STATE OF THE ART

GEOPHYSICAL EXPLORATION FOR GEOTHERMAL RESOURCES

by

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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 geothermal development. Currently, only the moderate- and high-temperature hydrothermal convective type of geothermal system can be economically used for electric power generation. 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 in order 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 may seem small compared to the estimated 8.4 x 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. It is very difficult to estimate the ultimate potential contribution of geothermal energy to mankind's needs for at least three reasons: 1) 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 the total heat stored above surface temperature in the earth to a depth of 10 km to be about 1.3 x 10^{27} J, equivalent to the burning of about 2.3 x 10^{17} barrels of oil. It is apparent that if even a small part of this heat could be made available, its contribution would be significant.

In the United States, commercial development of geothermal energy is pursued by private industry, and much 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 there is reasonably good access to data, although not all of it appears 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 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 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, while it is small compared to the corresponding petroleum or minerals infrastructures, it is making important contributions.

In this paper we seek to review the application of geophysical methods to geothermal exploration and development and to assess the current state of the art. Previous reviews of geophysical applications have been given by Pálmason (1976), McNitt (1976), Meidav and Tonani (1976), Ward (1983b), and Rapolla and Keller (1984), among others. There are many more published accounts of geophysical work in the geothermal environment than can be discussed or referenced in an article of this length. We have cited a few typical references for each of the applications discussed. We apologize in advance for emphasis on the U. S. literature, with which we are most familiar, and for omissions in recognizing the 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 those characterized by natural thermal waters, and are divided into those with significant large-scale convection and those without. 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 the hydrothermal resources have been developed to any extent, the other resources being presently uneconomic.

Classification of Geothermal Systems

Convective hydrothermal resources are geothermal resources in which the earth's heat is actively carried upward by the 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 whose temperature may be 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 *regional aquifers*. In the north-central U. S., the Madison and other formations contain thermally anomalous waters, whose origin is not fully understood, over an area of 25 000 km^2. 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 Gulf Coast area of the U. S., 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. Assessment of the effects of producing these resources is a topic of 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). Similiar experiments have been carried out in England (Batchelor, 1982). This work indicates that it is technologically feasible to induce 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.

As a matter of convenience, 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 in the range 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.

The preponderant use of geophysics, by a wide margin, has been in the exploration for and delineation of moderate- and high-temperature hydrothermal resources. For this reason, the remainder of this paper 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 for 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 to be 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 geological, geochemical or hydrological models of hydrothermal systems, it is appropriate in this review 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 are systems of hot, briny circulating fluids that are highly reactive chemically. Models for such systems have been discussed by 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 (≈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 the dissolution of primary minerals and the 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 the 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.A.; 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. 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 to interact with 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, it is the deep chloride water that 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 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 non-boiling 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) note that of 11 hydrothermal fields investigated in New Zealand, all of which have high temperatures (230°C to 300°C), five are non-productive 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 by leaching by the thermal fluids. Although none of the geophysical methods maps permeability directly, any geological, 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 and also shows the common geothermal targets for these methods. 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 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.

The interpretation of temperature, thermal-gradient, and heat-flow data and the evaluation of resource potential from these measurements can be quite complex, as discussed in detail by several authors (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 Mount 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 while 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 while heat flow values ranged 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 and several have provided methods to do so (Birch, 1950; Blackwell et al., 1980). Continuation of heat flow data for the purpose of determining subsurface isotherms in geothermal areas has been discussed by Brott et al. (1981), who conclude 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) give a review of previous work in this topic and report on numerical solutions 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). A great deal more work remains to be done at this interface between geophysics on the one hand and hydrology and reservoir engineering on the other hand.

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) have 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 affects the amount of energy absorbed, (5) variations in surface roughness, which affects 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 in 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 emerged 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 Valley, New Mexico. In the absence of near-surface cold-water flow, 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 were made hours to days after light to moderate snowfall. The thermally anomalous areas were visible because the snow melted faster over these areas than it did over non-thermal 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 hot springs that were already known and located other areas of hot ground that were previously unknown. Later ground-truth surveys determined that more than 90% of the areas indicated as anomalous on IR imagery had actual ground temperatures above ambient.

Dickinson (1976) gives 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 used 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 areal photographs and field checking helped to 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. Vegetation over thermally anomalous areas was also found to exhibit elevated temperatures, and so 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 increasing porosity, increasing salinity, and increasing 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, it may be the wet magmas that we seek, because they have a high enough content of volatile elements to produce the fracturing needed for hydrothermal convection to develop.

Thermal brine and alteration may 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) have recently reviewed the applications of electrical methods in mining exploration, and many of the points made in their article apply also to geothermal exploration.

Galvanic Resistivity. - The uses of the Schlumberger and Wenner arrays have been 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), among many others. The Schlumberger array is the most convenient one for depth sounding, i.e. estimation of the thicknesses and resistivities of the layers of a horizontally layered earth (Palmason, 1976). Successful use of the head-on Schlumberger method has been reported by Lezama (1984), among 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 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 two- or three-dimensional modeling techniques, there is often not enough data to do this. One must be very careful when using sounding techniques in areas of complex structure or lithologic variations.

The bipole-dipole array has been used in geothermal exploration by 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 achieved early success over broad areas of resistivity lows caused by hydrothermal alteration, but it has subsequently fallen into disfavor because of its failure to produce distinctive anomalies over some geothermal systems (Dey and Morrison, 1977). Also, the reduced resistivity values are strongly dependent on the local resistivity distribution in the vicinity of the transmitting dipole (Frangos and Ward, 1980).

Dipole-dipole arrays were used in surveys reported by Beyer (1977), Fox (1978a), Ward et al. (1978), Baudu et al. (1980), Patella et al. (1980), and Wilt et al. (1980) among many others. This array is widely used in geothermal, mineral and petroleum exploration because it is an efficient means of collecting a large number of data points from which lateral variations in resistivity can be separated from variations in resistivity with depth by suitable interpretation. 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 great advantage of the dipole-dipole technique in discriminating between vertical and horizontal resistivity boundaries and commented that resistivity surveys in general were 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. Above and flanking this region, resistivities showed a systematic decrease with time. These decreases were more difficult to explain, but there appeared to be a component related to ascending hot, more saline, fluids at the eastern edge of the producing zone.

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

Controlled-Source Electromagnetics (CSEM). - Keller (1970) made a baseline review of the applications of active and passive electromagnetic methods in geothermal exploration. Subsequently, a number of articles have appeared which illustrate the success and failure of these methods. Included are the articles by 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) describe 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 have been available for only the layered-earth, one-dimensional case. If the subsurface has a resistivity distribution that is two-dimensional or three-dimensional, as it usually does in hydrothermal environments, interpretation using one-dimensional 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 electric and magnetic field are measured (as in magnetotellurics). Sandberg and Hohmann (1982) have evaluated its use in the Roosevelt Hot Springs, Utah geothermal system. It offers advantages over galvanic resistivity methods in that it is sometimes faster and suffers less from the effects of lateral resistivity variations when providing sounding information (Ward, 1983a).

Scalar Audiomagnetotellurics (AMT). - The AMT method utilizes either natural or artificial electromagnetic 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, have reported its use in geothermal exploration. The method suffers from two particular problems. First, the natural fields occasionally are too weak to obtain useful information. Second, and far more important, the scalar data are totally inadequate for interpretation in two- and three-dimensional terrains, in which the tensor AMT method should be used. The CSAMT method is a substantial improvement over scalar AMT insofar as the direction of the inducing fields can be controlled, thus simplifying interpretation in two- and three-dimensional environments. 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). A comprehensive review of data acquisition, processing, and interpretation for the method, plus a full discussion of the problems it encounters in geothermal exploration, has been prepared by Ward and Wannamaker (1983).

