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SUBREGIONAL AND DETAILED EXPLORATION FOR
GEOTHERMAL-HYDROTHERMAL RESOURCES

N.E. Goldstein

May 1986

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SUBREGIONAL AND DETAILED EXPLORATION FOR GEOTHERMAL-HYDROTHERMAL RESOURCES

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May 1986

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INTRODUCTION

Whether by reason of fortuitous circumstance or careful planning, many regional geothermal-exploration activities routinely enter the subregional exploration phase, defined here as the stage between regional assessment and deep exploration/production drilling, logging, and well testing. The subregional stage begins once the regional exploration area has been reduced to one or more subareas and the commitment is made to focus activities on the search for drill targets. Because the geological risk could be very high at this stage, a manager might follow a simple phased program to gain maximum information for minimum time and money. In such a phased program, the initial objective may only be to determine whether the area warrants one or more deep exploratory holes, and if so, where and how deep to drill. If there are strong surface manifestations of a shallow heat source, these questions may be answered simply by the results of geological mapping and geochemical analyses of the thermal effluents. More often, however, surface manifestations may be weak and/or provide inadequate information for a deep drill hole.

There is no universally applicable formula for planning and conducting subregional exploration. Each project differs in geological-hydrogeological setting, physical characteristics (e.g., terrain, accessibility, rainfall, vegetation), and in the local availability/applicability of equipment and laboratory facilities. Furthermore, the results from existing geological, geochemical, and geophysical surveys may dictate a special course of action. On the basis of past personal experiences or case-history information, a manager may rely initially on a familiar, conservative exploration approach, avoiding newer or innovative techniques altogether or relegating them to a later phase. In this chapter we assume no exploration strategy; rather we discuss the basis for most of the more widely used methods and techniques, plus some lesser-known ones. We give examples of how each technique has been effective in certain situations. We also describe the limitations of the techniques and where they might not prove effective or give ambiguous, misleading information.

EXPLORATION TECHNIQUE OVERVIEW

Serious geothermal-exploration activities in the United States began in the late 1960s and early 1970s, somewhat later than in Italy, New Zealand, and France. Although the techniques used initially were influenced strongly by technology borrowed from the mining and petroleum industries, technology and strategies specific to geothermal reservoirs evolved rapidly in the U.S. because of a unique combination of circumstances and conditions. In contrast to conditions in other countries, prospective geothermal areas in the U.S. generally lacked strong surface manifestations, yet many of the areas were easily accessible. This situation contributed to the willingness of many geothermal developers to try any technique, however speculative. Furthermore, the easy availability of drills and experienced drillers encouraged more and earlier drilling of shallow- and intermediate-depth holes for stratigraphic and temperature information than was customarily done in other countries. One consequence of this approach is that it led to several disappointing exploration attempts and an erosion of confidence in geothermal-exploration methodologies. However, government-supported research programs through the Departments of Energy and Interior, along with the participation of private geothermal developers, have produced improved techniques, equipment, and interpretational methods, together with a better appreciation of their limitations.

Reviews of published papers and knowledge of specific exploration work seem to indicate that most practitioners around the world tend to rely initially on a basic set of techniques grouped in Table 1 according to the major discipline. Hydrogeology cross-cuts the three disciplines. With the exception of the shallow-to-moderate-depth temperature surveys, which require drilling equipment, the techniques in Table 1 are all relatively low cost and can be conducted by small field crews using fairly simple conventional instruments, available almost everywhere. Prior to the deep-drilling phase, additional subsurface information may be gained from the application of various techniques listed in Table 2 and discussed in this report.

This report is subdivided into three main sections, covering geological, geochemical, and geophysical techniques. This order corresponds to the order of increasing section length and detail of information; it also no doubt reflects the fact that the author is a geophysicist with an ill-concealed professional bias. Moreover, these section lengths are proportional to current exploration costs and levels of effort. Geophysical costs typically far exceed the costs of both geochemical and geological work; and depending on how one differentiates geochemical techniques, geochemical costs often exceed geological costs. While it is true that geophysics is expensive because of field crew size, cost, and complexity of both field equipment and data interpretation, it can also be argued that most exploration efforts will stress geophysical data acquisition because the data, if properly interpreted, can be extremely helpful in providing the subsurface information needed for decisions on where and how deep to drill.

TABLE 1 Basic Geothermal-Exploration Techniques	
Discipline	Techniques
Geology	• Surface and photogeologic mapping - Volcanic stratigraphy - Structure - Hydrothermal alteration
Geochemistry	• Sampling and analysis of surface discharges • Analysis of superficial deposits • Field measurements of - Temperature - pH - Flow rates of discharges • Geochemical geothermometry
Geophysics	• Shallow-to-moderate-depth drill holes for temperature surveys • Magnetic, gravity, and dc-resistivity surveys

TABLE 2
Supplemental Geothermal-Exploration Techniques

Discipline	Techniques
Geology	• Volcanic and plutonic geochronology • Paleomagnetism
Geochemistry	• Stable-isotope studies • Subsurface alteration and trace-metal distribution • Mixing and boiling models
Geophysics	• In-field heat-flux measurements • Magnetotelluric soundings • Controlled-source electromagnetic soundings • Self potential • Passive and active seismic surveys

GEOLOGICAL TECHNIQUES

The subregional geological assessment is concerned mainly with assembling a preliminary data base to guide geochemical, geophysical, and drilling activities. As the best guides to a geothermal reservoir and a heat source are surface thermal manifestations and the age and distribution of volcanic rocks, the geological work normally concentrates on those features. However, where thermal manifestations are weak, the geologist will attempt to draw inferences from the zonation of hydrothermal alteration, age relations between hydrothermal alteration and faults, and possible structural-stratigraphic and hydrogeological relations between faults, permeable rock units, and hot or warm springs in the area.

Geological data are usually compiled on maps of 1:24,000 or 1:62,500 scale, and the detailed data are later recompiled at a scale of 1:1000 to 1:5000. If available, air-photo imagery, such as black and white, color, or color-infrared photographs, supplement the field and library information. By combining local geology with regional geology and hydrology, the geologist can assess the probable locations of subsurface thermal zones and determine whether cold meteoric waters recharge and/or mask the system.

IGNEOUS-VOLCANIC ROCK ASSOCIATIONS

Smith and Shaw (1973, 1975) have laid a framework for the relationship of igneous rocks, volcano evolution, and geothermal potential that geologists have found to be extremely useful. Simply stated, the hypothesis of Smith and Shaw holds that magmas producing basic volcanics (basalts and andesites) form in the deep crust/upper mantle and, because of their low viscosity, ascend rapidly to the surface to form narrow dikes and small pipes. Individual basic magma pulses are volumetrically small, contribute little stored heat to the upper crust, and therefore rarely produce thermal anomalies of economic importance. On the other hand, continued basaltic underplating of the crust (Lachenbruch and Sass, 1978) can result in high-level magma chambers. Depending on the composition of the crustal rocks, the degree of host rock assimilation, magma mixing, and crystal fractionation within the chamber, these magma chambers may yield lavas of varying composition. Sustained underplating of basaltic magma beneath the chamber produces long-lived magmatic-hydrothermal systems with large geothermal potential (Eichelberger and Gooley, 1977). These systems are characterized by spatially and temporally coherent distributions of volcanic vents that produce episodes of cogenetic eruptive rocks varying from basalts through dacites to rhyolites, e.g., the Newberry volcano, Oregon (Higgins, 1973). Less common is the association of geothermal areas containing igneous rocks with a bimodal distribution of basalt and rhyolite, such as the youngest eruptions at the Coso volcanic field (Duffield et al., 1980), the Salton Sea geothermal field (Robinson and Elders, 1976), and the Medicine Lake volcano (Heiken, 1978), all in California. These systems have been characterized as extensional zones yielding basaltic ($\text{SiO}_2 < 55\%$) and rhyolitic ($\text{SiO}_2 > 70\%$) magmas from the fractional melting of a mantle peridotite. The cogenetic magmas appear uncontaminated by assimilated crustal host rocks. The frequency of volcanic episodes and the volume of eruptive rocks are believed to be closely linked with the rate of crustal extension, which results in the "bleeding off" of magma from a deeper chamber (Bacon, 1982).

The relationships between magma evolution, volcanism, and the rate of crustal extension have been described by Hildreth (1981) and illustrated, in part, by Figure 1. Under conditions of rapid crustal extension, but low thermal input (basaltic dike injections) from the mantle, rhyolitic domes and peripheral basaltic lava flows develop (Figure 1A). A large

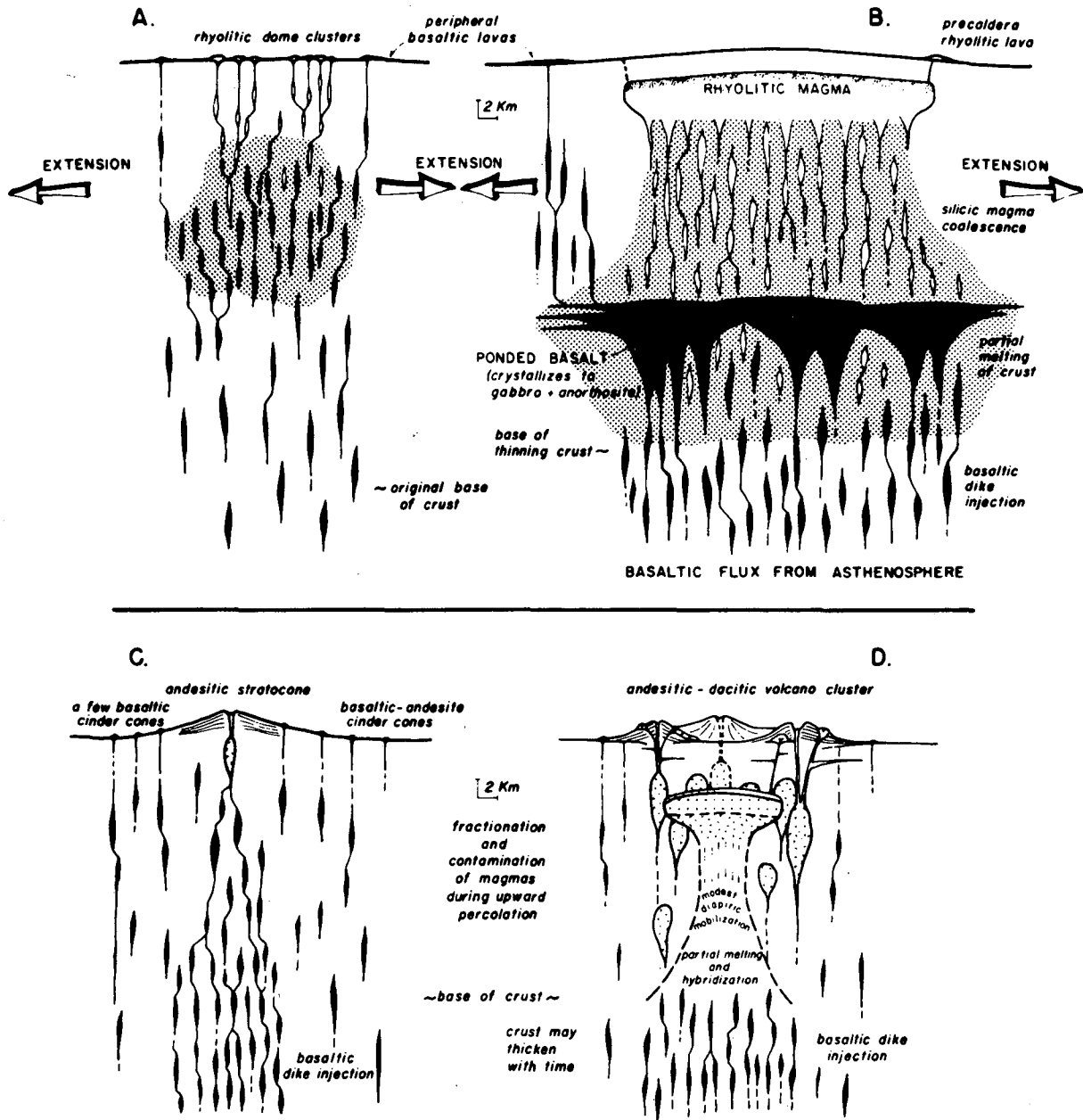


Figure 1.

Some contrasting styles of lithospheric magmatism. The two upper panels depict rhyolitic-basaltic magmatism under stress conditions favoring marked crustal extension: (A) modest power input; (B) large power input, advanced stage. Shaded regions indicate partial melting of crustal rocks sufficient to permit separation of rhyolitic magmas as gash veins and dikes. The two lower panels depict possible stages in development of volcanic systems where tectonic extension, if any, is subordinate and shallow: (C) early stage; (D) intermediate stage. This model applies to island arc, continental margin arc, and continental interior systems that produce abundant intermediate magmas. All four sketches are idealized and refer to no particular systems. The models are independent of the mode or site of generation of basaltic magma, but basalt is thought to provide the power supply for virtually all other magmatics (from Hildreth, 1981).
(XBL 849-3860)

amount of thermal flux eventually produces a large, zoned rhyolitic magma chamber (Figure 1B) and a silicic caldera, as discussed in the next paragraph. Subordinate crustal extension may result in basaltic cinder cones and flows around an andesitic stratocone (Figure 1C). In time, and given continued heat input at the base of the crust, the simple system in 1C may evolve into a cluster of overlapping andesitic-dacitic volcanoes and satellite vents fed from several discrete melt zones (Figure 1D).

Where the build-up of volatiles in the hood zone of a large and highly evolved chamber has led to voluminous silicic ash-flow eruptions, such as the Long Valley caldera, California (Bailey et al., 1976), subsidence and the formation of a caldera structure have outlined the location of the original chamber, the dashed line in Figure 2. Caldera formation followed the eruption of the Bishop Tuff (500 km³) 0.7 m.y. Subsequent eruptions record a progressive evolution and solidification of the underlying zoned chamber. This is evidenced by the trend toward more mafic (i.e., less quartz and potassium feldspar) and crystal-rich eruptions, from the early rhyolites associated with the resurgent dome, through the rhyolites that erupted around the resurgent dome (0.5 to 0.1 m.y.B.P.), the rim rhyodacites (0.2 to 0.05 m.y.B.P.), and finally the late basaltic lavas in the west and south moat area (0.2 to 0.06 m.y.B.P.). The Long Valley caldera is in the waning stage of development, but there is evidence that magma remains at a depth of 6 km or more. In systems such as The Geysers geothermal field, California (Hearn et al., 1981) and The Coso volcanic field (Duffield and Bacon, 1980), where there has been sustained tectonic activity and frequent magma leakage, we may observe only a widely distributed set of volcanic cinder cones, domes, and flows whose compositions and distribution bear evidence of the evolution and location of the one or more parental magma chambers that had been or may be present. Prominent volcanic edifices, such as the composite andesitic stratovolcanoes of the High Cascade Range in Oregon and Washington, stand out as obvious areas for exploration. Despite their size and the youthfulness of the eruptions—700,000 years to present (Williams et al., 1982)—they have not yet yielded evidence for an exploitable, high-temperature reservoir. In contrast to the Holocene volcanoes in Central America, for example, which also constitute part of the Circum-Pacific belt of andesite-rhyodacite volcanoes, the High Cascade volcanoes are much smaller in volume and lower in rate of volcanism. The lack of a major discovery beneath the flanks of these volcanoes may be due to several factors: the magma chambers are small and have evolved at depths greater than 8 to 10 km (Smith and Shaw, 1975; Blackwell and Steele, 1983), ascending magmas either freeze or erupt without imparting much stored thermal energy to the rock beneath the edifices, and volcanism is in the waning stage of a depleting mantle source of magma (McBirney, 1978). Figure 3 is a conceptual model (after Henley and Ellis, 1983) for a geothermal system associated with an andesite-dacite composite volcano, typical of active island arcs and the High Cascade Range. The heat source is a cooling neck-type conduit and subvolcanic stock. Isotherms are depressed on the high rainfall side of the volcano, but shallow, hot water aquifers in permeable pyroclastic units may occur beneath the flanks of the rain-shadow side. In addition, high-temperature resources may exist in permeable older volcanic and pre-volcanic units close to the central vent or in association with major graben faults.

Young volcanic rocks are not necessarily evident in the geothermal areas within the Basin and Range Province of the western U.S. The systems seem to be mainly related to deep fluid convection along fault zones within an area of hotter crust; the crust is heated by basaltic magmas that rarely break the surface.

