. N., Oppliger, G. L., Edwards, R. N., Lo, B. B. H., and Chessman, S. J., 1984, Cross-hole magnetometric resistivity (MMR): Geophysics, **49**, 1313–1326.

Newman, G. A., Wannamaker, P. E., and Hohmann, G. W., 1985, On the detectability of crustal magma chambers using the magnetotelluric method: Geophysics, **50**, 1136–1143.

Nielson, D. L., 1978, Radon emanometry as a geothermal exploration technique: theory and an example from Roosevelt Hot Springs KGRA Utah: Univ. Utah Res. Inst., Earth Sci. Lab Rep. No. 14.

Noble, J. W., and Ojiambo, S. B., 1976, Geothermal exploration in Kenya: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., **1**, 189–204.

Norton, D. L., 1984, Theory of hydrothermal systems: Ann. Rev. Earth Plan., Sci., **12**, 155–177.

Nur, A., and Simmons, G., 1969, The effect of saturation on velocity in low porosity rocks: Earth Plan. Sci. Lett., **7**, 183–193.

Okubo, Y., Graf, R. J., Hansen, R. O., Ogawa, K., and Tsu, H., 1985, Curie point depths of the Island of Kyushu and surrounding areas, Japan: Geophysics, **50**, 481–494.

Olmstead, F. H., 1977, Use of temperature surveys at a depth of 1 meter in geothermal exploration in Nevada: U.S. Geol. Surv. Prof. Paper, 1044-B.

Oppenheimer, D. H., and Iyer, H. M., 1980, Frequency-wavenumber analysis of geothermal microseisms at Norris Geyser basin, Yellowstone National Park, Wyoming: Geophysics, **45**, 952–963.

Oristaglio, M. L., 1985, A guide to the current uses of vertical seismic profiles: Geophysics, **50**, this issue, 2473–2479.

Paillet, F. L., 1980, Acoustic propagation in the vicinity of fractures which intersect a fluid-filled borehole: Presented at the 21st Ann. Logging Sympos., Soc. Prof. Well Log Analysts, Lafayette, LA, DD1-DD33.

Pálmason, G., 1976, Geophysical methods in geothermal exploration: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, 1975, **2**, 1175–1184.

Patella, D., Quarto, R., and Tramacere, A., 1980, Dipole-dipole study of the Travale geothermal field: Advances in European Geoth. Res., D. Reidel Co., 833–842.

Paterson, N. R., and Reeves, C. V., 1985, Gravity and magnetic surveys: the state-of-the-art in 1985: Geophysics, **50**, this issue, 2558–2594.

Poley, J. P., and Van Steveninck, J., 1970, Geothermal prospecting-delineation of shallow salt domes and surface faults by temperature measurements at a depth of approximately 2 meters: Geophys. Prosp., **18**, 666–700.

Rai, C. S., and Manghnani, M. H., 1978, Electrical conductivity of basalts to 1550 C, *in* Dick, H. J. B., Ed., Proc. of Chapman Conference on Partial Melting in the Earth's Upper Mantle, Oregon, Dept Geol and Min. Ind., Bull., **96**, 219–232.

Rapolla, A., and Keller, G. V., Eds., 1984, Geophysics of geothermal areas: State of the art and future development: Proc. of the 3rd Course held at the School of Geophysics, "Ettore Majorana" Internat. Centre for Sci. Culture, Erice, Italy, 1980; Colorado School of Mines Press.

Razo, A., Arellano, F., and Fouseca, H., 1980, CFE resistivity studies at Cerro Prieto: Geothermics, **9**, 7–14.

Reasenberg, P., Ellsworth, W., and Walter, A., 1980, Teleseismic evidence for a low-velocity body under the Coso geothermal area: J. Geophys. Res., **85**, 2471–2483.

Rijo, L., 1977, Modeling of electric and electromagnetic data: Ph.D. thesis, Univ. of Utah.

Risk, G. F., MacDonald, W. J. P., and Dawson, G. B., 1970, D.C. resistivity surveys of the Broadlands geothermal region, New Zealand: Proc. U.N. Sympos. on the Development and Utilization of Geoth. Res., Pisa, Geothermics, Spec. Issue 2, **2**, 287–294.