The tensor magnetotelluric/audiofrequency magnetotelluric method is usually too expensive to be used for mapping the resistivity distribution in the shallow parts of a geothermal system. Hence, it is more logically used to map regional structure, to map the deeper parts of convective hydrothermal systems, to attempt to map magma chambers, and to detect and delineate zones of partial melt in the deep crust and upper mantle. Some workers have used telluric current data in combination with a reduced number of MT stations in order to help 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 may be combined in calculations with electric data from the remote stations to 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 of the high-temperature resource exploration programs in the western United States. We attribute this to its advertised great depth of exploration and to a common assumption that it is able to detect the molten or partially molten source of heat. Neither of these attributes is necessarily correct. The conductivity of magma at elevated temperatures is strongly dependent upon the partial pressure of water (Lebedev and Khitarov, 1964) and so a dry partial melt will be more difficult to detect by MT than a wet partial melt. Depth of exploration depends to a certain extent on the near-surface resistivity structure. Also of great importance is the size and other characteristics of the magma body. Newman et al. (1985) have explored conditions under which crustal magma bodies can be detected. They conclude 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 two- or three-dimensional 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 anomalies over convective hydrothermal systems arise from the 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 self-potential 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) has developed interpretation techniques that have considerable potential for solving some of these problems.

Telluric Currents. - The telluric method is mainly suitable for reconnaissance of horizontal resistivity variations. It is based on the assumption that telluric currents flowing in extensive sheets are affected by lateral variations in the resistivity structure, which can be caused, for example, by variations in geological structure or by hydrothermal systems. The method requires the simultaneous measurement of the telluric electric field at two stations. From the ratio of the amplitudes of the electric field at the two stations, inferences may be drawn about variations in the underlying resistivity structure. By keeping the base station fixed and moving a field station about, one can thus map resistivity variations in a qualitative way (Pálmason, 1976).

The method has been used in geothermal exploration by Beyer (1977), Isherwood and Mabey (1978), Jackson and O'Donnell (1980), and others. It appears to be a convenient method for regional surveys in order to detect areas worthy of more detailed exploration by resistivity methods. The method suffers from a number of problems which include random noise, geological noise due to overburden, lack of resolution, and effects of topography, but the main problem is that it is a semi-quantitative method 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 that are 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 the location of 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 has been 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 have also been 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) reports a contiguous zone of gravity lows west of the High Cascades in central Oregon and notes 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) report similar interpretations. Williams and Finn (1982) report 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. Their use can be 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 being due to a conversion of magnetite to pyrite (Studt, 1964). A magnetic low occurs over a part of the hot spring area at Long Valley, and is interpreted by Kane et al. (1976) as due to 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 have examined from Coso Hot Springs, California, from Baltzaor, 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 (Palmason, 1976; Flovenz and Georgeson, 1982).

Magnetic data may also yield regional information of value in explanation. 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) interprets 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 are dependent on 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 in 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 can thus 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 did not detect body waves at Roosevelt Hot Springs. Liaw and McEvilly (1979) failed to find body waves at Leach Hot Springs, Nevada, but did find microseismic energy propagating as fundamental-mode Rayleigh waves from the vicinity of the thermal manifestations. Their paper, additionally, presents 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 must be done to produce valid results using microseismic techniques.

Microearthquakes. - Microearthquakes frequently are closely related spatially to major hydrothermal convection systems. Accurate locations of these earthquakes can provide data on the locations of active faults that may channel hot water toward the surface (Ward and Björnsson, 1971; Ward et al., 1979; Lange and Westphal, 1969; Hamilton and Muffler, 1972). Microearthquake (MEQ) surveys have been completed in several geothermal areas including those in Iceland (Ward and Björnsson, 1971), at 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 called "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 value for Q 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 Spring due to silica densification of sediments. Gertson and Smith (1979) found high Q over the geothermal system at Roosevelt Hot Springs, using refraction data.

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 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 microearthquake data, a good model of the subsurface velocity distribution is required. Lack of good velocity control is a principal problem in analysis of MEQ data. Some geothermal systems, such as Roosevelt Hot Springs, have a generally low, episodic 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, one can record passive seismic data for a two- or three-week period or longer and erroneously conclude that the geothermal system is unimportant since it is not seismically active during the time of recording.

Teleseisms. - If a sufficiently distant earthquake is observed with a closely spaced array of seismographs, changes in P-wave traveltime from station to station can be taken to 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, 1976b) 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. While one can speculate that relative P-wave delays are caused by partial melts or magmas, as may be the case at Coso, Long Valley, and The Geysers, they can also be caused by alluvium, alteration, compositional differences, lateral variations in temperature or locally fractured rock (Iyer and Stewart, 1977). Wechsler and Smith (1979) suggest that the P-wave delays found by Robinson and Iyer (1981) at Roosevelt Hot Springs may well be due to fluid-filled fractures or to a compositional change.

Refraction. - The seismic refraction and reflection methods can be used to map the depth to the water table, stratigraphy, faulting, intrusions, and geologic structure in general. They may also yield the subsurface distribution of seismic P-wave and S-wave velocities, attenuations and Poisson's ratio. Detection of a characteristic attenuation or a "bright" spot, as found over reservoirs in petroleum exploration, would be a useful feature (Ward 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; Ackerman, 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 does the seismic reflection method. Sentiment today calls for performing seismic refraction at the same time as seismic reflection, with little added cost. Some attempts have been made to map velocity and amplitude attenuation anomalies, of both P- and S-waves, coinciding with a geothermal system (Goldstein et al., 1978). Majer and McEvilly (1979) report 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 Spring, and Roosevelt Hot Springs, respectively. Majer (1978) interprets the high Q determined at Leach Hot Spring to result 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 appear to give good definition of Basin and Range border faulting and depths to the base of alluvial fill at Roosevelt Hot Springs, UT, Soda Lake, NV, San Emidio, NV, Dixie Valley, NV, and Grass Valley, NV. One seismic line which crosses the Mineral Mountains at Roosevelt Hot Springs shows little obvious lithologic or structural information within the range itself, or within the reservoir, but substantial structural information along the range front (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 extremely useful in delineating structure in Grass Valley, NV. At Soda Lake, in 1977, Chevron obtained 1 200% CDP seismic reflection coverage. The seismic data delineated a complex NE-SW trending graben from the shore of Soda Lake passing south of Upsal Hogback. The reflectors dip to the southwest, consistent with a small basin over the gravity low. The maximum depths of reliable seismic data are governed by a thin basalt unit and vary from 730 to 1 220 m (Hill et al., 1979).

Zoback and Anderson (1983) demonstrated the 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) was 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, Nevada. 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 are present in a geothermal system, they will be detected in the ^{214}Bi peak, since they also are daughter products of ^{238}U decay. An examination of hot-spring waters in Nevada indicates the presence of ^{226}Rn and ^{222}Ra, in varying abundances, in spring systems where $CaCO_3$ is the predominant material being deposited. Systems where silica predominates are relatively low in radioactivity (Wollenburg, 1976). The use of alpha-cup detectors for radon emanating from hydrothermal systems has been reported by Wollenburg (1976) and Nielson (1978). Surface radon emission surveys appear to be capable of detecting open channels that may conduct geothermal fluids. Nevertheless, very little use has been made of the method in geothermal exploration.

Geophysical Well Logging

Much research remains to be done in order to understand fully the responses of various well logs in geothermal 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 of the 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. Several papers have discussed these items and the 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 which we call 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 down-hole 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 3-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 shot points, 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 the tube waves (Paillet, 1980). Crampin (1978, 1984) and others have argued that VSP conducted with 3-component geophones might prove extremely useful for mapping the fractured conditions of rocks if one were to extract seismic anisotropy information from the shear-wave 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 (1985) 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. It was found that locating the buried electrode just outside the body did not significantly alter the results from those when the electrode is embedded in the inhomogeneity.

From the above studies we tentatively conclude the following: the cross-borehole method produces larger anomalies than does a single-borehole method; the cross-borehole anomalies using a pole-pole array are smaller than those for a cross-borehole dipole-dipole array; the cross-borehole mise-à-la-masse method produces larger anomalies than for the other cross-borehole methods, and the anomalies due to a thin sheet were generally much smaller than those for a sphere, as is to be 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 by 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, California. Integrated interpretation of the data yielded a cross-section of the caldera showing the ring fracture system near Minaret Summit, the configurations of post-caldera volcanics and welded Bishop Tuff and the interface with basement rock of the Sierra Nevada. Such detailed studies show great promise to provide structural details in the geothermal environment.