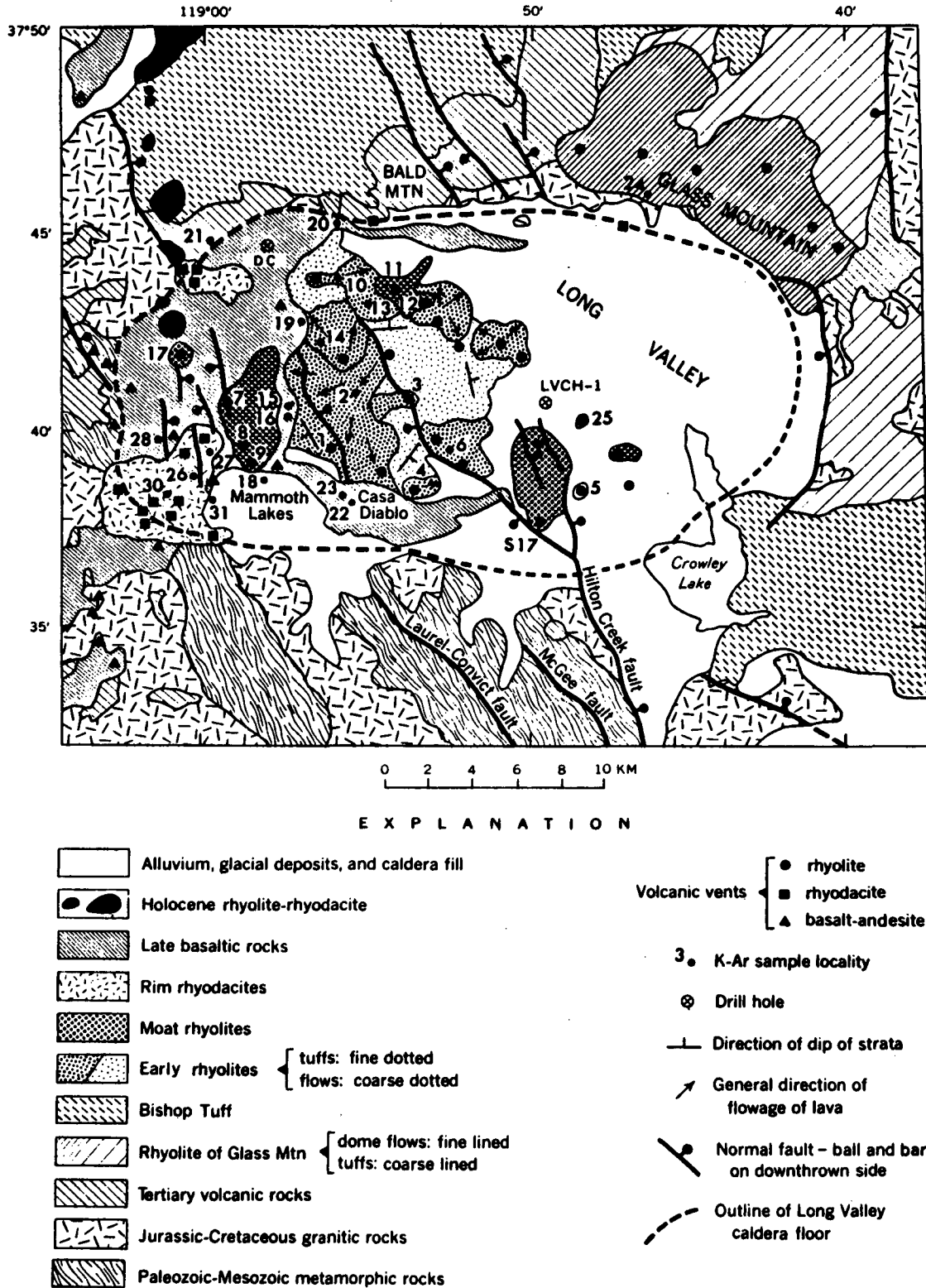


Figure 2. Generalized geologic map of Long Valley caldera (from Bailey et al., 1976). (XBL 849-3862)

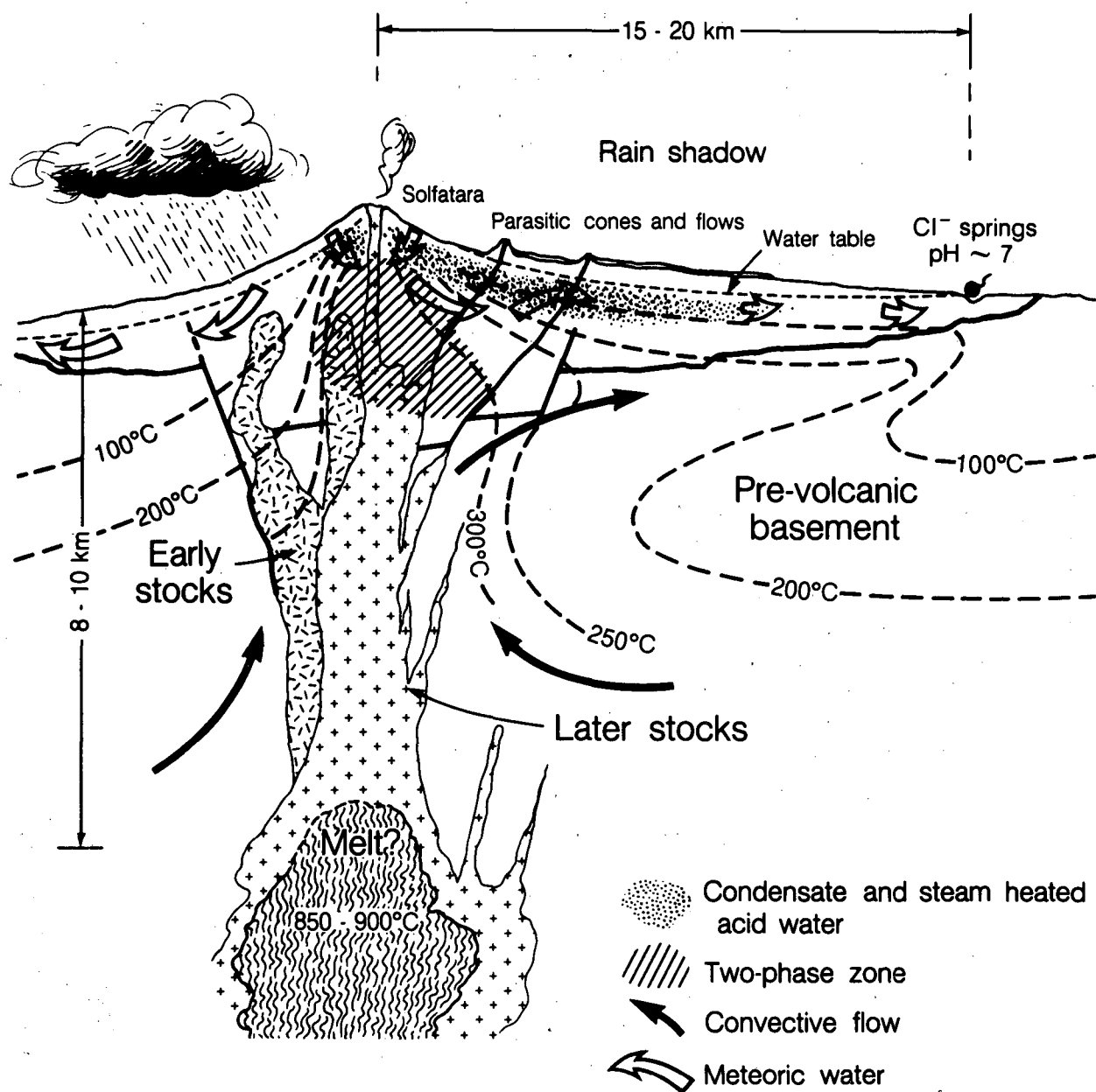


Figure 3. General schema of a possible hydrothermal-geothermal system associated with an andesite-dacite composite volcano typical of active island arcs (after Henley and Ellis, 1983). (XBL 8411-6164)

STRATIGRAPHIC AND STRUCTURAL INTERPRETATION

A combined literature search and field survey is often done to determine the stratigraphic relations between sedimentary and volcanic units and to assess the primary porosities and permeabilities of the units and the local hydrology. In contrast to the typical oil or gas reservoir, geothermal resources are most often encountered in rocks with low matrix permeability, such as crystalline rocks and metamorphosed sediments. Experience at several geothermal fields has shown that reservoir fluids are produced in important volumes only where a well has intersected narrow and infrequent zones containing fractures, usually sub-vertical, that are both open and part of a system that is well connected hydraulically. There are also cases where secondary porosity, due to the hydrothermal dissolution of quartz grains, is an important factor (Lippmann and Mañón, 1985).

The search for aquifers and potential reservoir rocks by studying volcanic structures and primary porosities has been particularly successful in Iceland, where uniformly high sub-surface temperatures diminish the usefulness of simple temperature surveys for targeting drill holes (Fridleifsson, 1979). Flow channels, dikes, and permeable faults forming the plumbing system are mapped in detail. From these and other careful studies of volcanic stratigraphy, pillow lavas were found to have higher permeability than other major rock units in the geothermal area, and those with olivine tholeiite composition are better reservoir rocks because of their larger primary porosity (Fridleifsson, 1979).

The mapping of faults and subsidiary fractures associated with faults and folds (Stearns and Friedman, 1972) and the mapping and dating of volcanic eruptions is fundamental to developing an idea of local stress conditions and the orientation and location of subsurface fractures. Careful mapping and three-dimensional (3-D) fault analysis was partially successful at the Redondo Canyon area of the Valles Caldera, New Mexico. Unfortunately, many of the major faults of the medial graben there were found to be sealed by hydrothermal minerals at depth (Hulen and Nielson, 1982), a fact which could not have been predicted from surface observations. On the positive side, there is evidence from the Larderello geothermal field, Italy (Gianelli et al., 1978), The Geysers geothermal field, California (McLaughlin, 1981), and Coso volcanic field, California (Brophy, 1984), that structural interpretations based on surface mapping, aided perhaps by remote sensing, can indicate where highly fractured rocks are more likely to occur. In all three areas productive fractures have been intersected by drilling near the crests of anticlinal folds or in horst blocks, presumably because extensional near-surface horizontal stress keeps fractures open to an appreciable depth. On the other hand, down-dropped graben blocks in calderas and beneath stratovolcanoes might typically be impermeable environments.

Recent studies by many workers on the origin and location of pull-apart basins show that these features develop within long strike-slip boundaries between rigid continental plates. Active pull-apart basins, such as those of the Salton trough, are also the loci for complex faulting, volcanism, and earthquake activity (Robinson and Elders, 1976; Sharman et al., 1976). High sedimentation rates can mask the thermal effects and make these basins difficult to detect (Mann et al., 1983). However, in areas of active deformation, photogeology and surface mapping may reveal the existence of features oblique to the master strike-slip faults. Exploration and development at the Cerro Prieto geothermal field, Baja California, during the 1970s was guided in part by this concept (de la Peña et al., 1979), which was later confirmed by geophysics and drilling (Vonder Haar and Howard, 1981; Lyons and van de Kamp, 1979).

Many geothermal systems are not so well concealed by sediments as those in the Salton trough. The neovolcanic zone of northern Iceland is a prime example of an oceanic rift zone where faults and fissuring associated with east-west extension, hot water emanations, and volcanic eruptions are clear (Björnsson et al., 1979). Less obvious are the relationships between faults and thermal features in the Basin and Range Province. Even though erosion and sedimentation during the last 10 million years have concealed all but the youngest earthquake faults, geologists have found that many of these geothermal systems occur at the intersections of major normal faults parallel to the ranges and older high-angle faults oblique to the ranges (Hose and Taylor, 1974; Beyer et al., 1976; Ross et al., 1982; Hulen, 1983). Some Basin and Range fault systems not only provide the permeable channel for fluid discharges, but also may produce part of the reservoir region, such as the permeable dilation breccia found by drilling at the Beowawe geothermal area, Nevada (Sibbett, 1983). The most thoroughly studied and documented structurally controlled geothermal system in the Basin and Range Province is at Roosevelt Hot Springs, Utah (Nielson et al., 1978, 1979; Ross et al., 1982). Tertiary plutonic and Precambrian metamorphic host rocks at the Roosevelt Hot Springs have very low primary permeability. The system is believed to be controlled by a north-northeast-trending set of young faults and fractures and their intersections with older high- and low-angle normal faults.

Complex faulting, some of which is undoubtedly important in providing fluid-flow paths, has been mapped at several young silicic calderas, such as the Valles Caldera, New Mexico (Smith and Bailey, 1968), and Long Valley caldera (Bailey et al., 1976), and at other volcanic centers, such as the Coso volcanic field (Duffield and Bacon, 1980) and the Medicine Lake volcano, California (Ciancanelli, 1983). Figure 4 shows the system of faults and lineaments mapped by Ciancanelli (1983) at the Medicine Lake volcano, a Quaternary bimodal shield volcano in the Cascade Range in northern California. A predominant set of north-south normal faults are mapped, and those that align with the extrusive vents are believed to be related to the eruption of voluminous rhyolitic lavas. Fink and Pollard (1983) believe that some of the faults may represent surface deformation above dikes that came to within only 100 m of the surface 1000 years ago, estimated on the basis of ^{14}C dating.

AIRBORNE REMOTE SENSING

Among the modern geological tools are various ground-imaging techniques known collectively as "remote-sensing" techniques because the data, electromagnetic in nature, are acquired at aircraft or satellite elevations. Because these methods provide rapid data acquisition over large areas, their primary use is in regional assessment. However, if these data are available, they should not be ignored during the subregional exploration phase. For this reason, a brief discussion of a few methods is appropriate here, even though these methods do not seem to be widely used in geothermal exploration. Except for Landsat imagery, available from the U.S. Geological Survey, and aerial photography shot with hand-held 35-mm cameras, data-acquisition costs can be high. As data processing and interpretation techniques have improved, the value of remote-sensing techniques has grown.

Photographic Techniques

High- and low-sun-angle black-and-white aerial photographs taken at different times of the year have been used to help map faults and to identify fracture systems. Radial, circular, and linear discontinuities visible in the photographs may be related to concealed recent dikes and larger plutons (Fink and Pollard, 1983). Anomalous patterns in snowmelt and vegetation type and vigor may also be useful indications of hydrothermal conditions along

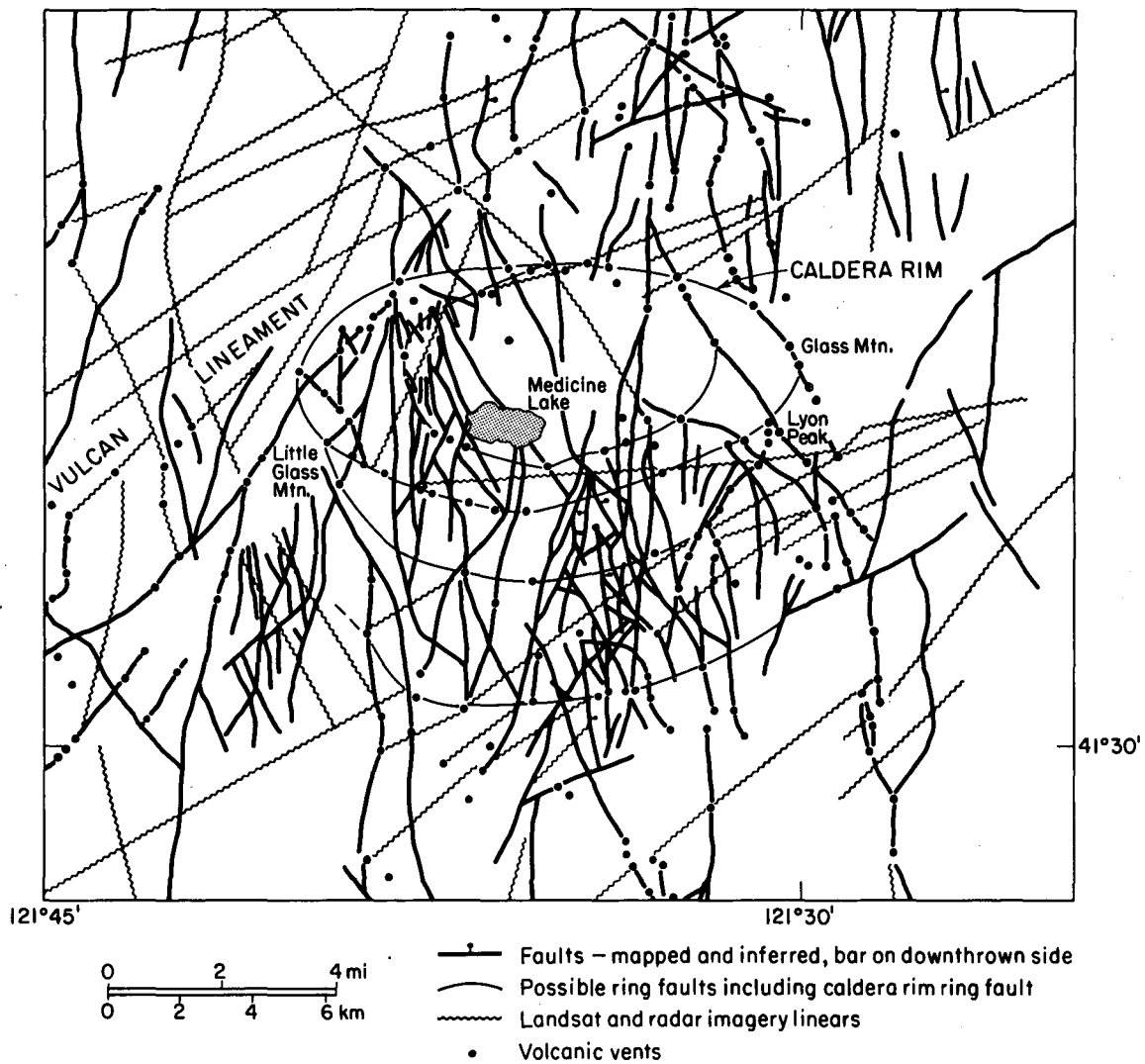


Figure 4. Faults and lineaments at the Medicine Lake volcano, in the Cascade Range, northern California (from Ciancanelli, 1983). (XBL 841-9516)

deep-seated zones of fractures. True-color aerial photographs may reveal the type and extent of hydrothermal alteration and the spatial relations between various volcanic units and other rocks. True-color aerial photographs show surface features similar to that perceived by the human eye. However, over vegetation most visible wavelengths are strongly absorbed by photosynthesis; only green and yellow are weakly reflected, and true-color photographs give limited information (Goetz et al., 1983). False-color infrared (CIR) film is more sensitive to the narrow bandwidth (0.75-0.9 μm) where vegetation is most reflective. For this reason, CIR photography is often used in geobotanical investigations and can be used with true-color photography to infer the state of health of vegetation. This technique could help identify areas of vegetation that are affected by recent thermal discharges (e.g., geochemical stresses induced by certain heavy metals in the soil, such as Ca^{+2} , Mg^{+2} , Cu^{+2} , and other bivalent cations) (Hewitt, 1963).