Robinson, R., and Iyer, H. M., 1981, Delineation of a low-velocity body under the Roosevelt Hot Springs geothermal area, Utah, using teleseismic *P*-wave data: Geophysics, **46**, 1456–1466.

Ross, H. P., Nielson, D. L., and Moore, J. N., 1982, Roosevelt Hot Springs geothermal system, Utah—Case study: Bull., Am. Assn. Petr. Geol., **66**, 879–902.

Ross, H. P., and Moore, J. N., 1985, Geophysical investigations of the Cove Fort-Sulphurdale geothermal system, Utah: Geophysics, **50**, November.

Rundle, J. B., Elbring, G. J., Striker, R. P., Finger, J. T., Carson, C. C., Walck, M. C., Ellsworth, W. L., Hill, D. P., Malin, P., Tono, E., Robertson, M., Kuhlman, S., McEvilly, T., Clymer, R., Smithson, S. B., Deemer, S., Johnson, R., Henyey, T., Hauksson, E., Leary, P., McCraney, J., and Kissling, E., 1985, Seismic imaging in Long Valley, California, by surface and borehole techniques: An investigation of active tectonics: EOS, **66**, 194–201.

Sandberg, S. K., and Hohmann, G. W., 1982, Controlled-source audiomagnetotellurics in geothermal exploration: Geophysics, **47**, 100–116.

Sanyal, S. K., Wells, L. E., and Bickham, R. E., 1980, Geothermal well log interpretation state of the art—Final report: Los Alamos Scientific Lab. Rep. LA-8211-MS.

Sass, J. H., Lachenbruch, A. H., Munroe, R. J., Greene, G. W., and Moses, T. H., Jr., 1971, Heat flow in the western United States: J. Geophys. Res., **76**, 6367–6413.

Sass, J. H., Blackwell, D. D., Chapman, D. S., Costain, J. K., Decker, E. R., Lawver, L. A., and Swanberg, C. A., 1981, Heat flow from the crust of the United States: *in* Tourlovkian, Y. S., Judd, W. R., Roy, R. F., Eds., Physical properties of rocks and minerals: McGraw-Hill Book Co., 503–548.

Schultz, P. S., 1985, Seismic data processing: current industry practice and new directions: Geophysics, **50**, this issue, 2452–2457.

Shankland, T. J., and Waff, H. S., 1977, Partial melting and electrical conductivity anomalies in the upper mantle: J. Geophys. Res., **82**, 5409–5417.

Shuey, R. T., Schellinger, D. K., Tripp, A. C., 1977, Curie depth determination from aeromagnetic data: Geophys. J. Roy. Astr. Soc.), **50**, 75–101.

Sill, W. R., 1983, Self-potential modeling from primary flows: Geophysics, **48**, 76–86.

Smith, C., 1983, Thermal hydrology and heat flow of Beowawe geothermal area, Nevada: Geophysics, **48**, 681–626.

Smith, L., and Chapman, D. S., 1983, On the thermal effects of groundwater flow, 1, Regional scale systems: J. Geophys. Res., **88**, 593–608.

Smith, M. C., and Ponder, G. M., 1982, Hot dry rock geothermal energy development program annual report fiscal year 1981: Los Alamos Nat. Lab., NM, (LA-9287-HDR).

Smith, R. J., 1985, Geophysics in Australian mineral exploration: Geophysics, **50**, this issue, 2637–2665.

Snyder, D. D., and Fleming, D. B., 1985, Well logging—A 25-year perspective: Geophysics, **50**, this issue, 2504–2529.

Souto, J. M., 1978, Oahu geothermal exploration: Trans. Geoth. Res. Counc., **2**, 605–607.

Stanley, W. D., 1982, Magnetotelluric soundings on the Idaho National Engineering Laboratory facility, Idaho: J. Geophys. Res., **87**, 2683–2691.

Stanley, W. D., Jackson, D. B., and Zohdy, A. A. R., 1976, Deep electrical investigations in the Long Valley geothermal area, California: J. Geophys. Res., **81**, 810–820.