Lager and Lytle (1977) adopted the technique for high frequency electromagnetic measurements between boreholes. Daily et al. (1982) and Daily (1984) describe the application of borehole-to-borehole measurements using radio frequency signals between 1 MHz and 40 MHz to map the electromagnetic attenuation of oil shale. The method appears to be 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) describe 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) present numerical modeling results which illustrate that cross-borehole resistivity measurements are much more effective than single-borehole measurements for detecting deeply buried fractures and ore deposits. Their model results suggest 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 very large amount of data available, we have chosen just one area for which the 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 my and $\leq$ 1.1 my (Duffield et al., 1980), and formed domes, flows and pyroclastic deposits which covered much of the 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) describe 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 m 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 ($15 \text{ m} \leq \text{depth} \leq 65 \text{ m}$). A map of heat flow for a deeper interval ($35 \text{ m} \leq \text{depth} \leq 300 \text{ m}$) 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) describe shallow temperature surveys completed at Coso. The 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 HFU contour of Combs (1980) shown in Figure 4. In their later paper, LeSchack and Lewis (1983) describe 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 audiomagnetotelluric soundings have been completed in an area of approximately 900 km^2 which includes the Coso geothermal area (Jackson et al., 1977). Jackson and O'Donnell (1980) present an interpretation of these and other data and note 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 the 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 Ω·m low-resistivity zone includes Coso Hot Springs, Devil's Kitchen and much of the $\leq$ 16 Ω·m resistivity low defined by 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 Ω·m) were mapped along the north side of Sugarloaf Mountain, a rhyolite dome. Rather uniform low apparent resistivities (5-10 Ω·m) to the east occur where alluvium overlies granitic rocks. The central portion of the line is dominated by low to moderate (10-30 Ω·m) resistivities, and resistivity increases with increasing separation. Occasional lower resistivities, 4-10 Ω·m, occur near Devil's Kitchen and along an ENE-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 clearance.

Basement lithologic and structural information are apparent even when the data are presented at a 200 nT contour interval, as shown in Figure 8. Fox (1978b) identified NE- and NW-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 is due in part to magnetite destruction resulting from hydrothermal alteration by the geothermal system, as well as to 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 by Walter and Weaver (1980). The seismicity of the Coso geothermal area has been reported in some detail by Combs and Rotstein (1976) and most recently by Walter and Weaver (1980) who provide 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 seventy-eight events which could be 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. Geological Survey studies to evaluate the geothermal resource potential. They recorded 4 216 local earthquakes (0.5 < m < 3.9) during the first 2 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 are 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 by Combs and Rotstein (1976). This 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 was considered insufficient to determine the geothermal production capability of the fault system.

In an accompanying paper, Young and Ward (1980) presented a three-dimensional 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 attenuation 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.

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 by 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. The locations of several successful wells drilled by California Energy Corporation are also shown. This drilling has 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 on-going 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. A total of 554 references was selected, and these 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 USA 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, saw significant utilization: galvanic resistivity sounding or VES (59%), gravimetric (52%), temperature gradient (50%), heat flow (48%), magnetic (39%), MT (35%), dipole-dipole resistivity (33%), reflection seismology (33%), MEQ (32%), remote sensing (28%), and bipole-dipole resistivity (26%). The least-used methods included CSAMT, IP, pole-dipole resistivity, earth noise, and geomagnetic soundings. The greatest utilization of geophysics is in the exploration for moderate- and high-temperature resources where the potential value of the resource justifies the application of costly exploration methods. The exploration for lower temperature resources, in addition, is often limited to the immediate vicinity of potential users, as in the Paris basin. An average of 6 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 familarity 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 one would expect. 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 has been successfully employed in one environment is tried in other geologic settings for which it is not appropriate. Ward (1983b) provides 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 were 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 exploration and assessment of geothermal resources is a relatively young field which has evolved rapidly by drawing on pre-existing petroleum and mining technology. Exploration to date has focused on high-temperature systems, many of which had some surface expression. It 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 attempt an evaluation of the present state of the art.

STATE OF THE ART

In this section we give a brief evaluation of 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 measuring 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 that will be 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 in order to understand observed thermal gradient and heat flow patterns. Continued development of two- and three-dimensional 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 exploration for concealed resources in rugged volcanic terrains. Galvanic resistivity surveys, while relatively easy to run and for which interpretation methods are reasonably well worked out, often lack adequate depth 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. As such it 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 two- and three-dimensional 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 would seem 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 appear to be adequately developed. Two-dimensional and three-dimensional interpretation algorithms are available but are still not in universal use. CSEM methods have a great deal of 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 two-dimensional and three-dimensional interpretation techniques. These geophysical techniques also need further field evaluation. Much of the data collected in the past has been interpreted using layered-earth models, and we consider these to be inadequate in most geothermal environments. The scalar AMT technique is adequately developed and tested so that we can conclude that it has only limited use in geothermal exploration. Tensor MT and AMT are still classed as largely 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 has long suffered from lack of interpretation techniques to facilitate the level of information needed in order to make a decision about whether or not to drill test an anomaly. Substantial gains in quantitative interpretation theory have been made since 1980, however, but the algorithms are probably in limited use. Provided that the technique can be understood better, more field evaluation might be warranted. We consider the telluric current method to have limited application because of its semi-quantitative nature and to be 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 to be made 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 about five years ago. Continued work on data processing and interpretation as well as further testing in geothermal environments appears to be warranted. Microearthquake surveys have potential to contribute to defining drill targets especially for deep or blind hydrothermal systems. Less expensive equipment is needed so that 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 to be adequate for routine application where appropriate, although we recognize that advances will continue to be made.

The *seismic reflection* method has potential for greater contributions to geothermal work than it has made to date. The method will always be expensive per unit of coverage, but if the information derived could be increased, an adequate payout may result. Portable, high resolution gear is now just becoming available for shallow reflection work in the hard-rock environment. 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 (see Stolt and Wegelein, 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 in the past.

Radioactive Methods

It is unlikely that conventional radioactive methods will play a significant role in geothermal exploration. We deem them to be adequately developed. See Grasty et al. (1985, this issue) for a discussion of current interpretation practices for multi-channel 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 have been summarized in Sanyal et al. (1980) and by 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 exist 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 petroleum industry (Snyder and Fleming, 1985, this issue) is available to the high temperature geothermal industry.

Borehole Geophysics

As a general statement, borehole geophysics has not undergone the development required even to assess its potential contribution to geothermal development. The *VSP techniques* have emerged as being important in petroleum exploration, and development for these purposes will have important spin-off for 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.

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FIGURE CAPTIONS

FIG. 1. Schematic geologic models for high-temperature systems.

1a) 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).

1b) 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).

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.

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

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

FIG. 4. Heat flow map of the Coso geothermal area, California (after Combs, 1980). Heat flow values and contours in $\mathrm{mW/m^2}$ for the upper zone ($15\ \mathrm{m} \le \mathrm{depth} \le 65\ \mathrm{m}$).