Thermal Infrared Imagery

Thermal infrared (IR) imagery gives a direct indication of thermal manifestations and can be extremely useful in sparsely populated and poorly accessible areas. Commercial Bendix or Daedalus scanners examine a swath of ground whose width is proportional to a 120° arc beneath the aircraft. Detectors are sensitive to thermal radiation in the bands where IR is not absorbed by the atmosphere, wavelengths of 3.0 to 5.5 μm and 8.0 to 14.0 μm . This technique can resolve thermal effects as small as a few meters in diameter (McNitt, 1976). Conventional IR scanners record surface temperature differences of 1 to 3°C , and are used qualitatively to map surface temperature anomalies of geothermal origin such as fumaroles and hot springs. An advanced airborne method, described by del Grande (1982), has the potential for resolving surface temperature variations of 0.24°C by measuring the two IR bands simultaneously at two different altitudes. The practical value of this method is that the corrected temperature maps, used together with temperatures in a series of shallow holes, may provide a more accurate 3-D picture of hot water in shallow aquifers. Airborne thermal IR surveys have seen limited use in the western U.S., probably because most areas with thermal discharges are reported in the literature. However, over the Black Rock Desert, Nevada, thermal IR imagery detected numerous hot springs, one only 1 m in diameter, and standing pools of hot water. Many springs were previously unreported in the literature (G.V. Keller, personal communication, 1978). When used after a volcanic eruption or major earthquake activity, thermal IR imagery could help determine changes in the location and intensity of thermal waters reaching the surface.

Landsat and Airborne Multispectral Scanning

Linear and arcuate patterns caused by faulting and possible concealed igneous intrusions may be obtained from the Landsat satellite multispectral scanner (MSS) data. Although most of the work has been applied to mineral exploration, MSS data have been used to map large areas of hydrothermal alteration. MSS data, acquired in four spectral bands (band 4 at 0.5 to 0.6 μm , band 5 at 0.6 to 0.7 μm , band 6 at 0.7 to 0.8 μm , and band 7 at 0.8 to 1.1 μm), have a spatial resolution (the pixel size) of about 80×80 m. Because there is usually a high degree of correlation between spectral bands, an examination of the raw data from all four bands or color composites made from any three of the raw bands does not provide much more information than can be obtained from a visual examination of one or two bands. To overcome this limitation, a numerical linear transformation can be made using a technique called the "principal components processor" (Anuta, 1977), which reduces the magnitude of the correlations among the four MSS bands and thereby

emphasizes minor trends in the data. In effect, the data are decorrelated (Siegal and Gillespie, 1980), allowing minor geological trends to show up better as subtle color variations when any three of the four enhanced data sets are recombined to form a blue, green, and red composite. In one application, processed MSS data delineated areas of hydrothermal iron oxides in Arizona porphyry copper districts. Color anomalies in shades of red and orange were found after ratios of bands 4/5, 5/6, and 6/7 were reproduced as blue, green, and red, respectively, and then superimposed (Abrams et al., 1983).

The limited spatial resolution of the Landsat MSS has been a problem, but the fourth Landsat, launched in 1982, carried a seven-channel scanner called the thematic mapper (TM), which provides better spatial resolution. Channels 5 and 6 of the TM are sensitive to wavelengths longer than 1 μm and are thus sensitive to the reflectance from hydrous, hence hydrothermally altered, minerals. An aircraft-borne, 24-channel MSS was flown over the Marysvale, Utah, mining area to assess the detectability of secondary argillic minerals, such as alunite and kaolinite, which are derived from reactions with acidic hydrothermal fluids (Podwysocki et al., 1983). Areas of abundant alunite and kaolinite, two minerals found over some geothermal systems, were identified by the intense absorption in the 2.17- to 2.22- μm band, as depicted in color composite images using various band ratios chosen to emphasize the spectral contrasts that exist between argillic versus nonargillic rocks. The patterns of intense alteration were interpreted as the remnant of a paleohydrothermal convection system that was produced by the emplacement of quartz monzonite stocks 23 million years ago.

While airborne multispectral techniques have not yet been applied to active geothermal areas, high-resolution absorption and reflectance spectral techniques may have applications in areas where vegetation obscures geological features. Reflectance from vegetation may be modified by changes in the internal structure of leaves, which may be affected by metal concentration, thus producing geobotanical anomalies (Labovitz et al., 1983). The distribution of plant communities may also be related to geological variations, including hydrothermal alteration (Milton, 1983).

FIELD INVESTIGATIONS

Once in the field, the geological team may first locate all known or possible thermal anomalies, confirming in that process anomalous features detected in data obtained by remote-sensing techniques. Ideally, the team should have a geochemist and a hydrologist to sample and study hot and cold waters in the area and to perform simple field measurements and calculations. This aspect is discussed in the next section. In less developed countries, the thermal discharges may be hard to find from existing maps and reports, and the prospecting team may have to question the local population (McNitt, 1973).

After locating the thermal manifestations, the geologist will sample and map the area, paying particular attention to the distribution of mineralized zones and old prospects, hot-spring deposits, type and extent of rock alteration, igneous-volcanic outcrops, possible faults and fracture patterns. Where possible, the geologist will measure the dimensions of a hot-spring deposit, from which its mass is estimated. When this is compared to the present rate at which deposition occurs at the discharge areas, one obtains a crude estimate of the age of the discharge system.

Studying the hot-spring deposits and rock alterations in conjunction with geochemical indicators provides temperature information on the system. Sampling and mapping igneous and volcanic rock units and hydrothermal alterations provide information on the age and evolution of the magmatic-volcanic system. Of particular concern are the ages and

distribution of volcanic rocks and alteration patterns, as these should be closely related to the hydrothermal-geothermal reservoir sought. Volcanic-igneous rock samples are collected for K-Ar age dating, a technique that is most generally regarded suitable for rocks older than 40,000 years. In areas of multiple thermal episodes, some rocks may yield radiometric ages that are too young. In these instances, the K-Ar age will reflect the age of the last thermal event that released Ar and reset the radiometric clock operating on the decay of ^{40}K to ^{40}Ar . K-Ar dating of potassic hydrothermal minerals such as adularia and sericite may be useful for distinguishing between multiple-alteration events and for relating the radioactive ages of igneous and host rocks to the ages of the alteration (Silberman and White, 1975). For recent thermal events, ^{14}C dating is sometimes helpful if charred vegetation can be found in ash-fall tuffs (Miller, 1985). This technique is generally limited to ages of less than 25,000 years. Improved laboratory techniques are reducing the gap between K-Ar and ^{14}C age dating. Other age-dating techniques are discussed in the section on Geochemical Techniques.

Hydrothermal Alteration As a Guide to Subsurface Conditions

The interaction between the circulating thermal fluids results in water-rock reactions that produce assemblages of secondary (hydrothermal) minerals that are often used as a guide to present or past subsurface temperatures. Permeability and porosity of the host rock strongly influence the water-to-rock volumetric ratio and the degree to which the original minerals are altered. The suite of secondary minerals is less controlled by rock type than by temperature and composition of the fluids (Browne, 1978). Several geothermal fields have been intensively mapped and drilled, and the cores and cuttings have been studied in enough detail to show that the hydrothermal minerals are zoned, sometimes crudely, with temperature in a manner similar to that observed in the host rocks enclosing hydrothermal ore deposits (Holland, 1967; Meyer and Hemley, 1967).

The typical hydrothermal alteration found in most geothermal systems is the propylitic type, which consists of several distinct mineral assemblages. In order of increasing depth and temperature these are as follows: (a) clay-zeolite, (b) calcite-mixed clays-chlorite, and (c) chlorite-epidote. Permeable zones may be capped by a blanket or patches of bleached rocks characteristic of acid-sulfate conditions that cause argillic alteration. Near-surface boiling, the release of H_2S into the gas phase and the influx of oxygenated meteoric water combine to yield abundant H^+ and an alteration assemblage consisting of aluminosilicates such as alunite (a sulfate), low-temperature K-mica, and kaolin, and by fine-grained pyrite and several varieties of silica such as cryptocrystalline quartz and cristobalite. Argillic alteration is associated with acid-sulfate springs and fumaroles at a number of geothermal fields such as at Valles Caldera, New Mexico (Hulen and Nielson, 1986) and is often associated with shallow precious-metal deposits in volcanics. Ore-grade gold-silver mineralization is believed to be depositing beneath the hydrothermal eruption craters of the Waiatapu geothermal field, New Zealand (Hedenquist and Henley, 1985).

We present two examples of the distribution of hydrothermal minerals, the first associated with a sandstone-siltstone reservoir in deltaic sediments, the second reservoir is in fractured volcanics, primarily andesites. Except for details, the mineral assemblages are similar to those observed in basalts at the Reykjanes area, Iceland (Truesdell, 1976), in rhyolites of Ohaki-Broadlands, New Zealand (Browne and Ellis, 1970), and ignimbrites of the Valles Caldera, New Mexico (Hulen and Nielson, 1986).

A Sandstone Reservoir: Cerro Prieto, Baja California

Several types of hydrothermal mineral zones, gradational with respect to depth, were recognized at Cerro Prieto (Elders et al., 1979) on the basis of cuttings from dozens of wells. The principal reservoir rocks are sandstones or silty sandstones. The simplest zonation pattern is one that shows a regular, progressive sequence of hydrothermal minerals in the sandstone units, as shown in Figure 5. This general pattern has been reported in other active geothermal areas in the Salton trough; at Wairakei, New Zealand; and at Reykjanes, Iceland, although each area produces a slightly different suite of minerals. Geologists at Cerro Prieto are reported to pay particular attention to the minerals in the sandstone cuttings as a guide to subsurface temperatures. The argillaceous rocks (siltstones, shales, mudstones) do not contain the same high temperature mineral assemblages, possibly due to their lower permeability, as the sandstones, but all rocks show reduced porosity due to the secondary minerals.

At depths where temperatures in the range of 175 to 250 °C occur there is a progressive decarbonation. Calcite in cement and vein fillings is destroyed and calc-aluminum silicates are formed; the principal ones are chlorite + epidote + prehnite + actinolite. At the highest temperatures encountered in wells, around 350 °C, hydrothermal biotite and vermiculite form. The calc-aluminum silicate zone at Cerro Prieto has a mineral assemblage similar to that found in greenschist facies rocks found in oceanic spreading centers, and is evidence for hydrothermal circulation of brines through the permeable sandstones. There does not seem to be any relation between the hydrothermal zones and faults or stratigraphic units.

An Andesite-Rhyolite Reservoir: Los Azufres, Mexico

An extensive study of cores and cuttings from around 40 wells has been made to detail the effects of hydrothermal alteration as a function of temperature and depth (Cathlineau et al., 1985). The primary rocks consist of spherulitic, glassy and pumiceous rhyolite tuffs near the surface, andesite flows ranging in texture from aphanitic to porphyritic, and minor basalts and dacites. The secondary mineralization is compatible with prograde metamorphism of Ca-Fe-Mg rich rocks. Signs of Na-K metamorphism, such as observed where geothermal brines interact with granites, rhyolites, and sandstones, seems to be absent at Los Azufres. The main mineral assemblages observed in veins and in the altered host rocks are listed below in sequence of increasing temperature. A zonation diagram is shown in Figure 6, and this information is also shown in Table 3 and in Figure 7. The amorphous silica, elemental sulfur, smectite, and alunite observed in the shallowest zone are typical of argillic alteration.

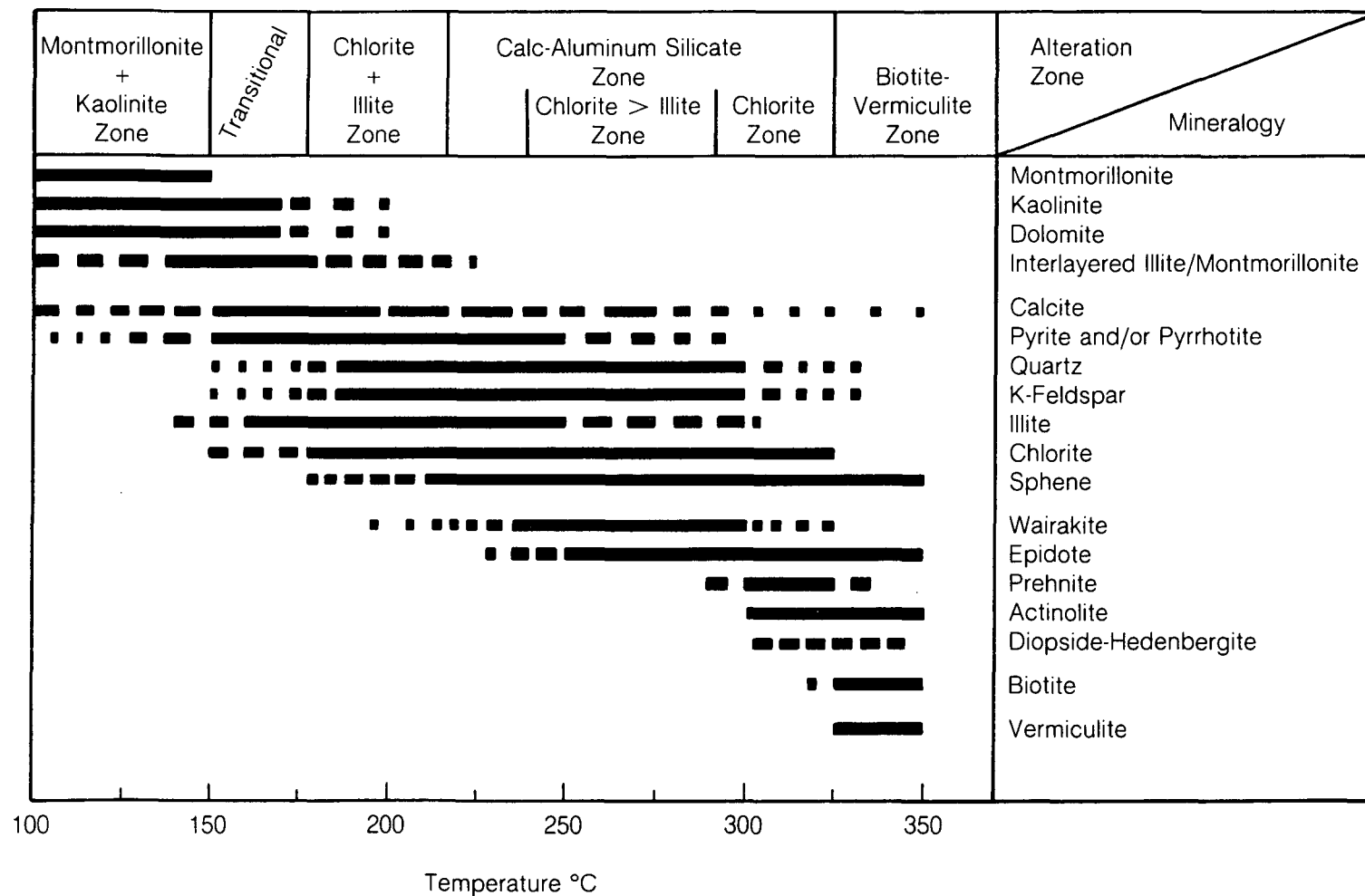


Figure 5. Temperature ranges for the occurrence of hydrothermal minerals in the sandstones of the Cerro Prieto geothermal field, Baja California (after Elders et al., 1979).
(XBL 863-10707)