Stanley, W. D., Boehl, J. E., Bostick, F. X., Jr., and Smith, H. W., 1977, Geothermal significance of magnetotelluric sounding in the eastern Snake River Plain-Yellowstone region: J. Geophys. Res., **82**, 2501–2514.

Steeples, D. W., and Iyer, H. M., 1976a, Low-velocity zone under Long Valley as determined from teleseismic events: J. Geophys. Res., **81**, 849–860.

——— 1976b, Teleseismic *P*-wave delays in geothermal exploration: *in* Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1199–1206.

Stolt, R. H., and Weglein, A. B., 1985, Migration and inversion of seismic data: Geophysics, **50**, 2458–2472.

Studt, F. E., 1964, Geophysical prospecting in New Zealand's hydrothermal fields: *in* Proc., United Nations Conf. on New Sources of Energy, **2**, 380.

Swanberg, C. A., 1976, The Mesa geothermal anomaly, Imperial Valley, California: A comparison and evaluation of results obtained from surface geophysics and deep drilling: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1217–1229.

Swift, C. M., Jr., 1979, Geophysical data, Beowawe geothermal area, Nevada: Trans., Geoth. Res. Counc., **3**, 701–703.

Tripp, A. C., Ward, S. H., Sill, W. R., Swift, C. M., Jr., and Petrick, W. R., 1978, Electromagnetic and Schlumberger resistivity sounding in the Roosevelt Hot Springs KGRA, Geophysics, **43**, 1450–1469.

Varet, J., 1982, Géothermie basse-énergie: Usage direct de la chaleur: Masson.

Wallace, R. H., Jr., Kraemer, T. F., Taylor, R. E., and Wesselman, J. B., 1979, Assessment of geopressured-geothermal resources in the Northern Gulf of Mexico Basin: *in* Muffler, L. J. P., Ed., Assessment of Geoth. Res. of the U.S., Geol. Surv. Circular 790.

Walter, A. W., and Weaver, C. S., 1980, Seismic studies in the Coso geothermal area, Inyo County, California: J. Geophys. Res., **85**, 2441–2458.

Wannamaker, P. E., Ward, S. H., Hohmann, G. W., and Sill, W. R., 1980, Magnetotelluric models of the Roosevelt Hot Springs thermal area, Utah: Univ. of Utah Res. Inst., Earth Science Lab., Rep. DOE ET 27002-8.

Ward, P. L., and Björnsson, S., 1971, Microearthquake swarms and the geothermal areas of Iceland: J. Geophys. Res., **76**, 3953–3982.

Ward, R. W., Butler, D., Iyer, H. M., Laster, S., Lattanner, A., Majer, E., and Mass, J., 1979, Seismic Methods, *in* Nielson, D. L., Ed., Program Review Geoth. Explor. and Assessment Tech. Program, Rep. DOE ET 27002-6, Univ. of Utah Res. Inst., Earth Sci. Lab.

Ward, S. H., 1983a, Controlled source electromagnetic methods in geothermal exploration: U.N. Univ. Geothermal Training Programme, Iceland, Rep. 1983-4.

——— 1983b, Geophysical studies of active geothermal systems in the northern Basin and Range: Geoth. Res. Counc., Spec. Rep. No. 13, 121–157.

Ward, S. H., Parry, W. T., Nash, W. P., Sill, W. R., Cook, K. L., Smith, R. B., Chapman, D. S., Brown, F. H., Whelan, J. A., and Bowman, J. R., 1978, A summary of the geology, geochemistry, and geophysics of the Roosevelt Hot Springs thermal area, Utah: Geophysics, **43**, 1515–1542.

Ward, S. H., and Sill, W. R., 1983, Resistivity, induced polarization, and self-potential methods in geothermal exploration: Univ. of Utah Res. Inst., Earth Sci. Lab., Rep. DOE ID 12079-90.

Ward, S. H., and Wannamaker, P. E., 1983, The MT/AMT electromagnetic method in geothermal exploration: U.N. Univ. Geothermal Training Programme, Iceland, Rep. 1983-5.