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

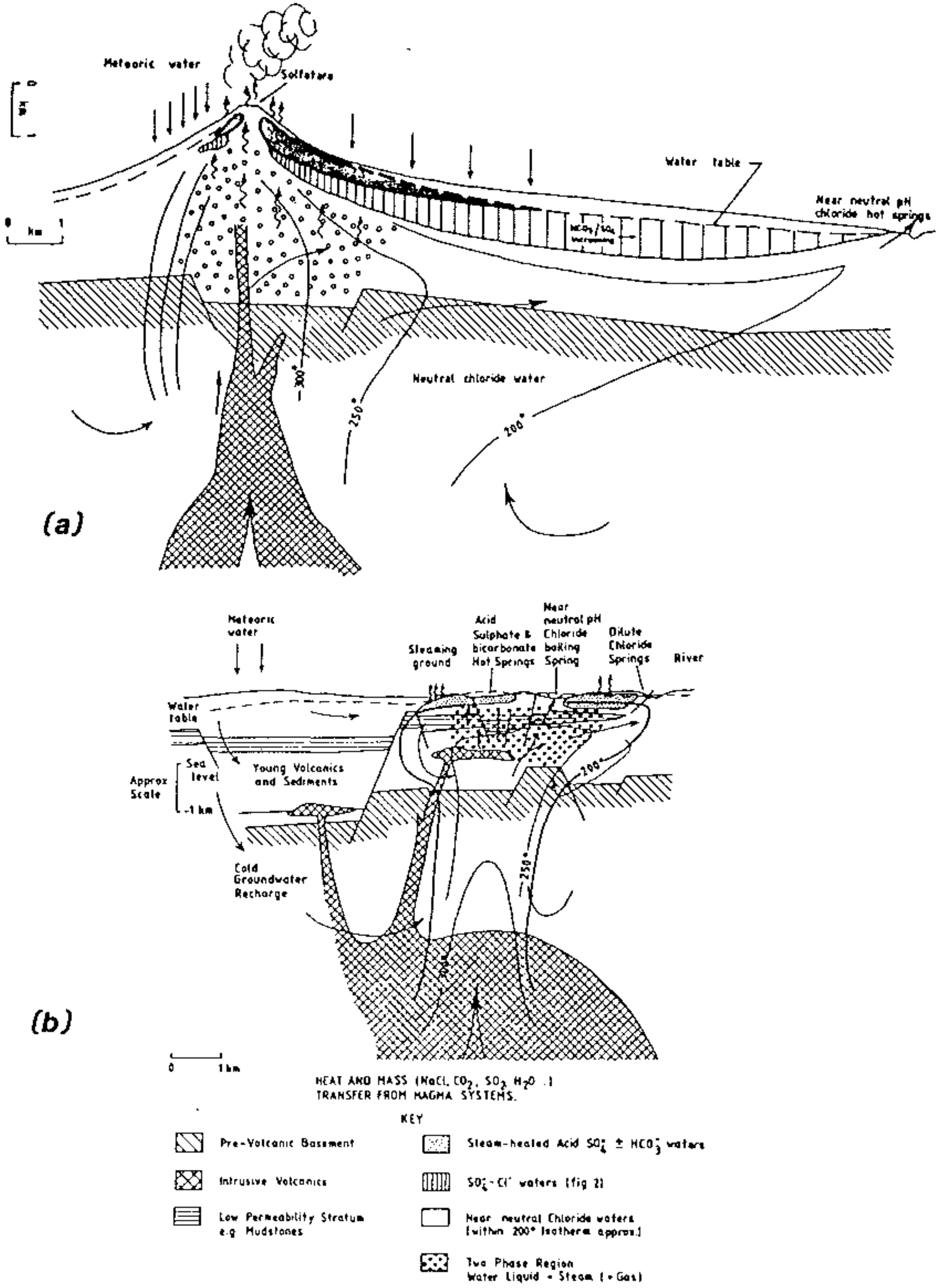
FIG. 6. Dipole-dipole electrical resistivity survey of the Coso geothermal area, California. Dipole lengths of 300 m and 150 m were used. The contoured apparent resistivity in $\Omega\cdot\mathrm{m}$ is shown for third separation (n=3) values of the 300 m dipole lines (after Fox, 1978a).

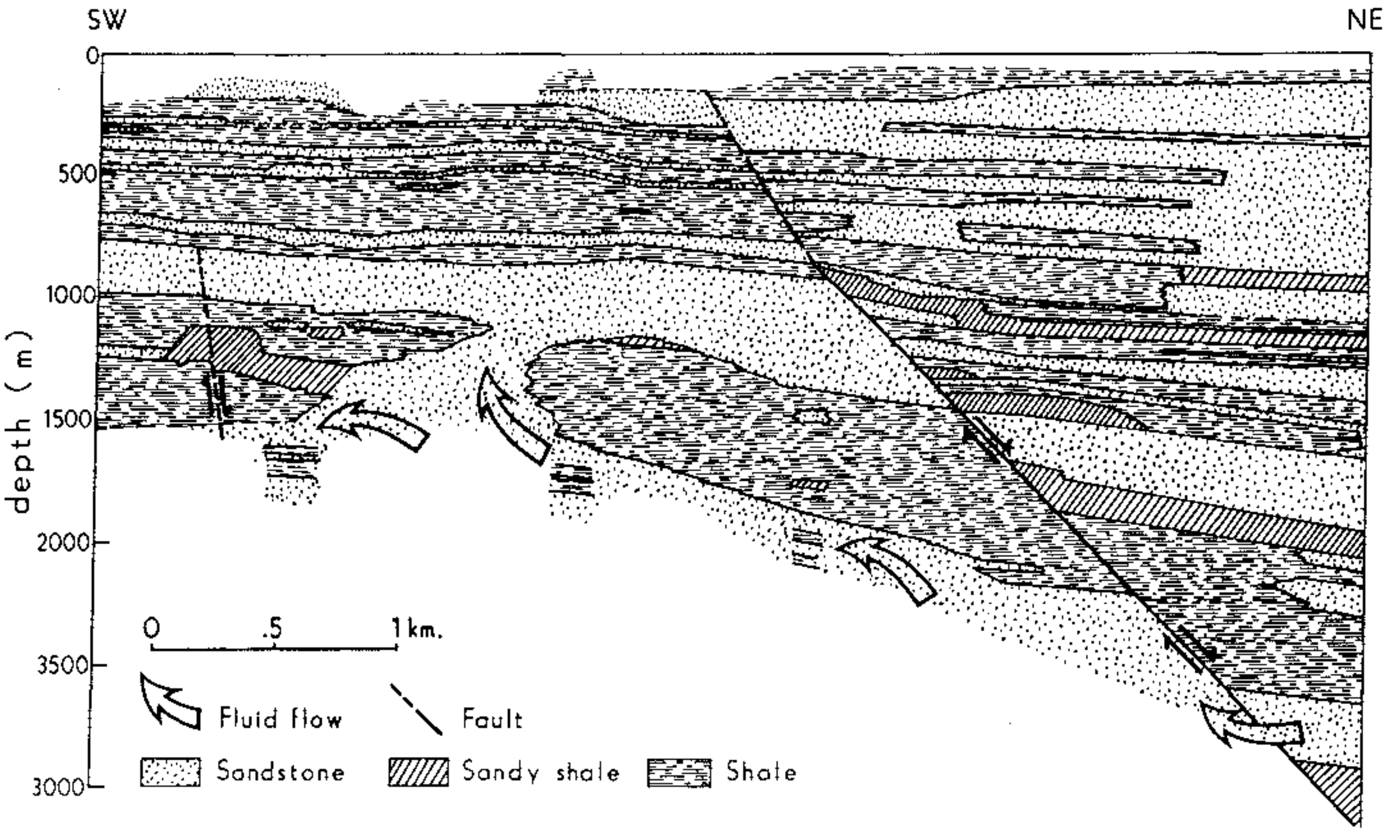
FIG. 7. Observed apparent resistivity ($\Omega\cdot\mathrm{m}$) and geologic cross-section for Line 1 (300 m dipoles) at Coso geothermal area. Line location shown in Figure 6. Geologic cross-section is modified from Hulen (1978): resistivity data from Fox (1978a).