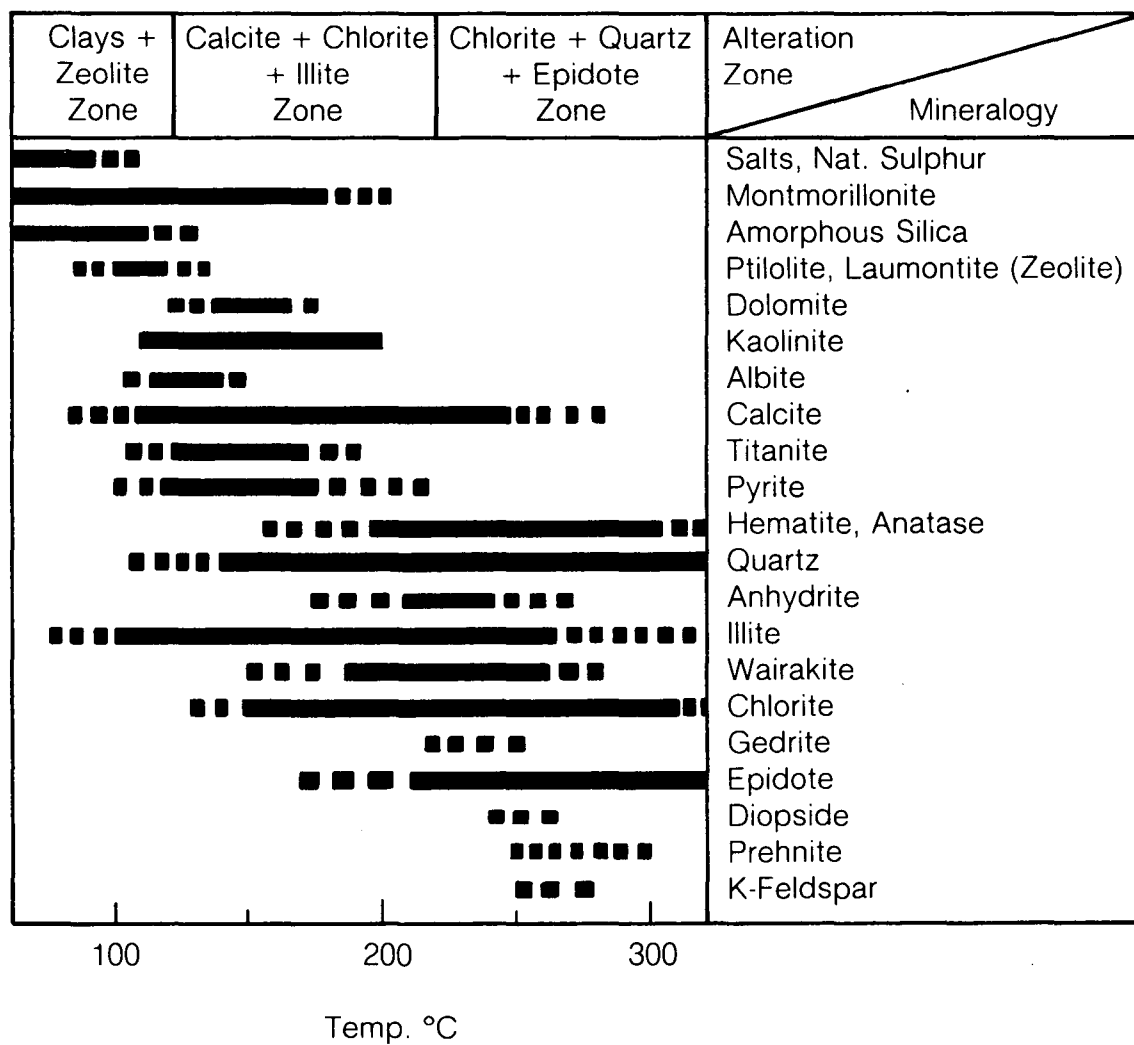


Figure 6. Temperature ranges for the occurrence of hydrothermal minerals in the andesites of the Los Azufres geothermal field, Michoacan (after Cathlineau et al., 1985).
(XBL 863-10706)

TABLE 3
Zonation of Hydrothermal Mineral Assemblages, Los Azufres, Mexico

Zone	Depth-Temperature	Mineralogy
Clay-Zeolite	Surface to 500 m and 100 °C	- amorphous silica + S + smectites + gypsum + alunite - smectites + SiO ₂ + Ca zeolites
Calcite Zone	500 m and 100 °C to ~ 1700 m ~ 215 °C	- calcite + chlorite + sphene + albite + pyrite - calcite + wairakite + SiO ₂ + chlorite + anhydrite
Epidote Zone	> 1500 m 210-300 °C > 2000 m 210-300 °C	- chlorite + epidote + SiO ₂ + hematite + anatase - epidote + amphibole (gedrite) + chlorite - quartz + microcline + prehnite + epidote - epidote + pyroxene (diopside) + SiO ₂

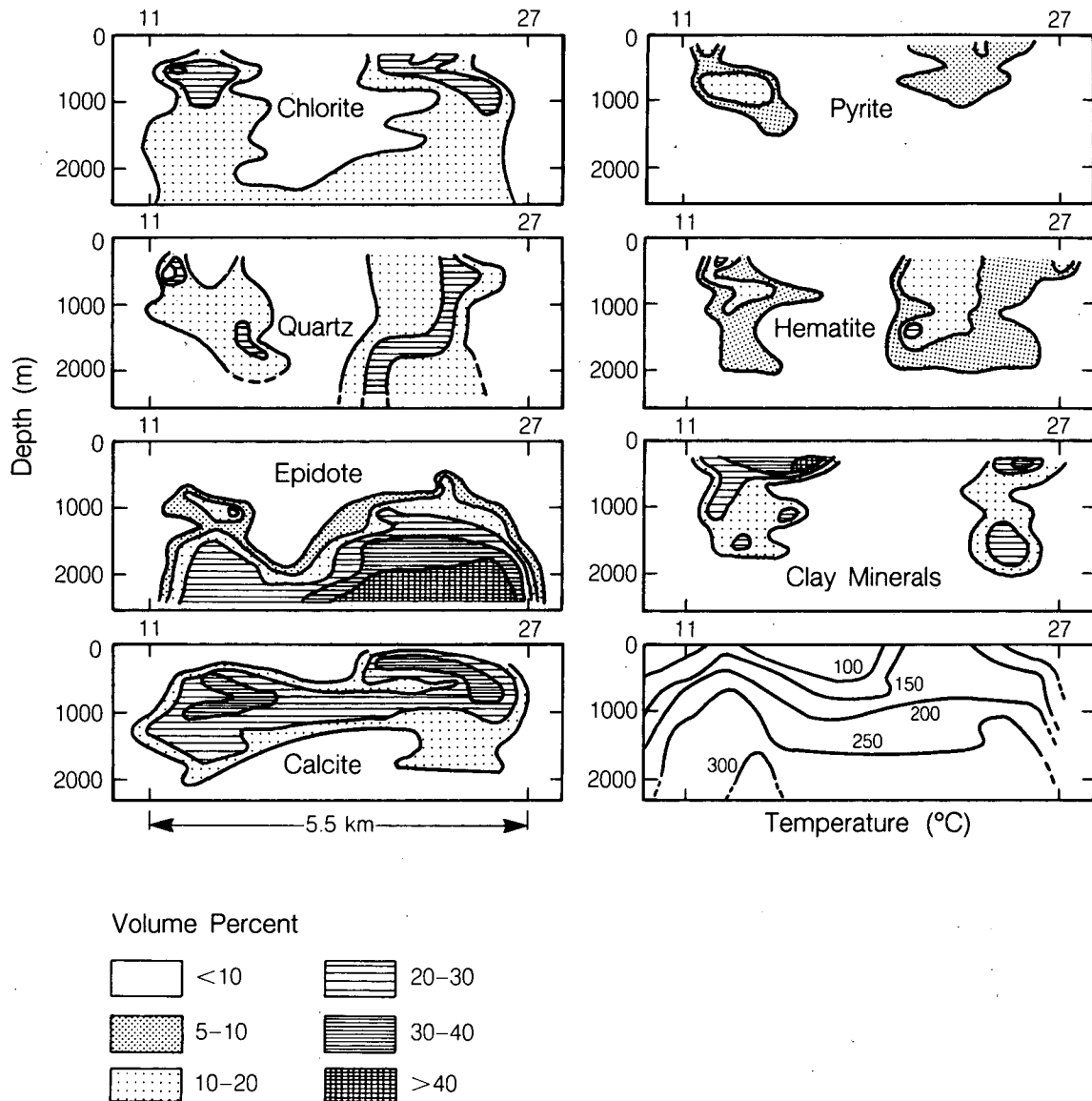


Figure 7. Distribution and concentration of hydrothermal minerals in the Los Azufres geothermal field (after Cathlineau et al., 1985). The cross-section extends from the Tejamaniles zone (well 11) on the south through the Maritaro zone and well 27 on the north. Horizontal and vertical scales are equal. (XBL 863-10705)

GEOCHEMICAL TECHNIQUES

Geochemical methods are used widely in both regional assessment and during the more detailed stages of exploration (Henley et al., 1984). Because of the relatively low cost of surface sampling and chemical analysis compared to most geophysical methods and drilling, geochemical techniques are used where possible to obtain information on present or past subsurface conditions. For example, a great deal of effort has been directed to the study and classification of naturally discharging geothermal waters, tabulating their constituents, and developing methods based on temperature-dependent equilibria for estimating subsurface temperatures. When integrated with information gained during regional assessment and from temperature gradient holes, detailed geochemical and supporting geological investigations are useful for planning and interpreting geophysical surveys and for providing information on subsurface conditions such as:

- (a) boiling or two-phase conditions at depth,
- (b) reservoir temperature and pressure, and
- (c) fluid origin, its circulation (migration) paths, and its dilution by meteoric waters.

If drill hole cores, cuttings and fluid samples are available, geochemical investigations can also provide information regarding

- (a) location of high-permeability zones for production and later fluid reinjection,
- (b) subsurface rock lithologies and the effects of water-rock interactions on major and trace elements in rocks and fluids,
- (c) accurate information on the percentages of noncondensable gases, for better reservoir assessment and modeling, and
- (d) age and thermal history of the system.

Geochemical sampling of soils, gases, and surface waters is usually confined to obvious discharge areas. These areas may be directly above or close to the thermal source and the reservoir region. However, in a significant number of cases, the discharge areas and the geochemical anomalies are displaced many kilometers from the reservoir region. The separations occur for reasons of hydrogeology and topography.

Among the more recently published works, a paper by Mahon (1976) presents a good general review on the hydrogeochemistry of geothermal (mainly volcanic) systems. Ellis and Mahon (1977) cover the same subject in more depth on the basis of their extensive experience in New Zealand and elsewhere, and they discuss the chemical nature of hydrothermal-geothermal systems and geochemical prospecting techniques. Fournier (1981, 1982) presents reviews of techniques and results of water geochemistry, specifically on the use of water chemistry to determine underground temperatures on the basis of chemical and isotopic geothermometers and to recognize boiling and mixing relations. Henley et al. (1984) have compiled a thorough tutorial guide to the use of geochemical techniques in geothermal reservoir analysis. Lastly, one cannot overlook Levinson's (1974) comprehensive and practical text dealing with exploration geochemistry; it contains a broad compendium of field and laboratory procedures developed for mineral exploration.

CLASSIFICATION OF WATER DISCHARGES

Central to geochemical exploration is the proper sampling and analysis of waters, gases, and condensates from fumaroles, hot and cold springs, and local surface drainage. Depending on the integrity of the cap rock and the reservoir pressure, reservoir gases and waters may discharge at the surface and carry information on reservoir temperature, as well as clues to the initial composition of the reservoir fluid and the physical and chemical processes (boiling, mixing, water-rock reactions) that have modified fluid chemistry between reservoir and surface.

On the basis of isotopic evidence, it is now generally accepted that local meteoric water is the principal water found within geothermal systems. Juvenile (magmatic) water, if present, has not been recognized (Truesdell, 1976), but some systems contain a significant amount of connate water, such as the partially evaporated sea water in the Cerro Prieto sediments (Truesdell et al., 1981).

On the basis of many chemical analyses of thermal waters and gases made over the years by workers at most of the major geothermal areas, White (1957a, 1957b), White et al. (1963), Ellis and Mahon (1964), Henley and Ellis (1983), and others have grouped hot-spring waters into the following chemical types and have discussed the origins of those discharges.

Sodium Chloride Water

Chloride-rich waters are common and generally representative of large systems of circulating hot water. The waters are nearly neutral ($\text{pH} \sim 7$) because of the H^+ reactions with silicate minerals, but may be slightly alkaline at the surface if boiling occurs. Sodium is the principal cation, and the $\text{Cl}^-/\text{SO}_4^{2-}$ ratio is large. The highest Cl^- concentrations (tens to hundreds of thousands ppm) are in waters from sedimentary rocks, particularly those containing marine and evaporite deposits.

Acid Sulfate-Chloride Water

The acidity of waters that contain substantial SO_4^{2-} in addition to Cl^- is due to oxidation of H_2S gas or less commonly sulfide minerals to SO_4^{2-} by air or circulating oxygenated meteoric water. There is now evidence that similar waters originate in part from solution of volcanic volatiles (SO_2 and HCl) and occur in the deep parts of active hydrothermal systems associated with relatively young andesite volcanism, such as in the volcanic areas of the southwest Pacific (Fournier, 1983).

Acid Sulfate Water

Found in fumarolic areas, acid sulfate waters have very low pH. These waters are mainly a condensate from steam and indicate subsurface boiling conditions. Much of the acidity may arise from reactions involving H_2S (rarely SO_2) being oxidized to form sulfuric acid. This water normally has low Cl^- but may contain a wide variety of cations derived from acid leaching of near-surface rocks.

Calcium Bicarbonate Water

The principal anion of this water, HCO_3^- , forms when CO_2 -rich waters react with silicate minerals. At temperatures below 200°C , the solubility of calcite is relatively high, and calcium is easily leached from volcanic rocks. When these bicarbonate waters emerge at the surface, they lose their remaining CO_2 , causing travertine (a variety of calcite) to deposit.

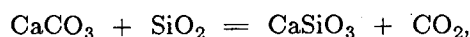
Bicarbonate water and travertine deposits usually indicate low-temperature reservoirs (Holland, 1967).

CLASSIFICATION OF GAS DISCHARGES

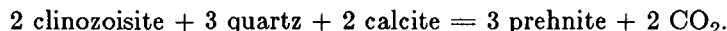
While most of the gas (> 95%) discharging at steam vents and fumaroles is steam (H₂O), various noncondensable gases are present (e.g., CO₂, H₂S, CH₄, H₂, N₂) whose relative abundances have been studied as a guide to the thermal potential of geothermal systems. Gas chemistry has not been particularly successful as an exploration guide because of the wide variations in gas concentrations observed from the same geothermal field (D'Amore and Panichi, 1985). Gas geothermometers depend on a knowledge of the gas/steam ratios of vapor-dominated systems and the steam/water ratios of water-dominated systems. These ratios usually cannot be determined for surface emanations because the gas and water rarely reach the surface together and the phases may have undergone chemical reactions en route to the surface from the reservoir region. Geothermal gas discharges represent a complex process of gas evolution including (a) gas separation from a crystallizing magma (Carmichael et al., 1974), (b) gases (He and Rn) released by radioactive decay, (c) gases (CO₂, SO₂, and H₂S) from thermal metamorphism of rock containing calcite and sulfides, (d) gas (mainly CH₄) evolved from organic material and the serpentinization of ultrabasic rocks, and (e) gases from rock-fluid and fluid-fluid reactions in the more oxidizing near-surface (or reservoir) environment. To this list we can also add the contaminating atmospheric gases (mainly CO₂ and N₂) that are picked up by meteoric waters. On the basis of many studies of dissolved gases in thermal waters, gas discharges at volcanic vents and fumaroles in geothermal areas, and gases driven from igneous rocks by heating in the laboratory, there have been numerous attempts to use the relative proportion of certain noncondensable gases to classify geothermal systems. For example, high-temperature waters, particularly those heated by igneous intrusions, may not only have a high dissolved-solids content but also can be rich in all or a combination of dissolved CO₂, H₂S, He, and H₂ (Arnórsson, 1974). Steam arising from active volcanism is distinguishable from steam related to a circulating hot-water system in that the former contains SO₂ and gaseous HCl and HF. SO₂ has a noticeably acrid smell, but HCl and HF will irritate the eyes and skin (Ellis and Mahon, 1977).