Wechsler, D. J., and Smith, R. B., 1979, An evaluation of hypocenter location techniques with applications to Southern Utah: Regional earthquake distributions and seismicity of geothermal areas: Univ. of Utah, Dept. of Geol. and Geophys., Rep. IDO DOE ET 28392-

32.
West, G. F., and Edwards, R. N., 1985, A simple parametric model for the electromagnetic response of an anomalous body in a host medium: Geophysics, **50**, this issue, 2542–2557.
White, D. E., 1965, Geothermal energy: U.S. Geol. Surv. Circ. 519.
——— 1969, Rapid heat-flow surveying of geothermal areas, utilizing individual snowfalls as colorimeters: J. Geophys. Res., **74**, 5191–5201.
White, D. E., Muffler, L. J. P., and Truesdell, A. H., 1971, Vapor-dominated hydrothermal systems compared with hot water systems: Econ. Geol., **66**, 75–97.
White, D. E., and Williams, D. L., Eds., 1975, Assessment of geothermal resources of the United States-1975: Geol. Surv. Circ. 726.
Williams, D. L., and Von Herzen, R. P., 1974, Heat loss from the Earth: New estimate: Geology, **2**, 327–328.
Williams, D. L., and Finn, C., 1982, Evidence from gravity data on the location and size of subvolcanic intrusions: Preliminary results (abstr.): Geophysics, **47**, 425.
Williams, P. D., Mabey, D. R., Zohdy, A. R., Ackermann, H., Hoover, D. B., Pierce, K. L., Oriel, S. S., 1976, Geology and geophysics of the southern Raft River valley geothermal area, Idaho, USA: Proc. 2nd U.N. Sympos. on Development and Use of Geoth. Res., San Francisco, **2**, 1273–1282.
Wilt, M. J., Beyer, J. H., and Goldstein, N. E., 1980, A comparison of dipole-dipole resistivity and electromagnetic induction sounding over the Panther Canyon thermal anomaly, Grass Valley, Nevada: Trans., Geoth. Res. Council, **4**, 101–104.
Wilt, M. J., and Goldstein, N. E., 1981, Resistivity monitoring at Cerro Prieto: Geothermics, **10**, 183–193.
Wilt, M. J., Stark, M., Goldstein, N. E., and Haught, J. R., 1981, Electromagnetic induction sounding at geothermal prospects in Nevada: Univ. of California, Lawrence Berkeley Lab., Rep LBL-12100.
Wodzicki, A., and Weissburg, B. G., 1970, Structural control of base metal mineralization at the Tui Mine, Te Aroha, New Zealand: New Zealand J. Geol. and Geophys., **13**, 610.
Wollenburg, H. A., 1976, Radioactivity of geothermal systems: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1282–1292.
Wright, P. M., 1981, Gravity and magnetic methods in mineral exploration: Econ. Geol., 75th Anniversary Volume, 829–839.
Yang, F. W., and Ward, S. H., 1985, Single- and cross-borehole resistivity anomalies of thin ellipsoids and spheroids: Geophysics, **50**, 637–655.
Young, C.-Y., and Ward, R. W., 1980, 3-D $Q^{-1}$ model of Coso Hot Springs, KGRA: J. Geophys. Res., **85**, 2459–2470.
Zablocki, C. J., 1976, Mapping thermal anomalies on an active volcano by the self-potential method, Kilauea, Hawaii: Proc. 2nd U.N. Sympos. on the Development and Use of Geoth. Res., San Francisco, **2**, 1299–1309.
Zandt, G., McPherson, L., Schaff, S., and Olsen, S., 1982, Seismic baseline and induction studies, Roosevelt Hot Springs, Utah and Raft River, Idaho, Final Technical Report: Univ. Utah Research Inst., Earth Sci. Lab., Rep. 100.
Zoback, M. L., and Anderson, R. E., 1983, Style of Basin and Range faulting as inferred from seismic reflection data in the Great Basin, Nevada and Utah: Geoth. Res. Counc., Spec. Rep. No. 13.
Zohdy, A. A. R., Anderson, L. A., and Muffler, L. J. P., 1973, Resistivity, self-potential, and induced polarization surveys of a vapor-dominated geothermal system: Geophysics, **38**, 1130–1144.