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

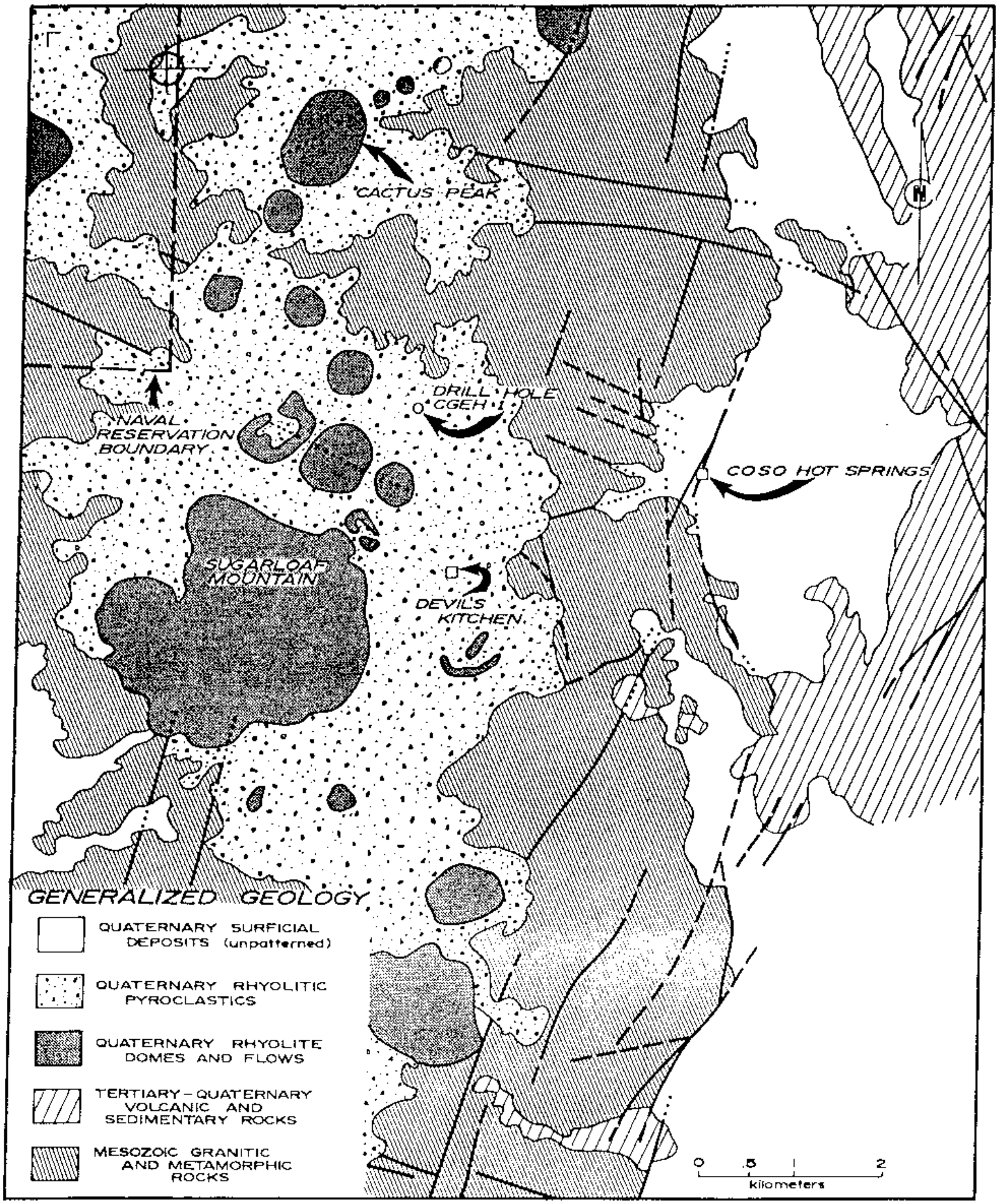
FIG. 9. Located earthquakes in the Coso Range, September 27, 1975 to September 30, 1977. Box outlines geothermal subarea shown in other figures. R indicates Red Hill seismic zone, B locates seismic zone on west side of Coso Basin. Cross-hatching designates Sierra Nevada on west and Argus range on the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

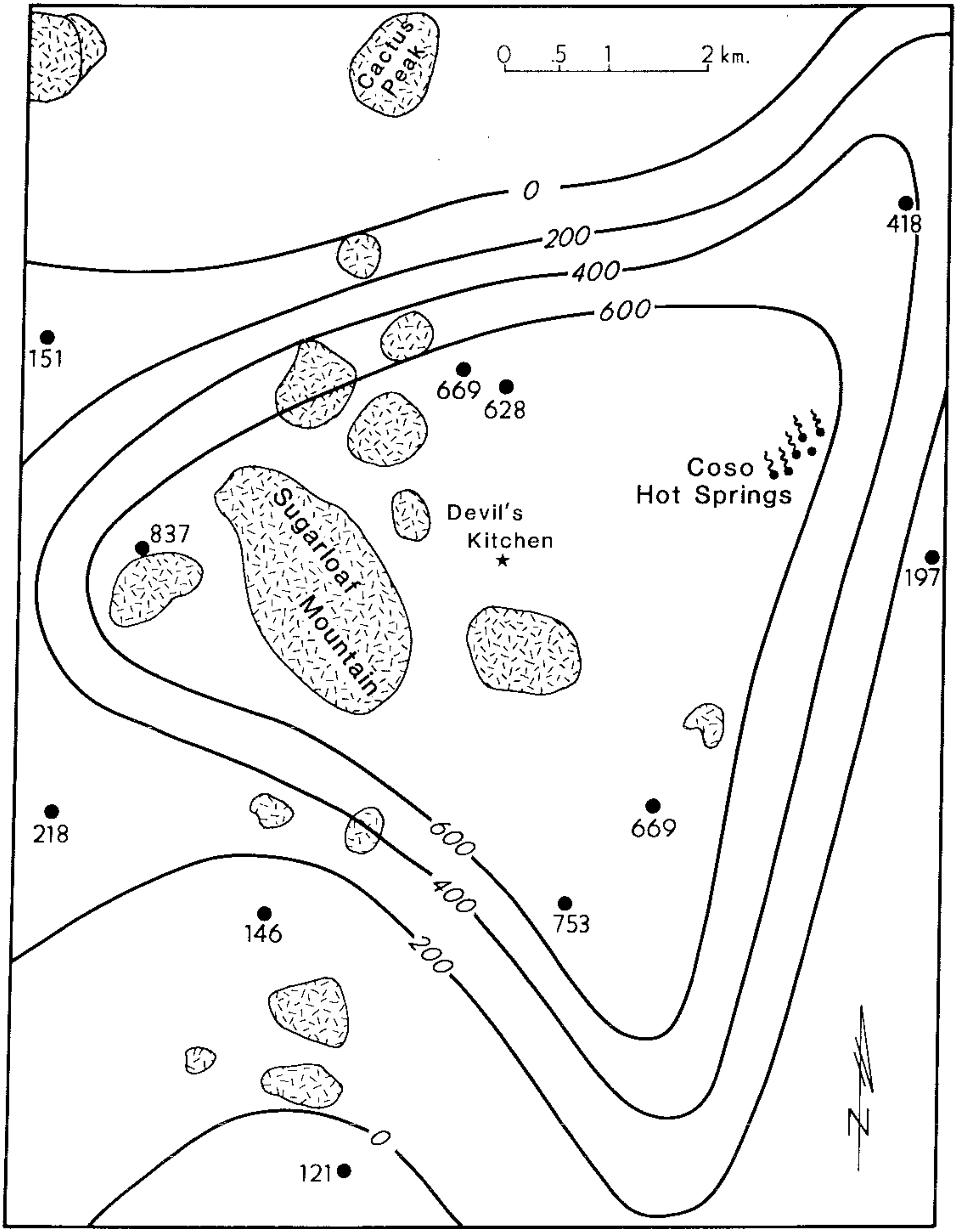
FIG. 10. Geophysical anomaly summary for the Coso geothermal area. Thermal manifestations, alteration areas and drill holes are shown for reference (after Hulen, 1978).