Some gas partial pressures, such as H₂, H₂S, and SO₂, are controlled by mineral buffers. For example, H₂S, SO₂ may be in chemical equilibrium with iron sulfide mineral phases and SO₄²⁻ ions. Armannsson et al. (1982) argued that in Iceland the absence of SO₂ above a magma was due to the precipitation of sulfide minerals in the deeper part of the hydrothermal system.

Ivanov (1967) divided hydrothermal waters into several broad categories on the basis of the dominant noncondensable gases present. N₂-CO₂ waters often represent low-temperature volcanic systems. N₂ is derived from air dissolved in the circulating waters, and CO₂ is derived from a variety of sources, including magmatic and thermal metamorphism of limestone. H₂S-CO₂ waters, on the other hand, are more representative of hydrothermal systems with abyssal heat recharge and are typical of most high-temperature geothermal fields. CO₂ is the principal noncondensable gas present in most thermal areas, and may occur as a result of various reactions. It is generally believed that the major source of CO₂ is from the thermometamorphism of rocks containing calcite or carbonate minerals and silica or various silicate minerals (D'Amore and Nuti, 1977). One simple reaction of this type is



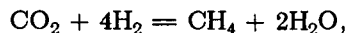
but there are many others that produce CO₂ plus various metamorphic minerals such as epidote, diopside, and prehnite. Cavaretta et al. (1982) showed that the CO₂ partial pressures at the Larderello and Serrazzano geothermal fields are close to equilibrium for the reaction



In such a case it is possible to fit CO₂ partial pressures to an empirical relation involving temperature, but this circumstance does not hold for all geothermal systems even though it may work perfectly well at Larderello (D'Amore and Panichi, 1985).

Recently, evidence for a magmatic CO₂ component has been sought because CO₂ is much less soluble in magma than water (steam) and other noncondensable gases. Isotopic analyses of C in CO₂ collected at Casa Diablo Hot Springs, Long Valley caldera, California, show that the ¹³C/¹²C ratio is consistent with values found in fluid inclusions in igneous rocks. The apparent lack of sedimentary sources of CO₂ beneath the caldera suggests that the CO₂ is exsolving from a contemporary magma (Taylor and Gerlach, 1983).

In addition to CO₂, many geothermal systems contain CH₄ and NH₃. If there is excess CO₂ present, methane may be generated through the Fischer-Tropsch reaction:



but this reaction does not explain the chemical concentrations and isotopic characteristics of CH₄ observed in most geothermal environments (Panichi et al., 1976). Where high-temperature waters and/or steam react with algal deposits or organic-rich sediments, the fluids are likely to contain high percentages of CO₂, CH₄, and NH₃ (Ellis and Mahon, 1977). The close association of CH₄ with organic-rich sediments and the highly variable concentrations of CH₄ observed in volcanic emissions, hot springs, and fumaroles (Ward, 1978; Graeber et al., 1979; Reitsema, 1979) has detracted from the value of CH₄ as an exploration guide. To complicate the matter, there is speculation that outgassing of primordial abiogenic methane from mantle depths is a global process that has been going on since the formation of the planet (Gold and Soter, 1980, 1982).

Inert atmospheric gases such as Ne and ³⁶Ar serve as references for the meteoric water contribution, because they are not produced from rocks in significant quantities. For example, Gunter (1973) found that the total N₂/Ar ratio for the steam plus liquid outflows at Yellowstone was similar to the proportions expected from circulating meteoric water. The very high N₂/Ar ratios and excess H₂ found by Hulston and McCabe (1962) in New Zealand suggest an organic source for some of the N₂ and H₂.

Gaseous emanations in Iceland contain a few to tens of percent H₂. The high concentration of hydrogen is considered typical of active hydrothermal systems associated with volcanic rifts. Most of the geothermal H₂ may result from water disassociation at high temperature and pressure conditions (D'Amore and Nuti, 1977). The actual H₂ in discharges can be greatly modified by other near-surface reactions.

Helium is also of special interest, because the isotope ³He is generally believed to have only a mantle source; i.e., it does not occur as a daughter product in any radioactive-decay series of crustal elements. The use of He as an exploration guide is discussed in the next section.

Finally, there has also been interest in the chemistry of certain metallic elements found in steam condensates. Because geothermometers based on SiO₂ and Na-K-Ca do not work at

fumaroles over vapor-dominated systems, Koga and Noda (1976) have used concentrations of Hg, As, and B in steam condensates to estimate subsurface temperatures.

SAMPLING AND ANALYSIS OF LIQUID AND GASEOUS DISCHARGES

Many parameters can be determined at flowing springs using simple, inexpensive equipment: air and water temperature, pH, electrical conductance, total alkalinity, chloride, and sulfide content. The various techniques for water sampling and analysis are summarized by Mariner et al. (1975) and by Ellis and Mahon (1977). Flow rates of streams and springs are based on simple calculations or visual estimations. Because these parameters can vary depending on season, rainfall rate, and rate of evapotranspiration, a prospect area is usually sampled at different times during a year.

Filtered samples from both warm and cold springs are collected at points as close to the orifices of springs as possible. The samples are immediately acidified for cation analysis to assure that metals remain in solution. Acid is not added to samples taken for anion analysis. Samples for silica analysis are diluted by 1:5, 1:10 or 1:20 with distilled deionized water to prevent silica polymerization.

In contrast to the common practice of sampling waters from hot and cold springs, sampling of gases from low-temperature boiling springs and fumaroles is rarely done because of the difficulty in obtaining samples uncontaminated by atmospheric gases. A gas sampling technique developed and used in New Zealand (Ellis and Mahon, 1977) for many years uses a 1-m-long metal tube driven into the area of the fumarole. Uncontaminated gases are withdrawn into an evacuated bottle containing a concentrated NaOH solution that absorbs the acid gases (CO_2 , HCl , SO_2 , H_2S) while the other gases (H_2 , CH_4) remain undissolved. Gases are then analyzed by gas chromatography and other methods. A similar sampler developed by the Sandia National Laboratory has a glass dewar insert within the metal tube that avoids condensation of H_2O in the tube. Where fumarole gas pressures are low, one is then able to obtain a measure of H_2O in the gas.

Water samples are returned to the laboratory for traditional analysis of major cations (Ca^{2+} , Mg^{2+} , Na^+ , K^+), major anions (Cl^- , SO_4^{2-} , HCO_3^- , F^- , Br^-), silica, and trace metals (B is the only one usually analyzed) that are associated with hydrothermal-volcanic-magmatic systems. In addition to the traditional analytical methods, Bowman et al. (1976) showed that neutron-activation analysis (NAA) and x-ray fluorescence (XRF) techniques yield reliable analyses of major elements and some trace elements. NAA and XRF techniques have not caught on because of the specialized equipment needed.

Helium-Isotope Ratios

Helium-isotope ratios ($^3\text{He}/^4\text{He}$) determined from gas emanations and dissolved gases in hot springs may indicate the origin of helium, since all ^3He is believed to have only a mantle source (i.e., high ^3He in relation to normal ^4He from radioactive decay of crustal U and Th indicates a mantle source). Helium-isotope results are reported both as an absolute $^3\text{He}/^4\text{He}$ ratio and as the sample ratio normalized to the ratio in air:

$$R/R_a = \frac{(^3\text{He}/^4\text{He})_{\text{sample}}}{(^3\text{He}/^4\text{He})_{\text{air}}}$$

Neon is sometimes measured as well to correct the measurements for air incorporated into the sample during sampling or as a consequence of natural hydrologic processes (Torgersen

and Jenkins, 1982). However, this correction is not significant in samples with a high He concentration. The $^3\text{He}/^4\text{He}$ ratios in gas emanations and tholeiitic basalt at oceanic spreading centers (Kilauea, Iceland, and the Galapagos Rift) are 7 to 20 times greater than the normal atmospheric ratio. Craig et al. (1978) found an enrichment in ^3He in gases from a typical continental-margin orogenic province (Mt. Lassen in the California Cascade Range) and in gases from a mid-continent hot spot (Yellowstone Park volcanic caldera). At Mt. Lassen the acid hot-spring gases have a much higher proportion of mantle helium than is found in the neutral-to-alkaline springs, and the acid-gas ratios are similar to ratios found at other convergent-margin volcanoes. The isotopic ratio in the gas phase at Mt. Lassen is twice as high as in the liquid phase, indicating that it would be better to sample the gas phase whenever possible. Craig et al. (1978) also found some high $^3\text{He}/^4\text{He}$ ratios at Yellowstone; they were surprising in that it leads one to question how ratios about 15 times the atmospheric ratio can be maintained over what appears to be a large silicic magma chamber, enriched in U and Th and, consequently, in ^4He .

Table 4 shows the ranges of R/R_a measured at various geothermal and normal crustal areas reported by Craig and Lupton (1976), Craig et al. (1978), and Torgersen and Jenkins (1982). In interpreting R/R_a ratios, one should be mindful that the ratio is lowered by He introduced by country rock weathering and U and Th series decay, a so-called crustal overprint (Torgersen and Jenkins, 1982). In older magmatic systems, where magma has become isolated from its source, ^4He will also accumulate *in situ* because of U and Th series decay within the aging but still hot plutonic body. Finally, fluids that have picked up tritium (^3H) will produce ^3He by decay and may thus give a magmatic appearance.

Helium isotopes do not seem to be used widely in routine exploration, possibly because helium and neon measurements require mass spectrometry. A good summary of sampling and analysis techniques is given by Mazor (1976) and Torgersen and Jenkins (1982).

Noble Gases

Mazor (1976) studied atmospheric noble gases (He, Ne, Ar, Kr, Xe) in thermal waters from several areas in the world. He found that their relative abundances are a good indicator of whether a water discharge is derived from a boiling zone at depth or from a cooler reservoir that has not experienced boiling. Nonboiling waters retain the noble gases derived from the atmosphere, while boiling waters at depth are depleted in their noble gases, which partition into the gaseous phase.

Oxygen-Hydrogen Isotopes

Clues to the thermal age of a circulating hydrothermal system have been sought from the ratios of $^{18}\text{O}/^{16}\text{O}$ and D/H in cold- and hot-spring waters. Plotting these ratios in parts-per-thousand (‰) change from standard mean ocean water (SMOW), geochemists find that normal (nonthermal) meteoric waters are depleted in both heavier isotopes, ^{18}O and deuterium. This is because the heavier isotopes tend to remain in the oceans, and proportionally fewer of these atoms find their way into the precipitation that falls on land. The depletions also vary with latitude, altitude, and distance from the sea. For example, most of the ^{18}O and D in the atmosphere falls closest to the ocean, with progressively fewer atoms falling over inland areas. Consequently, the δD and $\delta^{18}\text{O}$ values of meteoric waters lie close to a linear depletion curve called the meteoric water line (Figure 8); the more inland sites lie further down the line to the left. Craig (1963) showed that the thermal waters for the same localities had about the same deuterium as meteoric waters but showed an "oxygen shift" to

TABLE 4
Range of R/R_a Ratio at Geothermal and Normal Crustal Sites.

Type of Site	Example	R/R_a
Active hot spots	Hawaii, Iceland	14-25
Mid-ocean ridges and spreading centers	Juan de Fuca Ridge	8-10
High-temperature geothermal systems	The Geysers Lassen Peak fumaroles Yellowstone	6.6-9.5 ~ 8 5-15
Subduction margins	Circumpacific belt	5-8
Hydrothermal systems	Steamboat Springs, Nevada Late Cenozoic, Basin and Range system	3.7-6.1
Low- to moderate-temperature geothermal systems	Raft River, Idaho	0.13-0.17
Stable crustal areas; U and Th decay	Gas and oil wells	~ 0.1

Data from Craig and Lupton, 1976; Craig et al., 1978; Torgersen and Jenkins, 1982.

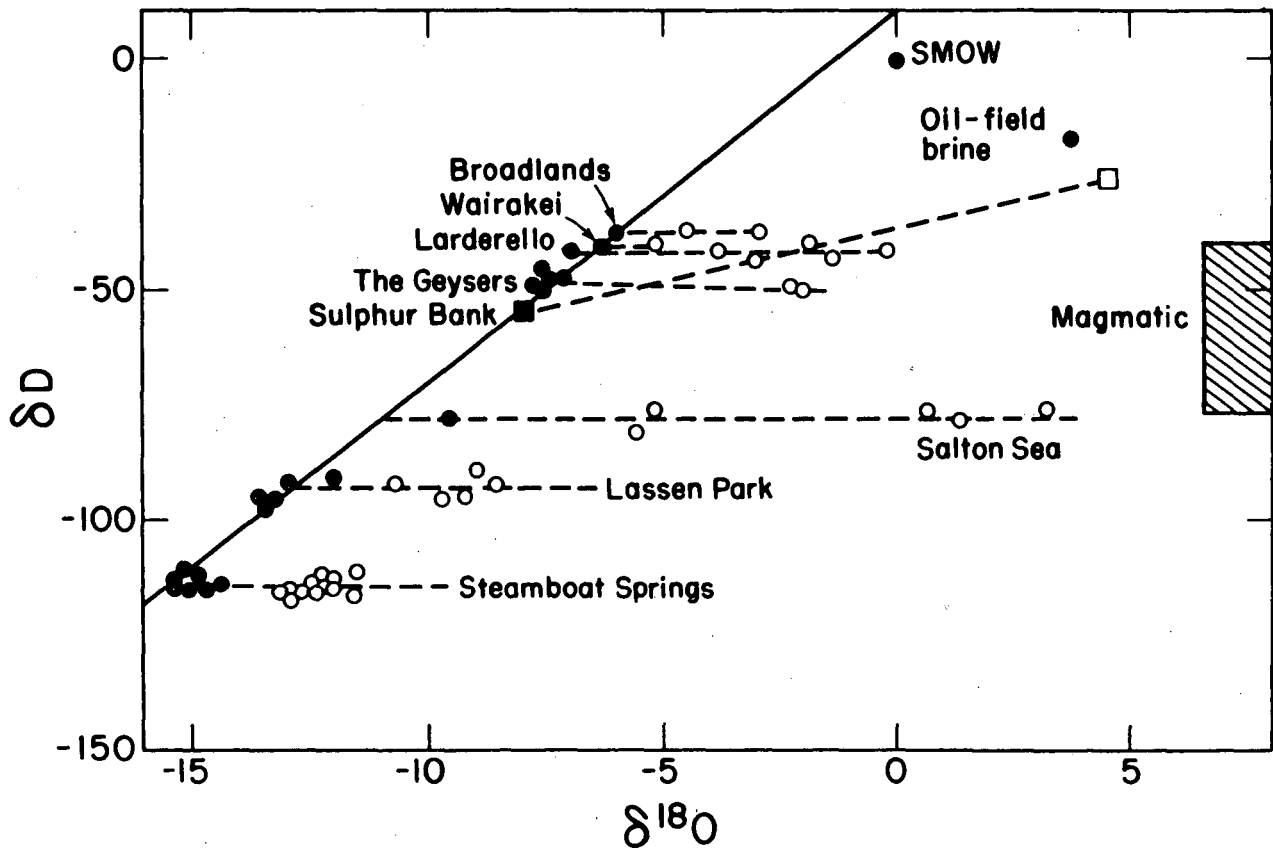


Figure 8. Values of $\delta D/\delta^{18}O$ for various meteoric waters, showing the meteoric water line (solid line) and the "oxygen shift" (broken lines). Filled circles and squares show points for meteoric waters. Open circles and squares show points for thermal waters (from Ellis and Mahon, 1977). (XBL 8312-2441)

higher amounts of ^{18}O . This effect is due to exchange reactions between the ^{18}O -depleted meteoric waters and the ^{18}O -rich silicate minerals. On the other hand, most rocks contain little deuterium for exchange.

The amount of oxygen shift in thermal waters varies considerably between geothermal areas; even within a specific area—the Salton Sea, for example—the oxygen shift can vary over a large range. In the simplest model, a small oxygen shift corresponds to an older geothermal area, one having already experienced considerable throughput of circulating meteoric water. A larger oxygen shift would correspond to either a young system or one through which there has been less water flow over geologic time. This simple model doesn't apply everywhere. For example, if the circulating meteoric waters mixed with ancient sea water, both δD and $\delta^{18}\text{O}$ values would be shifted toward values observed for connate water in compacted marine sediments. White (1965, 1974) reported that formation waters from these rocks gave δD values in the range of -10 to -20 ‰ and $\delta^{18}\text{O}$ values of $+3$ to $+5$ ‰ with respect to SMOW.