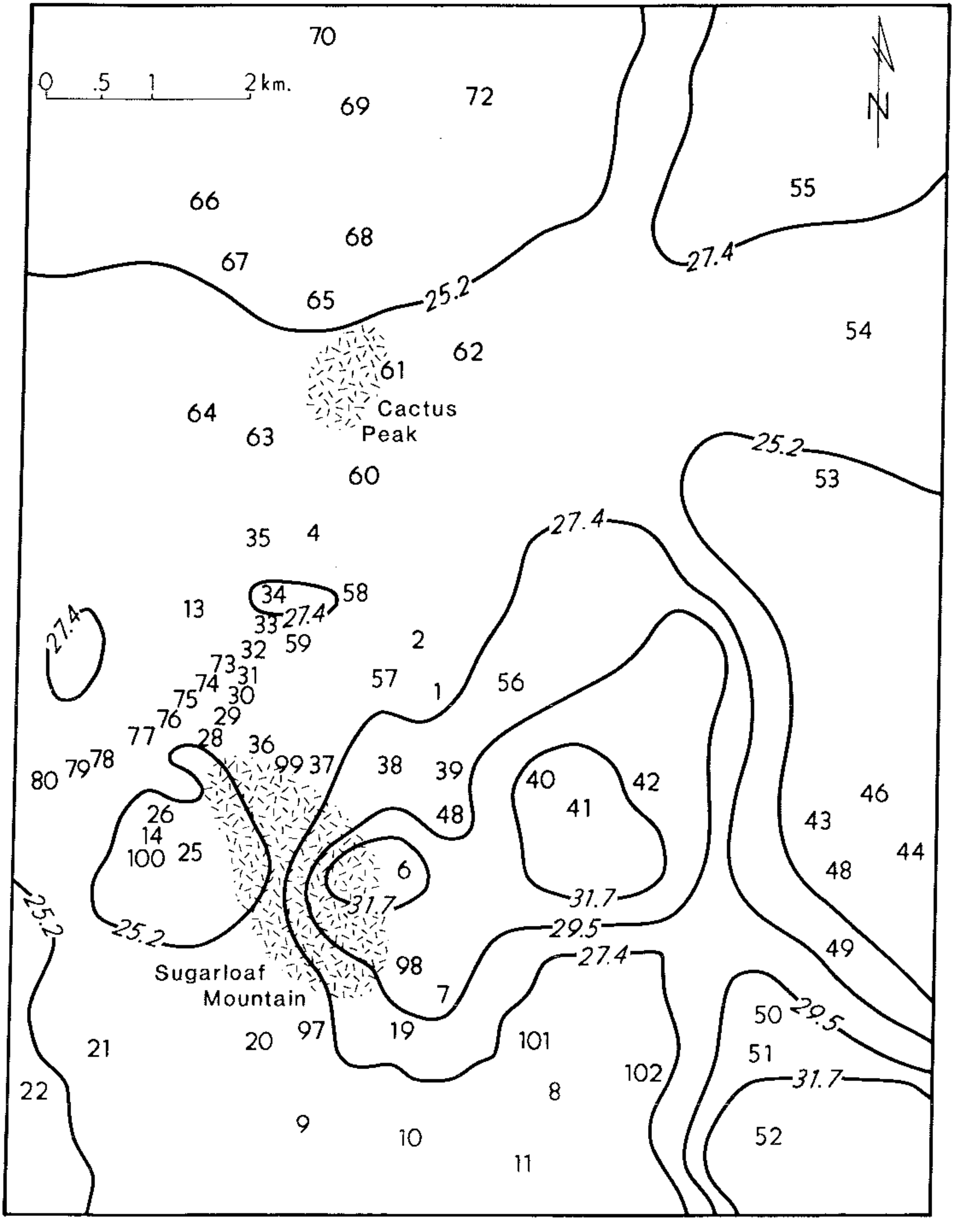

Meteoric water
Solfatara
1 km
0 km 1
Water table
Near neutral pH chloride hot springs
HCO3/SO4 increasing
Neutral chloride water
300°
250°
200°
(a)
Meteoric water
Steaming ground
Acid Sulphate & bicarbonate Hot Springs
Near neutral pH Chloride boiling Spring
Dilute Chloride Springs
River
Water table
Sea level
Approx Scale
-1 km
Young Volcanics and Sediments
Cold Groundwater Recharge
200°
250°
300°
(b)
0 1 km
HEAT AND MASS (NaCl, CO2, SO2, H2O ..) TRANSFER FROM MAGMA SYSTEMS.
KEY
Pre-Volcanic Basement
Intrusive Volcanics
Low Permeability Stratum e.g Mudstones
Steam-heated Acid SO4= ± HCO3- waters
SO4=-Cl- waters (fig 2)
Near neutral Chloride waters (within 200° Isotherm approx.)
Two Phase Region Water Liquid + Steam (+Gas)


SW
NE
0
500
1000
1500
2000
3500
3000
depth (m)
0
.5
1 km.
Fluid flow
Fault
Sandstone
Sandy shale
Shale

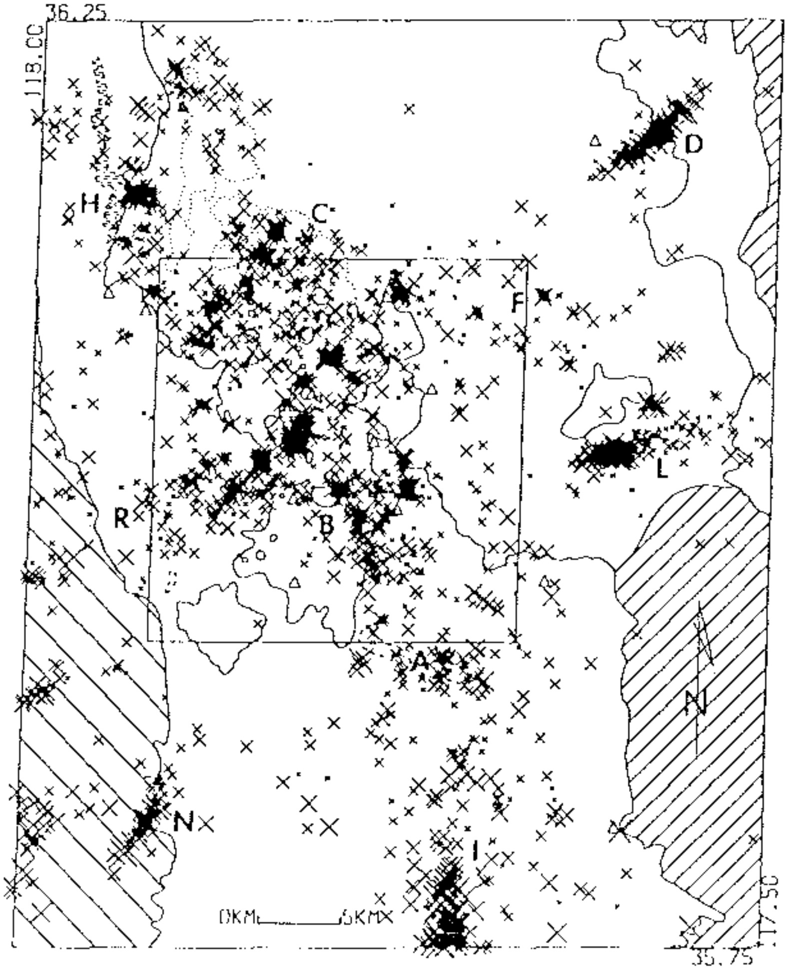

CALIFORNIA
0
40
80
MILES
N
COSO HOT SPRINGS KGRA
Redding
San Francisco
Fresno
Independence
Bakersfield
Santa Barbara
Los Angeles
San Bernardino