Although oxygen isotopic shifts have been studied at both hydrothermal ore deposits and geothermal systems, the data are difficult to interpret quantitatively for exploration purposes. If the circulating meteoric water has reached isotopic equilibrium with the rocks before reemerging at the surface, one might in theory be able to differentiate between an old, cooler system (small shift) and a young, active system (large shift). Cole (1983) found compelling evidence that oxygen exchange in geothermal systems at 200 – 300°C should occur in the process of propylitic alteration, which leaves a mineral assemblage of smectite, chlorite, epidote, albite, quartz, and carbonate. He also found, from studies of oxygen isotopic fractionation between rocks and fluids in geothermal systems, that equilibrium is rarely achieved and that the degree of equilibration can vary considerably over distances of only a few meters within the system. The general failure of the silicate reactions and the isotopic exchanges to reach equilibrium led him to conclude that local self-sealing of the fracture plumbing system for part or all of the thermal event could produce the disequilibrium conditions found. Using a rate model developed for studying water-rock equilibration, he found that the times required to produce the isotopic shifts seen in geothermal systems are typically less than 200 years and as brief as 10 years. Because these times are short compared to the total lifetime of a major convecting hydrothermal system (10^4 – 10^6 years), the residence times predicted from the model may represent only the time during which the meteoric water came into contact with high-temperature rocks.

Tritium

One indication of the youthful age of geothermal fluids is the presence of tritium (^3H or T). Tritium, a radioactive isotope of hydrogen with a half-life of 12.5 years, occurs in all meteoric water because of (1) nuclear reactions induced by cosmic-ray reactions with hydrogen in the upper atmosphere, and (2), more importantly, recent thermonuclear atmospheric explosions. Using the T/H ratio and a mixing model for mixing between young and older waters, one can estimate how long the water has been away from the atmosphere.

Geochemical Geothermometry

Samples of freely flowing fumaroles or hot springs are routinely collected and analyzed to estimate reservoir temperatures. Too often, unfortunately, this may be the extent of the geochemical evaluation. Although there are many different chemical and isotopic reactions that may be used as a guide to the temperature at which water and rock equilibrated, the

more widely used techniques are silica concentration, Na^+/K^+ ratio, $\text{Na}^+/\text{K}^+/\text{Ca}^{2+}$ relationship, and fractionation of the oxygen isotope $\delta^{18}\text{O}$ between HSO^- and H_2O . These geochemical geothermometers, summarized by Fournier and Truesdell (1973; 1974), Fournier (1977, 1981), and Truesdell (1984), work as well as they do because the chemical species reequilibrate slowly with the rock as they move from the reservoir region to the cooler surface.

In general, the different geochemical geothermometers (Table 5) can be expected to indicate different reservoir temperatures, and thus the calculated temperatures must be analyzed in terms of subsurface effects such as:

- (a) mixing of waters from different parts of the reservoir,
- (b) possible dilution from near-surface meteoric waters,
- (c) subsurface boiling,
- (d) effect of pH and salinity on quartz solubility, and
- (e) residence time of fluids in the reservoir.

McNitt (1976) noted that most geochemical geothermometers underestimated temperatures of known reservoirs for reasons stated above, and therefore it would be imprudent to reject a prospect on the basis of indicated reservoir temperature alone. In addition, errors will also be introduced if improper sample collection caused sample contamination. Most techniques seem to work best over some limited temperature range, generally above 100°C but below 250°C .

The Na/K and Na-K-Ca methods are less affected by subsurface mixing and boiling than other common geothermometers, provided there is little Na, K, or Ca in the diluting water. Thus they apply to many high-temperature volcanic systems. However, in systems where CaCO_3 is being precipitated due to the loss of dissolved CO_2 after the waters leave the reservoir, the diminished concentration of calcium in the water will lead to an incorrectly high reservoir temperature. To check if the Na-K-Ca geothermometer is sensitive to possible CaCO_3 precipitation, one can double the measured calcium concentration and recalculate the equilibration temperature. If the initial and recalculated temperatures differ by no more than a few degrees, then calcium precipitation is not strongly influencing the temperature estimate (Fournier and Truesdell, 1973). In some waters, magnesium interferes with the Na-K-Ca geothermometer, and a correction must be applied (Fournier and Potter, 1978).

Silica geothermometers are based primarily on the temperature-dependent solubility of quartz, chalcedony, alpha cristobalite, or amorphous silica (Fournier, 1973). The quartz-solubility relation is used for all high-temperature waters (180 to 250°C) and for lower-temperature water in granitic (i.e., high-silica) rocks. The chalcedony-solubility relation is often appropriate for low-temperature reservoirs and may be the controlling silica mineral in basaltic (low-silica) rocks up to 180°C (Arnórsson, 1975).

Recently Fournier and Potter (1982) presented a revised quartz geothermometer for conductive cooling from temperatures as high as 330°C . In cases where the fluid cools conductively from any temperature below 330°C during ascent to the surface, the improved quartz-solubility relation should provide an accurate estimate of that reservoir temperature as long as the solutions are dilute.

High-salinity brines alter quartz solubility; the effect is relatively minor below 300°C . Of more practical concern are the separate effects of adiabatic cooling (boiling) and the mixing of ascending high-temperature waters with cold meteoric water to give warm springs

TABLE 5
Equations for Some of the Commonly Used Geothermometers

Geothermometer	Relation between chemical concentration and temperature†	Effective temperature range
Silica Geothermometers (SiO₂ concentration in mg/kg)		
Quartz—no steam loss (conductive cooling)	$T(^{\circ}\text{C}) = \frac{1309}{5.19 - \log \text{SiO}_2} - 273.15$	0 – 250 °C
Quartz—maximum steam loss (adiabatic cooling)	$T(^{\circ}\text{C}) = \frac{1522}{5.75 - \log \text{SiO}_2} - 273.15$	0 – 250 °C
Chalcedony	$T(^{\circ}\text{C}) = \frac{1032}{4.69 - \log \text{SiO}_2} - 273.15$	0 – 250 °C
α-Cristobalite	$T(^{\circ}\text{C}) = \frac{1000}{4.78 - \log \text{SiO}_2} - 273.15$	0 – 250 °C
β-Cristobalite	$T(^{\circ}\text{C}) = \frac{781}{4.51 - \log \text{SiO}_2} - 273.15$	0 – 250 °C
Amorphous silica	$T(^{\circ}\text{C}) = \frac{731}{4.52 - \log \text{SiO}_2} - 273.15$	0 – 250 °C
Alkali Geothermometers (Na, K, and Ca concentrations in mg/kg)		
Na/K (Fournier)	$T(^{\circ}\text{C}) = \frac{1217}{\log(\text{Na/K}) + 1.483} - 273.15$	> 150 °C
Na/K (Truesdell)	$T(^{\circ}\text{C}) = \frac{855.6}{\log(\text{Na/K}) + 0.8573} - 273.15$	> 150 °C
Na-K-Ca	$T(^{\circ}\text{C}) = \frac{1647}{\log(\text{Na/K}) + \beta * [\log(\sqrt{\text{Ca/Na}}) + 2.06] + 2.47} - 273.15$	

Source: Fournier, 1981; Henley et al., 1984.

* $\beta = 4/3$ for $\sqrt{\text{Ca/Na}} > 1$, $T < 100^{\circ}\text{C}$; $\beta = 1/3$ for $\sqrt{\text{Ca/Na}} < 1$, $T > 100^{\circ}\text{C}$

†Throughout this report T is the only symbol used for temperature ($^{\circ}\text{C}$).

with so-called "mixed" waters. Truesdell and Fournier (1977) developed a silica-mixing-model correction based on enthalpy/silica-concentration diagrams. The same procedure can also be used to correct silica concentrations for adiabatic cooling, assuming single-stage steam loss at any temperature. As an example, consider the enthalpy-silica curve (*AB*) that has been experimentally determined over the temperature range from 50 to 350 °C (Figure 9). Point *E* represents the presumed silica-enthalpy conditions of the water at a vigorously boiling spring at an elevation where water boils at 90 °C. From steam tables, the enthalpy of liquid water at 90 °C is 376.9 J/g, while that of the coexisting steam is 2660 J/g. Because steam contains less than 1 mg/kg dissolved silica at temperatures less than 265 °C, point *E'* in Figure 9 characterizes the steam that separates from the boiling spring water. If the water cools entirely by boiling, and all the steam remains with the water until the mixture reaches the surface, then the silica concentration and enthalpy of the original water can be determined by finding point *C*, the intersection of a straight line between *E-E'* and the quartz-solubility curve *AB*.

SAMPLING AND ANALYSIS OF SURFACE SOILS AND ROCKS

Detailed sampling of soils and rocks is rarely done because of the time and costs involved. These techniques may be employed where zones of vertical hydraulic conductivity are obscured by recent alluvium. The careful and detailed mapping of hydrothermal-alteration intensity and patterns, supported by grid soil/rock geochemistry, may serve as a guide to geothermal-hydrothermal convection circulation (Bamford, 1978; Bamford and Christensen, 1979). Following the methods proposed by Bamford (1978), the hydrothermal-geochemical effects may be enhanced by separating for analysis the nonmagnetic and dense (+3.3 g/cm³) fraction from whole rock. This eliminates rock-forming and alteration silicates, enriching the sample in hydrothermal oxides and sulfides that carry the more interesting trace elements (Pb, Zn, Hg, As).

SAMPLING AND ANALYSIS OF SURFACE VOLATILE TRACE ELEMENTS

Studies of the volatile trace elements Rn, He, and Hg in soils and soil gases over geothermal areas have been described by many workers; e.g., Rn (Wollenberg, 1974; Stoker and Kruger, 1976; Nielson, 1978), He (Bergquist, 1979; Hinkle and Kilburn, 1980), and Hg (Matlick and Buseck, 1976; Klusman et al., 1977; Klusman and Landress, 1978; Phelps and Buseck, 1980; Varekamp and Buseck, 1983). The advantages of these techniques are that costs are low and results can sometimes be obtained using portable field equipment. A disadvantage, perhaps minor, is that they usually require a soil cover into which a sampling tube can be driven or from which soil samples can be easily taken.

Radon

Radon (²²²Rn), a gaseous daughter product of ²³⁸U, has been used as a guide to convecting hydrothermal systems. Solid-state nuclear-track detectors are employed in the Track Etch[®] technique (Fleischer et al., 1975), patented by Terradex Corporation (Alter and Price, 1972) and the General Electric Vallecitos Laboratory to detect alpha particles from the decay of ²²²Rn (half-life 3.8 days). Inverted plastic cups with a specially treated dielectric alpha-track detector are buried in shallow backfilled holes to detect the alpha particles emanating directly below the inverted cup mouth. After several weeks of exposure the cups are retrieved and the tracks are counted in the laboratory.

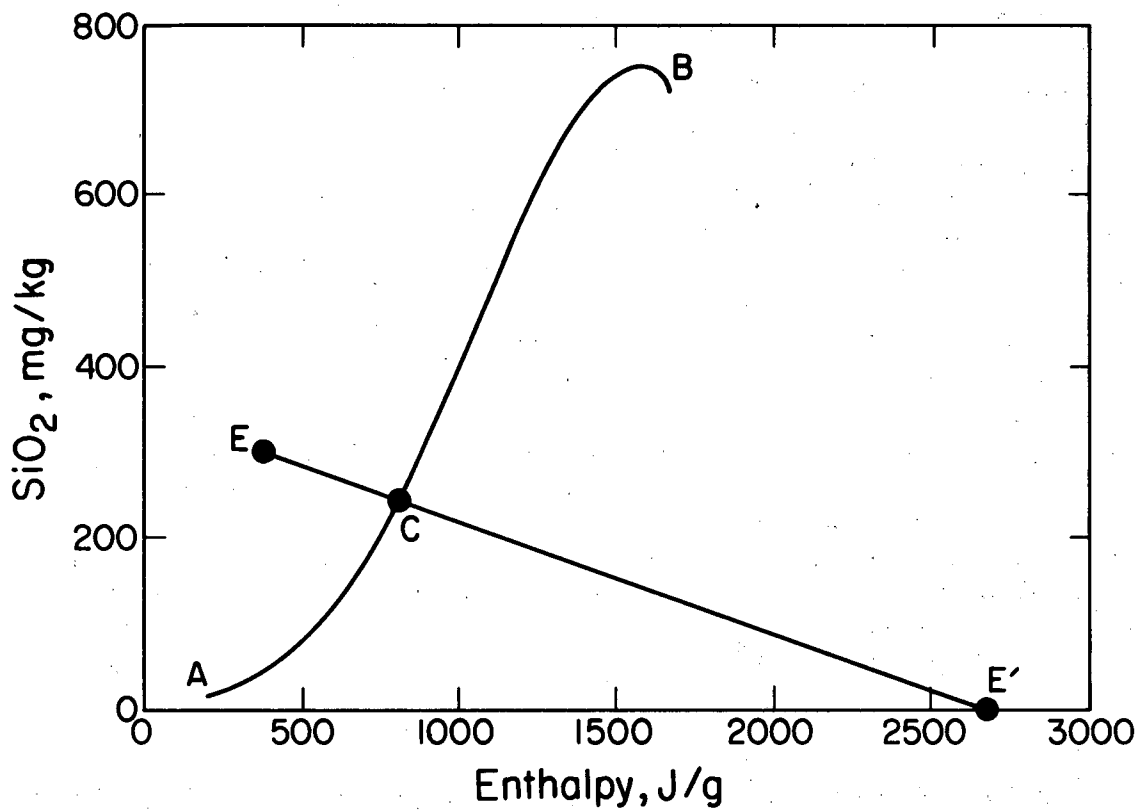


Figure 9.

Enthalpy/silica-concentration diagram used for correcting silica concentrations for adiabatic cooling, assuming single-stage steam loss at any temperature. Point *C* represents the silica concentration and enthalpy of the original water. See text for meaning of *AB* and *EE'* lines (from Truesdell and Fournier, 1977).
(XBL 8312-2440)

Radon is believed to migrate through the overburden by a combination of diffusion and advection (Tanner, 1964, 1980). If diffusion were the predominant mechanism, the migration distance of ^{222}Rn would be only a few meters in a dry soil with average porosity, and less for a saturated or a compacted soil (Kristiansson and Malmquist, 1982). Under such conditions, ^{222}Rn concentrations would reflect only the ^{238}U content of the local soil/rock. However, there is growing evidence that long-range transport of ^{222}Rn occurs in volcanic areas (Cox, 1980; Cox et al., 1980) and under both geothermal and nongeothermal conditions (Mogro-Campero and Fleischer, 1977). To explain the long-range transport phenomenon, Kristiansson and Malmquist (1982) proposed a model in which ^{222}Rn is carried in a stream of carrier gas composed of atmospheric gases and gases liberated by chemical reactions. Radon anomalies in geothermal areas might then be caused by a combination of fluid circulation, permeable paths to the surface, and a high availability of carrier gases.

Wollenberg (1974) found high track densities (several hundred to several thousand tracks/mm²) close to thermal mounds and pools in Nevada, but density values varied considerably at short distances from the springs. Corrections for background Rn in the soil around the cups were made to ascertain the ^{222}Rn ascending from depth along faults. In general, the background values could be explained by the ^{238}U content of the local valley fill and suballuvial rocks. Higher track densities occurred where a thin alluvial veneer covered rhyolitic ash-flow tuffs whose ^{238}U contents are higher than that of basaltic cones and flows in other parts of the valley.

Whitehead et al. (1983) reported on a detailed radon survey using an improved version of Track Etch[®] over the Craters of the Moon area, New Zealand. The entire area has a high ground temperature (as high as 60 °C), and Rn values are also high. However, the imperfect correlation between Rn and soil temperature was attributed to local variations in type and thickness of pumice and in the type and density of vegetation whose root system increases near-surface permeability. The authors found that isolated ^{222}Rn anomalies correlated well with known faults and that, when the technique is used for fault detection, the cups should be placed no farther apart than 60 m.

Mercury

Field-portable instrumentation has been developed for measuring Hg in soils and soil gas. The initial application was for mineral exploration; e.g., the Scintrex Model HGG-3, a Hg spectrometer that works on the principle that Hg vapor attenuates the Hg emission line at 254 nm from a Hg lamp. To collect a sample, soil gas is hand-pumped via a sample tube driven into the ground (Robbins, 1973).

Perhaps the most thorough published study and evaluation of Hg in soils as a geothermal-exploration method was done by Varekamp and Buseck (1983). Soil samples were collected from A1 horizons at several places in the western U.S. In arid areas, samples were collected at 5–10 cm depths or from a clay-rich part of the soil. Sample separation varied from 500 to 1000 m during the initial reconnaissance, and then 50- to 100-m intervals were used once an area of interest was identified. Hg analyses were performed using a Jerome Instrument Corporation detector, Model 301 (McNerney et al., 1972). Samples of 250–500 mg were heated, and the released vapors were deposited on a gold-plated wire coil. The amount of Hg was measured by the change in electrical resistance of the coil. The lower limit of detection was approximately 10 ppb. Hg soil anomalies were found associated with all geothermal areas studied. Peak enrichment occurred in the hot-spring environment, around which occurred a broader aureole population (Figure 10).