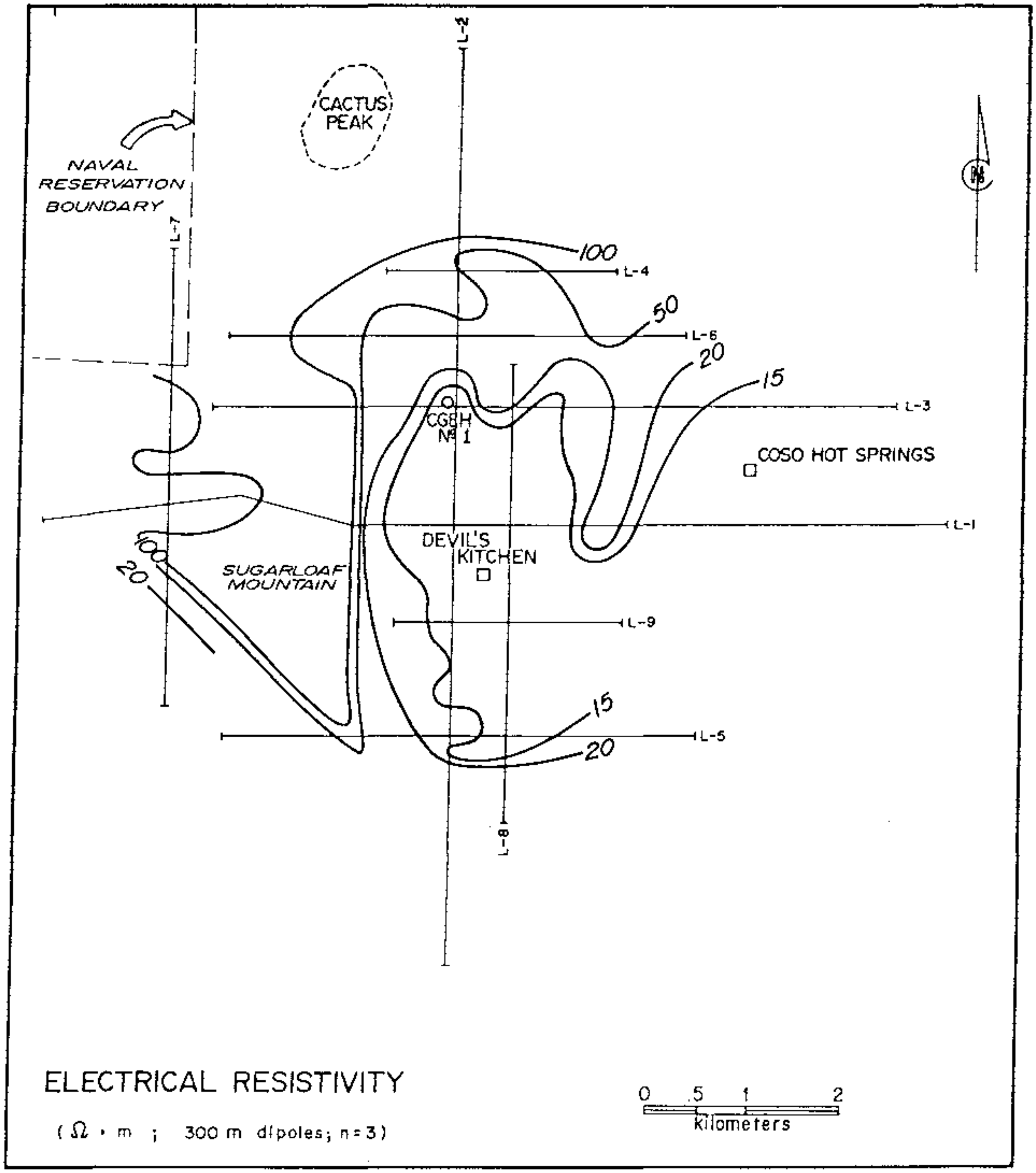

CACTUS PEAK
DRILL HOLE CGEH 1
NAVAL RESERVATION BOUNDARY
COSO HOT SPRINGS
SUGARLOAF MOUNTAIN
DEVILS KITCHEN
N
GENERALIZED GEOLOGY
QUATERNARY SURFICIAL DEPOSITS (unpatterned)
QUATERNARY RHYOLITIC PYROCLASTICS
QUATERNARY RHYOLITE DOMES AND FLOWS
TERTIARY-QUATERNARY VOLCANIC AND SEDIMENTARY ROCKS
MESOZOIC GRANITIC AND METAMORPHIC ROCKS
0 .5 1 2
kilometers


Cactus Peak
0 .5 1 2 km.
0
200
400
600
418
151
669
628
Coso Hot Springs
Devil's Kitchen
837
Sugarloaf Mountain
197
218
669
600
400
200
146
753
0
121
N


0 .5 1 2 km.
N
70
69
72
66
68
67
65
25.2
27.4
55
54
62
61
Cactus Peak
64
63
60
25.2
53
35
4
27.4
34
58
27.4
33
13
32
59
2
73
31
57
1
56
74
30
75
76
29
77
28
36
99
37
38
39
78
79
80
40
42
46
26
48
41
43
14
100
25
6
44
48
31.7
31.7
25.2
25.2
29.5
49
98
27.4
Sugarloaf Mountain
7
50
29.5
97
19
21
20
101
51
102
31.7
22
8
9
10
52
11

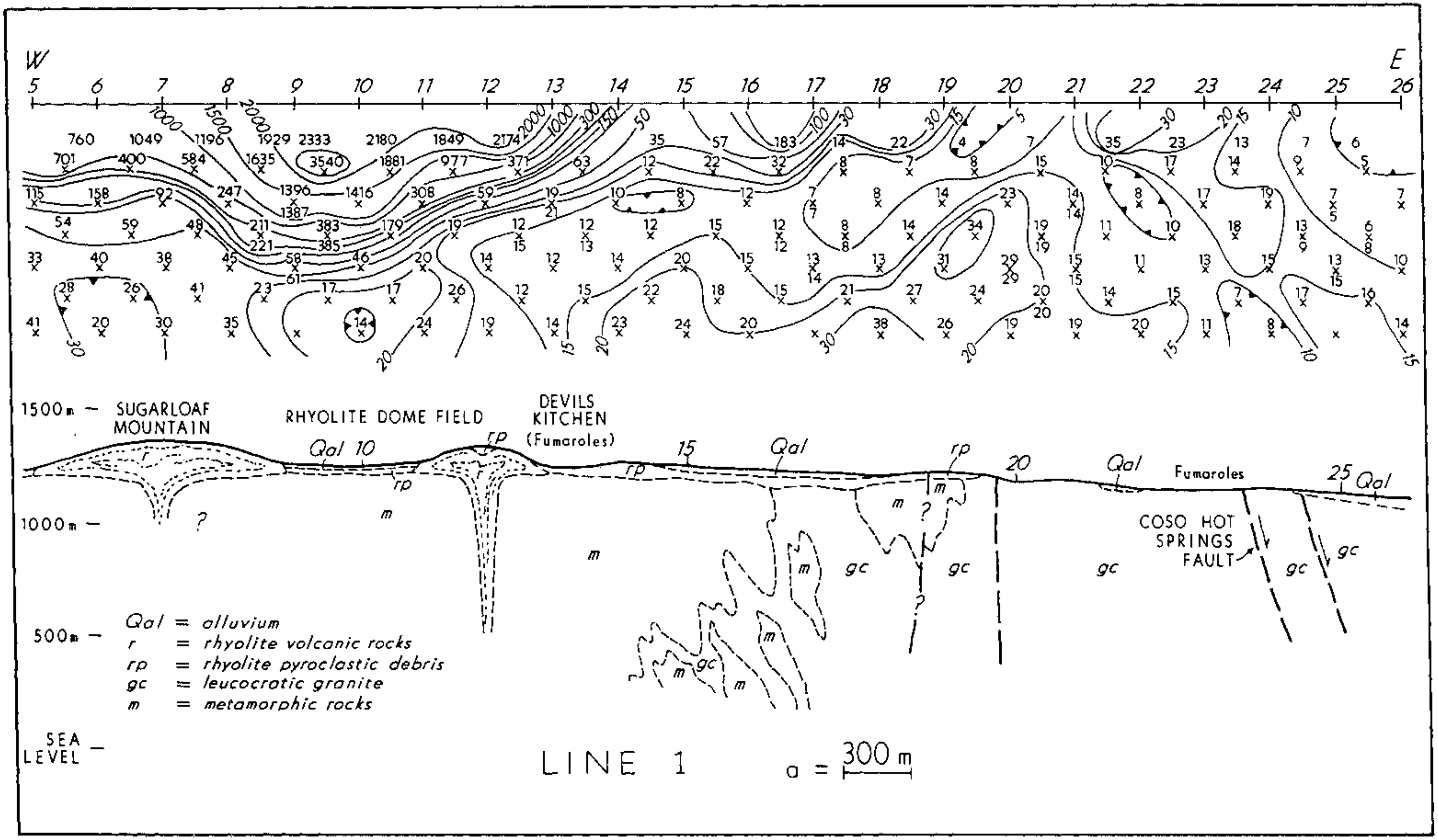
SEISMICITY 9-75 9-77





ELECTRICAL RESISTIVITY

($\Omega \cdot m$; 300 m dipoles; n=3)


W
E
5
6
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9
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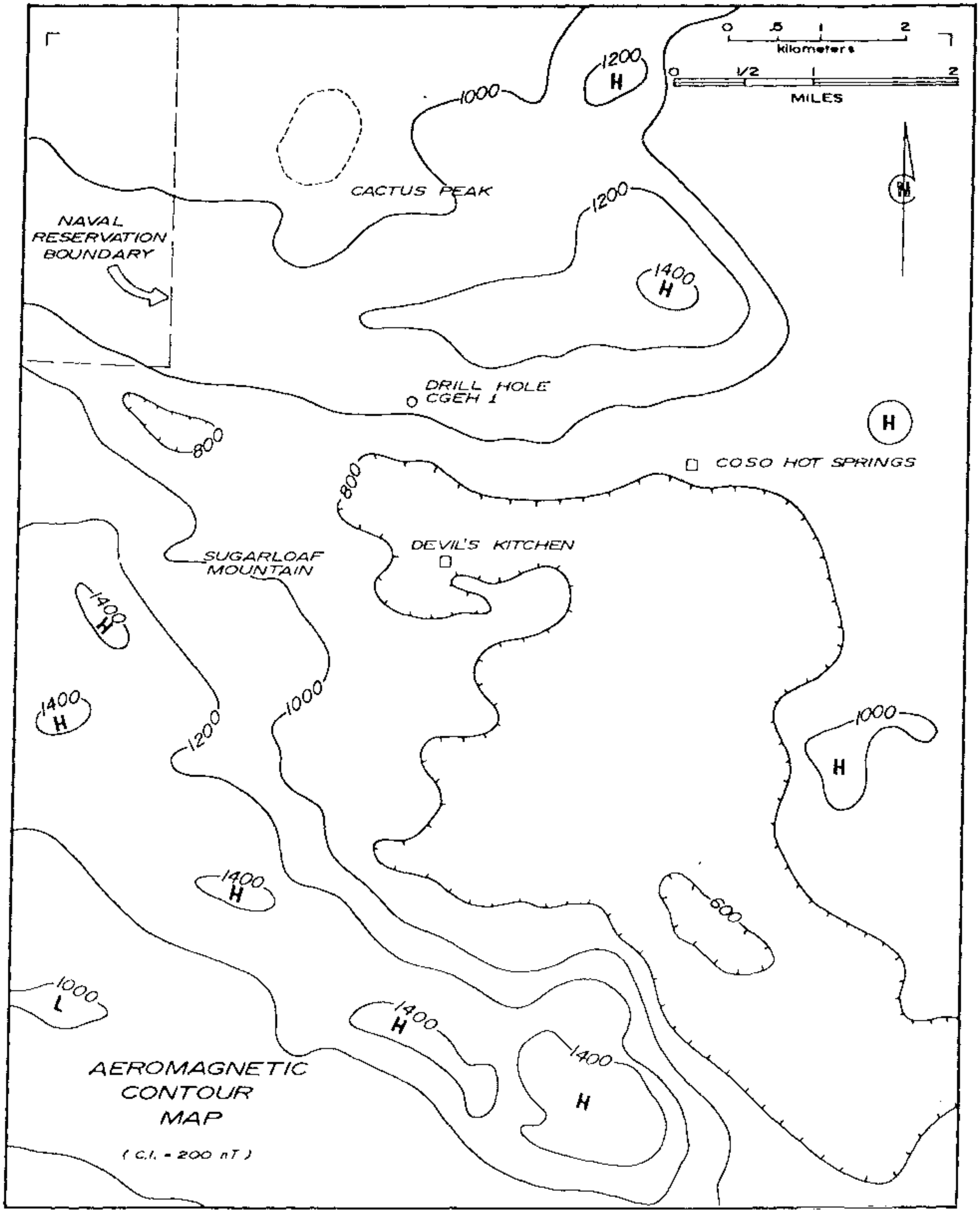
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
1000
1500
2000
300
150
50
100
30
15
5
20
10
3540
1500m
1000m
500m
SEA LEVEL
SUGARLOAF MOUNTAIN
RHYOLITE DOME FIELD
DEVILS KITCHEN (Fumaroles)
Fumaroles
COSO HOT SPRINGS FAULT
Qal
r
rp
m
gc
?
Qal = alluvium
r = rhyolite volcanic rocks
rp = rhyolite pyroclastic debris
gc = leucocratic granite
m = metamorphic rocks
LINE 1
a = 300 m



AEROMAGNETIC
CONTOUR
MAP

(C.I. = 200 nT)

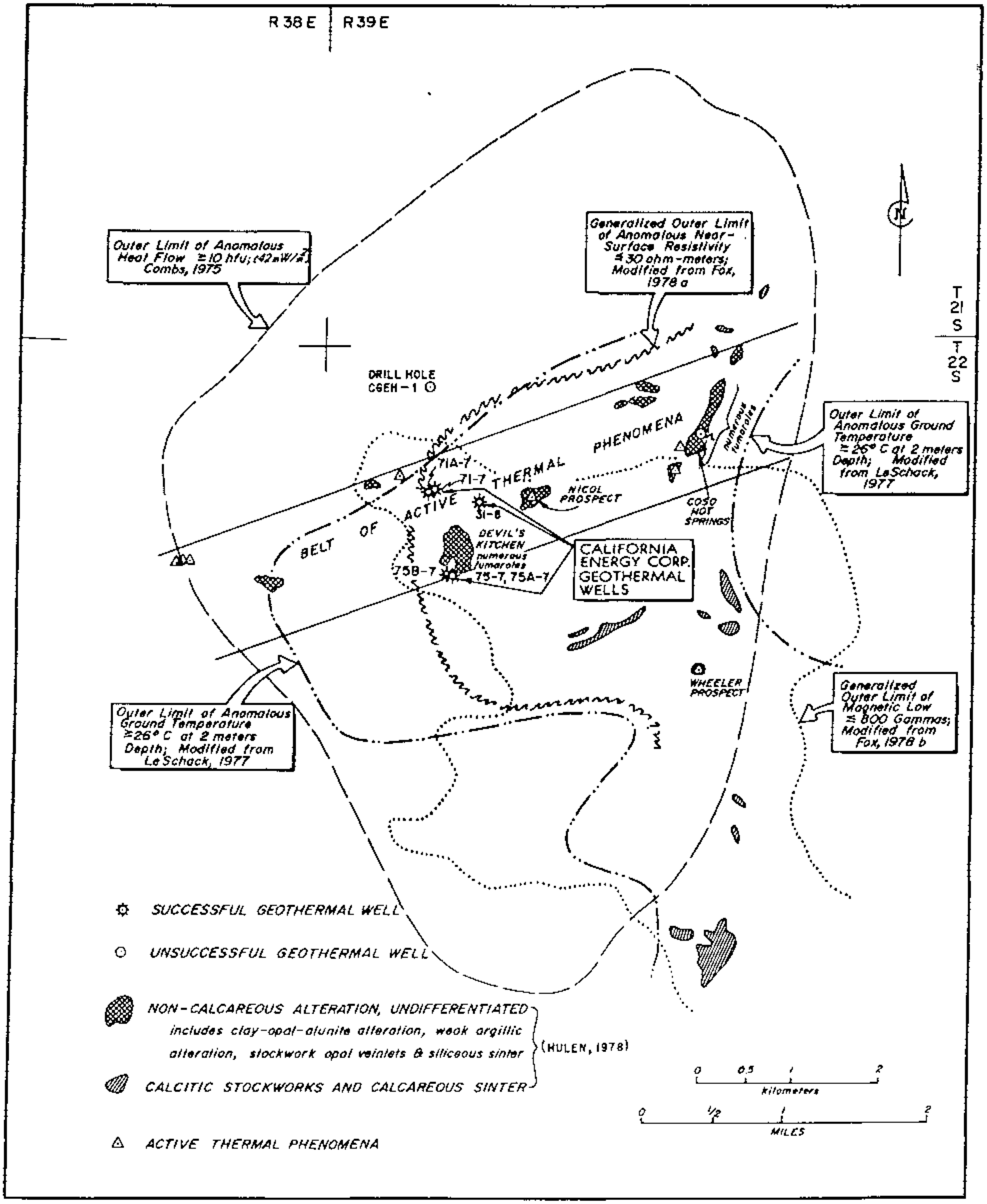

R 38 E
R 39 E
T 21 S
T 22 S
N
Outer Limit of Anomalous Heat Flow ≥10 hfu; (42 mW/m²) Combs, 1975
Generalized Outer Limit of Anomalous Near-Surface Resistivity ≤30 ohm-meters; Modified from Fox, 1978 a
DRILL HOLE CGEH-1
BELT OF ACTIVE THERMAL PHENOMENA
71A-7
71-7
31-8
75B-7
75-7, 75A-7
DEVIL'S KITCHEN numerous fumaroles
NICOL PROSPECT
numerous fumaroles
COSO HOT SPRINGS
CALIFORNIA ENERGY CORP. GEOTHERMAL WELLS
Outer Limit of Anomalous Ground Temperature ≥26° C at 2 meters Depth; Modified from LeSchack, 1977
WHEELER PROSPECT
Generalized Outer Limit of Magnetic Low ≤800 Gammas; Modified from Fox, 1978 b
Outer Limit of Anomalous Ground Temperature ≥26° C at 2 meters Depth; Modified from LeSchack, 1977
☼ SUCCESSFUL GEOTHERMAL WELL
○ UNSUCCESSFUL GEOTHERMAL WELL
NON-CALCAREOUS ALTERATION, UNDIFFERENTIATED includes clay-opal-alunite alteration, weak argillic alteration, stockwork opal veinlets & siliceous sinter
CALCITIC STOCKWORKS AND CALCAREOUS SINTER
(HULEN, 1978)
△ ACTIVE THERMAL PHENOMENA
0 0.5 1 2
kilometers
0 1/2 1 2
MILES

TABLE CAPTIONS

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

Table 2. Geophysical targets in geothermal exploration.

Table 3. Utilization of geophysical methods in geothermal exploration.

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

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

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

TABLE 3

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

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