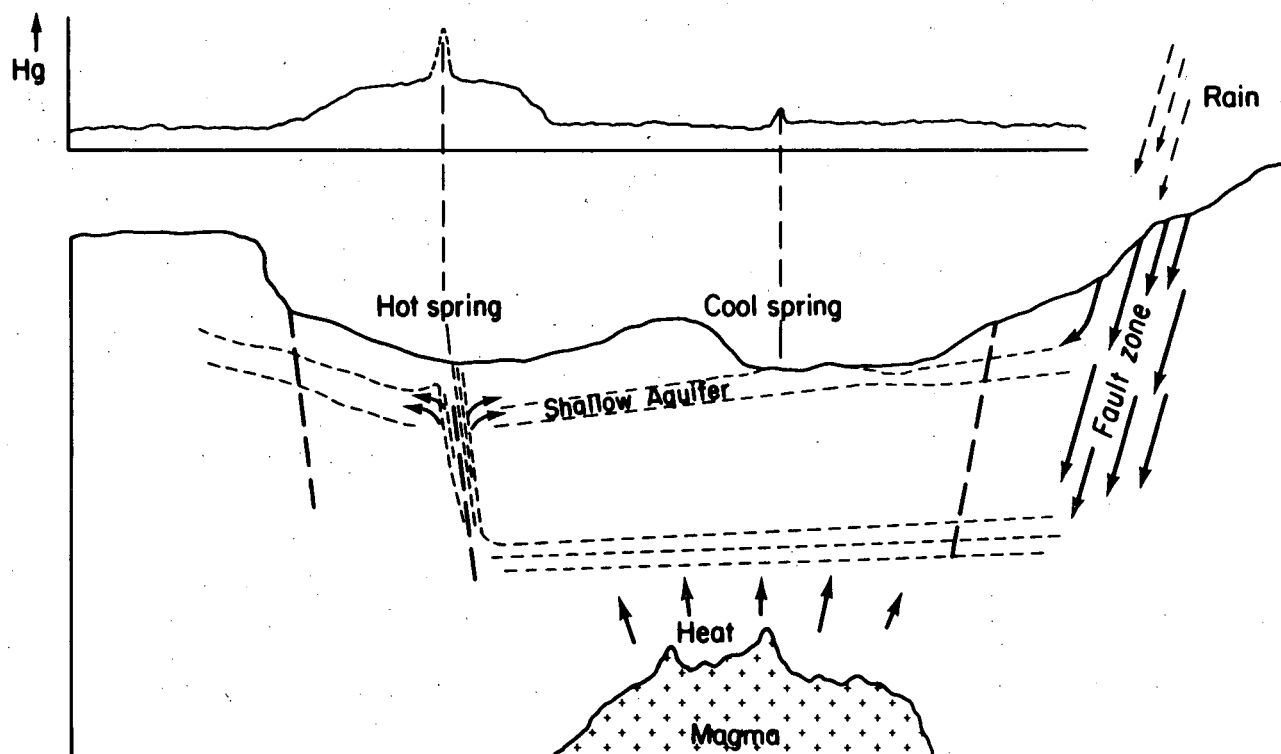


Figure 10. Schematic of a Hg anomaly at a geothermal area, showing peak Hg concentration (top curve) above a zone of ascending hot water (from Varekamp and Buseck, 1983). (XBL 8312-2442)

Because Hg partitions strongly into the vapor phase, fumaroles and their associated acidic hot spring will be enriched in Hg. Copp (1981) found anomalous concentrations of Hg over a 16-km² area at the Coso Hot Springs geothermal area. Comparisons of the Hg anomaly with heat flow (Combs, 1980) and electrical resistivity (Jackson and O'Donnell, 1980) revealed a good concordance between Hg and a convective heat-flow anomaly and an area of low resistivity. This is an active system with a steam cap at about 350 m above a hydrothermal-magmatic system. Because steady Hg loss is a dynamic process, fossil hydrothermal systems may not show Hg enrichment.

Varekamp and Buseck (1983) concluded that Hg soil surveys can be a cost-effective method of locating promising areas, because broad aureole anomalies above ascending zones of enriched thermal water are detectable by widely separated samples. However, at Roosevelt Hot Springs, a liquid-dominated reservoir, Hg anomalies occurred only in close coincidence with hot-spring deposits and near a shallow producing well (Capuano and Bamford, 1978). In contrast, background Hg values were found in nearby locations lacking indications of subsurface thermal activity, leading Capuano and Bamford (1978) to conclude that Hg anomalies were specific to structures that control fluid flow in geothermal systems and useful in their identification.

At the Meager Creek geothermal area, British Columbia, where a thick groundwater-saturated overburden overlies fractured crystalline and metamorphic basement rocks on the south side of the Meager volcanic complex, Fairbank et al. (1981) reported inconclusive results from both Rn and Hg surveys. Rn anomalies may have been suppressed by the thick overburden, and Hg content was found to depend on organic content in the soil.

⁴Helium

Measuring ⁴He in soil gas is simpler in some respects than studying the ³He/⁴He ratios of fluid emanations. The basis for this technique is that He is soluble in hot water, and the He produced by the normal radioactive decay of U and Th in crustal rocks will be scavenged by hydrothermal fluids. As the water approaches the surface and cools, a certain amount of He will be released and diffuse upward into the soil (Roberts et al., 1975). Because early measurements of ⁴He in soil gases near various hot-spring areas in the western U.S., including Yellowstone National Park, showed anomalous concentrations, Roberts et al. (1975) applied a portable helium detector to the problem of mapping He anomalies around a hot spring near Idaho Springs, Colorado. They used a small (1-cm radius) truck-mounted mass spectrometer that was set to collect ⁴He ions and reduce interference from C³⁺ ions. Gas was slowly pumped out of the ground via a tubular steel probe driven to a depth of about 0.5 m. The mass spectrometer was periodically switched between the soil gas and a standardized sample of ⁴He in compressed air. The comparison technique gave a detection threshold of 50 ppb; an entire measurement could be obtained in 3 to 4 minutes. Figure 11 shows the He isopleths plotted on a logarithmic scale near the spring. Background readings in the area showed the 5.2 ppm concentration typical of the atmosphere.

Compared with Rn and Hg field methods, measuring He in soil gas requires more sophisticated equipment; this might preclude the general use of the technique in parts of the world. McCarthy (1983) reports that sample analysis with a highly modified helium-leak-detector mass spectrometer by Dynatech, Inc., provided cost-effective soil-helium exploration in the Animas Valley, Colorado. The survey cost, including use of the sampling equipment, was \$25 per sample.

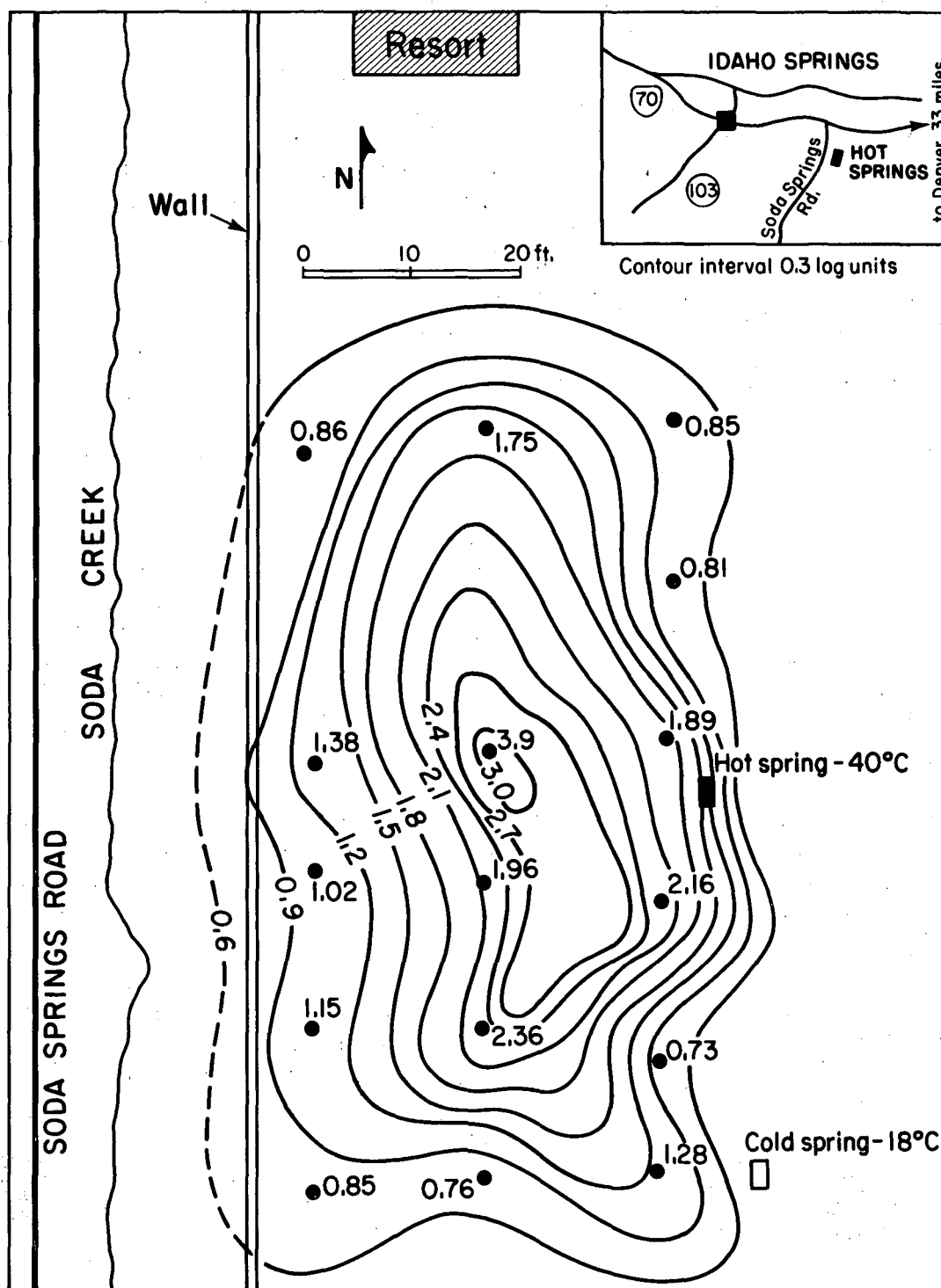


Figure 11. Logarithm of ^4He isopleths near a 40°C hot spring in Colorado (from Roberts et al., 1975). Notice that the peak of the helium anomaly is displaced from the springs.
(XBL 841-397)

SAMPLING AND ANALYSIS OF SUBSURFACE ROCKS

Present Subsurface Temperatures and Field Boundaries

The analysis of rock chips and cores from shallow to moderate-depth holes drilled for temperature information provides late-stage exploration information that can be helpful for guiding field development. Bamford (1978) found that As and Hg seemed to delineate most clearly both the Roosevelt Hot Springs and The Geysers reservoirs. The elements gave broad surface anomalies in permeable zones over the hottest or shallowest parts of the reservoir and also produced small subsurface anomalies close to thermal fluid entries in geothermal wells. However, in a follow-up study at The Geysers, Moore et al. (1982) found that Hg in wells drilled to depths of up to 3 km concentrated primarily in the outer, cooler portions of the thermal systems. The relative depletion of Hg within the reservoir rocks compared with concentrations near the margins of the field could be a useful exploration guide in active, vapor-dominated hydrothermal systems.

Present subsurface temperature distribution can be determined in a variety of ways. As discussed in other parts of this report, detailed studies of alteration minerals, geochemical geothermometry, and direct measurements of temperatures in boreholes are the most commonly applied techniques. The relation between mineral assemblages in cores and the compositions of formation fluids from deeper wells should provide information on thermal processes, such as whether the system is waxing or waning, and possibly some information on water-rock reactions and circulation. That information is extremely important for evaluating the chemistry and temperature of surface discharges. One might, therefore, argue for deeper drill holes and sampling at the earliest opportunity, particularly in areas lacking vigorously flowing discharges and where fluid mixing is suspected.

Dating Thermal Events

One method used to determine the age of a single, current thermal event is to look for the point at depth where fission tracks, defects left in minerals by the natural decay of ^{238}U , are completely erased or annealed. Knowing the temperature at that depth and assuming isothermal conditions have been maintained since the onset of heating, one can then estimate the duration of heating.

Sanford (1981) and Sanford and Elders (1981) found that the absence of fission tracks in apatite at a depth of 1940–1980 m in well T-366 at Cerro Prieto show that heating at 160–180 °C has been going on for 10^3 to 10^4 years. This age for the geothermal system is consistent with ages from paleopole positions determined in young dikes cutting the nearby Cerro Prieto volcano (de Boer, 1979).

Sanford (1981) used apatite because it is a ubiquitous detrital mineral in the Salton trough, and its annealing behavior is well studied at temperatures of 210 to 350 °C and for heating times of a few minutes to months. Data also exist on the geological annealing of fission tracks in apatite at temperatures between 100 and 150 °C and times of 10^5 and 10^8 years. Both laboratory and field data characterizing annealing can be extrapolated to lower temperatures and heating times, thus providing a means of estimating the age of younger geothermal events. A limitation of this technique is that it gives a reliable age only for simple events, i.e., instantaneous heating of the rock to the present temperature followed by a constant temperature. This assumption is valid in geological situations where the time over which temperature increased to the present elevated temperature is short compared to the total duration of heating. Where the temperature increase proceeded in a slow step-wise

fashion due to episodic volcanism, fission-track annealing would underestimate the age for the onset of heating.

Vitrinite reflectance is the basis of another technique that provides information on the thermal history of sedimentary rocks. The ratio of reflected to incident light from dispersed, detrital plant remains known as phytoclasts varies with the degree of thermal alteration, an irreversible process, and thus the ratio is related to the maximum temperature experienced by the rock (Piller, 1977). Vitrinite reflectance was applied to shale cuttings from four wells at Cerro Prieto (Barker, 1979; Barker and Elders, 1979, 1981) and to sedimentary rocks in other liquid-dominated geothermal systems in the U.S. (Barker, 1983). The Cerro Prieto wells showed good correlation between the reflectance profiles and downhole temperature logs, together with consistent temperature estimates from fluid-inclusion and oxygen-isotope geothermometry. The results indicate that the Cerro Prieto system is currently at its maximum temperature. Barker (1983) also showed that at six liquid-dominated systems, vitrinite reflectance could be fitted to the maximum temperature experienced by the rock through a regression equation (Figure 12):

$$R_m = 0.435 \exp 0.00683 T (^{\circ}\text{C}).$$

The coefficient of determination (r^2) is 0.8, suggesting that the technique, notwithstanding errors from measurement and sampling and the effect of heating time, has merit as an exploration method if used to predict undisturbed temperatures in a well from cuttings before thermal equilibrium is regained. Barker (1983) argues that time (i.e., duration of the organic-matter metamorphism) is secondary to maximum temperature. He feels that thermal stability in the hydrothermal organic-matter reactions is reached in about 10^4 years or less.

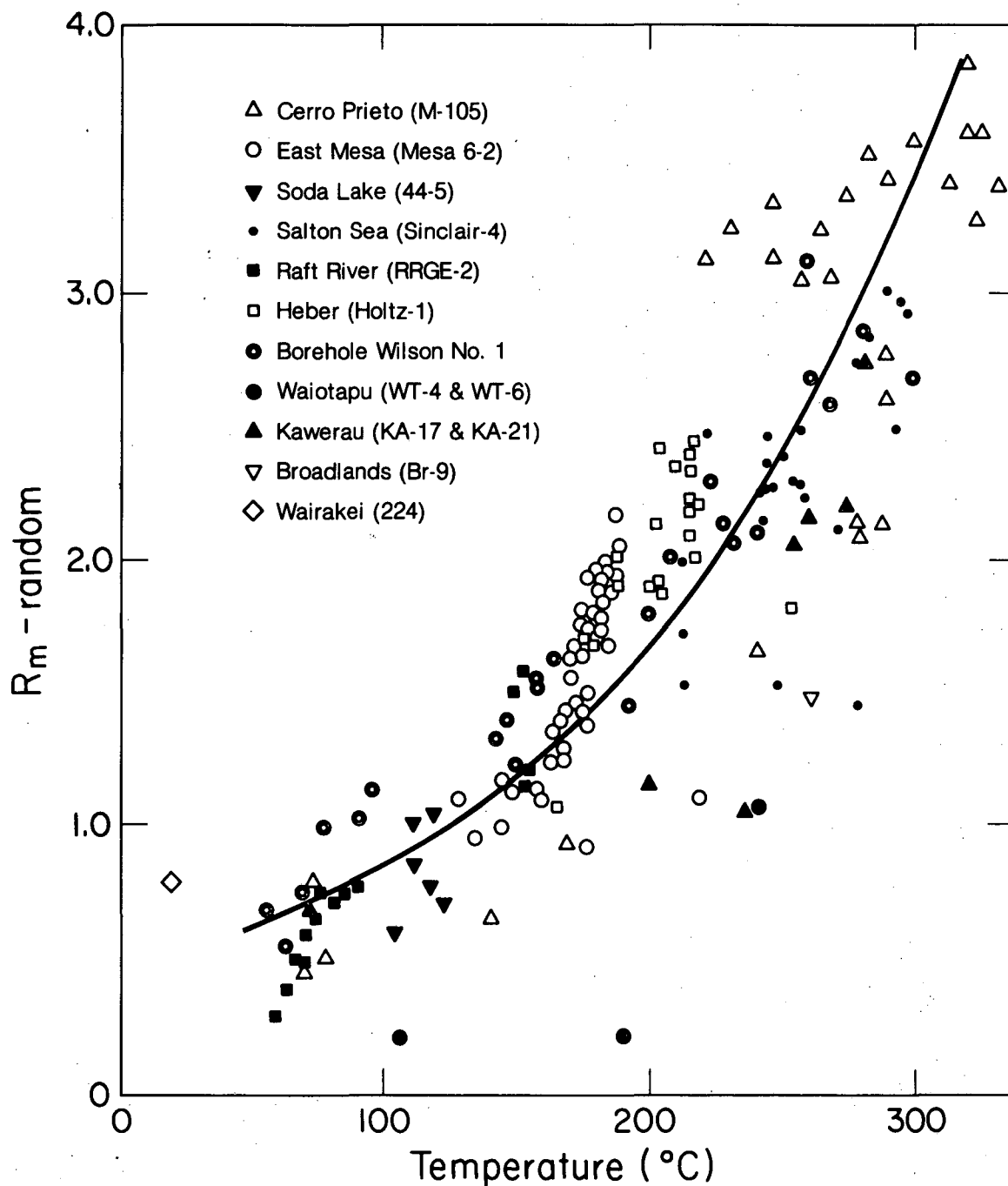


Figure 12. Vitrinite reflectance (R_m) versus temperature for several geothermal areas. The curve describes the regression equation $R_m = 0.435 \exp 0.00683 T (^{\circ}\text{C})$ (from Barker, 1983). (XBL 8312-2444)

GEOPHYSICAL TECHNIQUES

Because geothermal exploration is concerned with deducing depth and dimensions of a reservoir of hot water and/or steam, all principal geophysical-exploration studies have involved measurements of subsurface temperatures and heat flux. Coupled with geological and geochemical studies, temperature measurements made in shallow boreholes have sometimes proved to be a simple strategy for delimiting targets for deeper drilling. This simple strategy does not apply everywhere, nor can it locate reservoir rocks with sufficient porosity and permeability for sustained production. Thus other geophysical techniques have been tested and evaluated. As a result of these studies, industry and research organizations now have many geophysical tools and methodologies that can provide additional indirect information on thermal conditions and locations of potentially favorable reservoir rocks. As this section attempts to bring out, geophysical data require careful processing and interpretation to avoid erroneous and misleading conclusions. The relative success of geophysical investigations also depends on establishing realistic expectations between geologist, geophysicist, and management on what geological information can be obtained from the geophysical methods available.

THERMAL AND TEMPERATURE SURVEYS

Thermal techniques provide the only direct measure of the target sought. For this reason there has been more research on heat generation and transport and thermal measurements than on any other aspect of geothermal exploration. Kappelmeyer and Haenel (1974) and Rybach and Stegena (1979) provide detailed discussions and extensive bibliographies.

Basic Heat-Flow Equations

It is self-evident that exploration for high-temperature resources will be directed to locales within younger orogenic areas characterized by a higher-than-average regional heat flow and recent volcanism. The internal heat of the earth escaping to the atmosphere is called the surface heat flow (Q), and in the absence of a shallow crustal heat source (e.g., a cooling pluton) it is expressed as the sum of the radiogenic heat production in the crust $A(z)$ plus the mantle heat flow (Q_m); viz.,

$$Q \text{ (mW/m}^2\text{)} = Q_m + \int_0^{z_m} A(z) dz, \quad (1)$$

where z_m is the thickness of the crust in which the radioactive elements U, K, and Th are concentrated. Heat-flow data for the Sierra Nevada Province (Lachenbruch, 1968) and the eastern U.S. (Birch et al., 1968) seem to obey a linear relation, which has led workers to express Equation (1) simply as

$$Q = Q_r + DA_o,$$

where Q_r is called the "reduced heat flow," i.e., the heat flow for zero radioactive heat from the crust. D is the characteristic depth of radioactive heat production for the area in question (e.g., $D = 10$ km has been used to interpret heat-flow data in the Basin and Range Province). Although the accuracy of this simple relationship has been questioned (Lachenbruch and Sass, 1978), it is generally believed to apply to provinces where heat-flow density from the lower crust is uniform and where local effects from hydrologic and magmatic

convection are unimportant. These conditions and assumptions do not generally apply in geothermally prospective areas. In such areas, measured values of Q may be widely dispersed about the regional value, and a local population of values may have a mean 50% higher than the regional value. One such case is found in Nevada, where the average heat flow in the Battle Mountain heat flow high is 3 heat-flow units (HFU) (126 mW/m^2)*, compared with 2 HFU for the surrounding region. Lachenbruch and Sass (1977) concluded that heat-flow variations in the Basin and Range Province are caused by convective processes and their related transients and that these are far more significant, by a factor of 3 or 4, than effects of lateral variations in crustal radioactivity.

Most explored geothermal systems show significant heat transfer by convection, a process in which cold meteoric surface waters at temperature T_s are heated to temperature T_r at reservoir depth and then ascend buoyantly along a fault-fracture system, often discharging as hot springs or fumaroles. The convective heat-flow component can be estimated from the discharge of the thermal manifestations:

$$Q_{conv} = \frac{Q_G \dot{d}}{Area},$$

where $\dot{d}$ is the volumetric discharge rate (L/s), $Area$ is the discharge area, and Q_G is the volumetric heat gain of the water:

$$Q_G = \rho(T_r) \cdot [h(T_r) - h(T_s)],$$

where $\rho(T_r)$ is water density at reservoir depth, and $h(T_r)$ and $h(T_s)$ are the fluid enthalpies at temperatures T_r and T_s , respectively (Rybach, 1981).

From both a scientific and practical viewpoint, one would like to eliminate the thermal masking effects caused by near-surface convective transfer by hot and cold groundwaters in order to determine the actual geothermal conditions at depth. This requires measuring the conductive temperature gradient dT/dz or the related conductive heat flow Q_{cond} ,

$$Q_{cond} = \kappa \frac{dT}{dz}, \quad (2)$$

by drilling a number of holes to sufficient depths that the thermal gradients dT/dz are demonstrably constant. The technique also requires that the rock thermal conductivity κ be measured either *in situ* or in the laboratory, and that the thermal gradients be corrected for terrain effects (Blackwell et al., 1980).

For reasons discussed in this section, heat-flow measurements are not routinely made in subregional or detailed-stage exploration. Most thermal measurements are limited to measuring bottom-hole temperatures with a maximum-reading thermometer or to obtaining a continuous temperature profile with a suitable logging system. These efforts are designed to outline areas of highest near-surface temperatures and to develop a preliminary concept of the hydrogeology, such as the influence of faults and aquifers on the flow of cold and heated waters.

*1HFU = $10^{-6} \text{ cal} \cdot \text{cm}^{-2} \cdot \text{s}^{-1} = 41.8 \text{ mW} \cdot \text{m}^{-2}$.

Shallow Temperature Surveys

In spite of the fact that thermal discharges may be displaced horizontally from the thermal source and the reservoir rocks, considerable attention has been directed toward shallow temperature surveys. This is understandable in view of the high cost of deep drilling relative to the immense geological uncertainties at the early stages of subregional exploration. Consequently, techniques have been developed and used to obtain thermal information from shallow holes, 1 to 5 m deep, that can be auger-drilled quickly and inexpensively (Lee, 1977; Olmsted, 1977; Lange et al., 1982; LeShack and Lewis, 1983). Where cold groundwater does not mask thermal effects, a common and simple technique involves burying single-thermistor probes in shallow auger-drilled, backfilled holes. After the probes equilibrate, usually after 24 hours, the temperatures at depth z_0 are read and analyzed. To a close approximation, the observed temperature $T(z_0)$ can be expressed algebraically as

$$T(z_0) = T(0) + z_0 \frac{dT}{dz} + T_d + T_s.$$

$T(0)$ is the mean annual surface temperature, dT/dz is the thermal gradient, T_d is the diurnal temperature variation, and T_s is the seasonal temperature wave. Whereas T_d is small below 1 m, T_s will distort temperatures to depths of 10 to 20 m, and a correction must be made for it in order to obtain a better estimate of the thermal gradient. Lange et al. (1982) represented the seasonal wave as

$$T_s = T_a e^{-mz} \cos(\omega\Delta t - mz),$$

where

- $m = \sqrt{\pi/\tau\alpha},$
- $\omega = 2\pi/\tau,$
- $\tau = \text{the period of the wave} = 1 \text{ year},$
- $\Delta t = \text{the elapsed time since the summer maximum},$
- $T_a = \text{seasonal temperature amplitude},$
- $\alpha = \kappa/c\delta = \text{thermal diffusivity},$
- $\kappa = \text{thermal conductivity},$
- $c = \text{specific heat of the rock, and}$
- $\delta = \text{rock density}.$

As long as all the thermal measurements are made over a short time interval and α is constant over the survey area, then it is relatively simple to estimate and apply uniform T_s correction to all holes. As a practical consideration, this approach works best if the area surveyed is small enough that α is approximately constant and the required number of temperature probes is installed and read at one time.

Separations between holes can vary from tens of meters to 1 km or more, depending on the size and depth of the thermal source. One could use the same approach to planning a thermal survey as a gravity survey, because the associated anomalies are a related potential-field function (Simmons, 1967). For example, if one were attempting to delineate a fault zone along which thermal waters ascend, the hole spacing might be as little as 10 to 50 m (Kappelmeyer and Haenel, 1974). On the other hand, if the survey area is large, a temperature survey could require hundreds of holes, requiring a precise correction for the seasonal temperature wave. Rather than using a single thermistor, multiple thermistors can be

attached to rigid and cable probes. These require deeper holes, at least 3 m deep, but they provide temperature-gradient information directly, as well as a temperature profile (Lee, 1977; Lange et al., 1982). Lange et al. (1982) describe instrumentation and a technique in which thermal anomalies were outlined as well from 3-m holes as from 40-m holes. They used seven thermistors spaced 0.25 m apart between 1.5 and 3.0 m in a sealed PVC pipe. Multiple thermistors allowed them to detect a significant part of the seasonal wave, from which they were also able to estimate the diffusivity of the near-surface material. This allowed them to detect and separate diffusivity variations from true thermal anomalies. Figure 13 shows a comparison of a thermal anomaly mapped at the east side of Dixie Valley, Nevada. The shaded area represents the Augusta Mountains. Figure 13a shows the corrected temperatures at the bottom of 40-m-deep holes. Figures 13b and 13c show the corrected temperatures estimated from 3-m-deep holes on two dates 38 days apart. Because of the higher sampling density of the shallow holes, the shallow-hole temperature data show a more complex anomaly pattern.

Shallow-hole temperature surveys have been done mainly in the Basin and Range and Salton Sea Provinces, where ease of vehicle access and a thick soil cover make it possible to auger-drill many holes economically and where there is no problem from near-surface groundwater flow.

Temperature-Gradient Measurements

In areas where, for reasons of rock outcrop and near-surface hydrology, shallow-hole methods are inappropriate, moderate-depth holes are often drilled (to several hundred meters) by means of conventional mud or air rotary drilling or wireline coring if lost-circulation zones occur and a suite of rock cores is desired for study. Such holes may range in depth from 50 to 600 m; deep enough, as a rule, to penetrate completely through near-surface zones of cold- or hot-water convective flow and thus provide a reliable estimate of the conductive thermal gradient.

Combs (1980) describes one method for completing this type of hole for accurate temperature measurements. To prevent hole collapse after drilling, a watertight polyvinyl chloride (PVC) casing or black iron pipe, sealed at both ends, is put into the hole and held in place by backfill or grout. The casing is filled with water so that a stable temperature distribution results. Adequate time must be allowed for the temperatures to reequilibrate. A rule of thumb is to allow 10 to 20 times the drilling time for the disturbances in temperature caused by drilling to decay to 1% of their original values.

Temperature measurements are made either continuously or at intervals by a wireline temperature probe, usually a platinum resistance electrode with an accuracy of about 0.5°C (Ross et al., 1977) and a surface read-out. It is common practice to run several temperature logs days to weeks apart after circulation is stopped to get multiple-temperature reequilibration profiles that help identify hot- or cold-water entries. Temperature profiles in the deeper rotary holes may also be used to check and supplement thermal data from the 1- to 3-m-deep holes. However, numerous problems are associated with obtaining temperature data from deeper, rotary-drilled wells (e.g., the need for casing and cementing, repeated visits, and long waits to obtain equilibrium temperature values).

In hydrologically complex areas, such as the Long Valley caldera (Lachenbruch et al., 1976; Sorey et al., 1978) and the Newberry Volcano, Oregon (Sammel, 1983), moderate-depth test drilling and temperature logging reveal a great deal of hydrological information regarding lateral heat and mass transfer within permeable volcanic units. Figure 14 shows

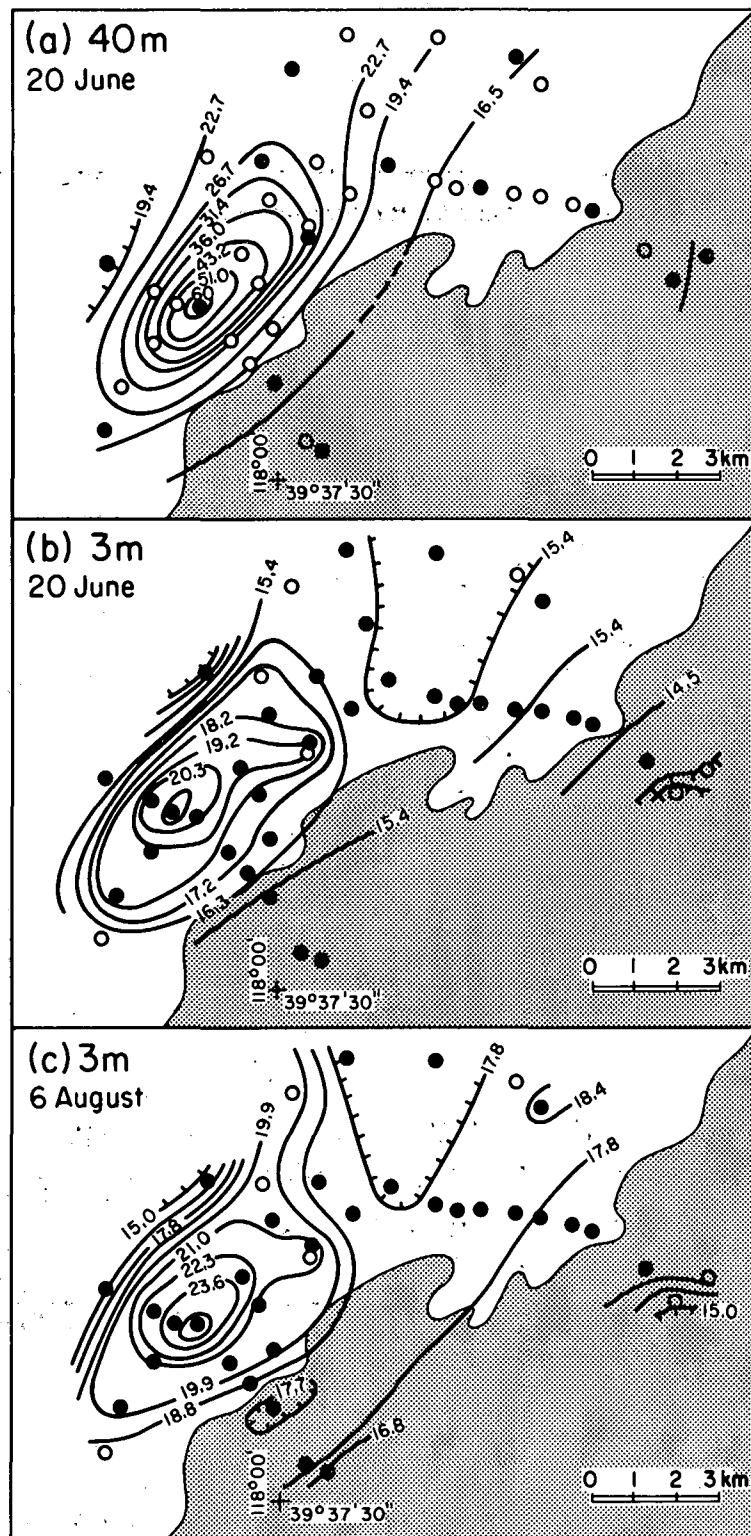


Figure 13.

Thermal anomaly mapped at the east side of Dixie Valley, Nevada. (a) Corrected temperatures at the bottom of 40-m-deep holes. (b,c) Corrected temperatures estimated from 3-m-deep holes on two separate days (from Lange et al., 1982). (XBL 841-9517)

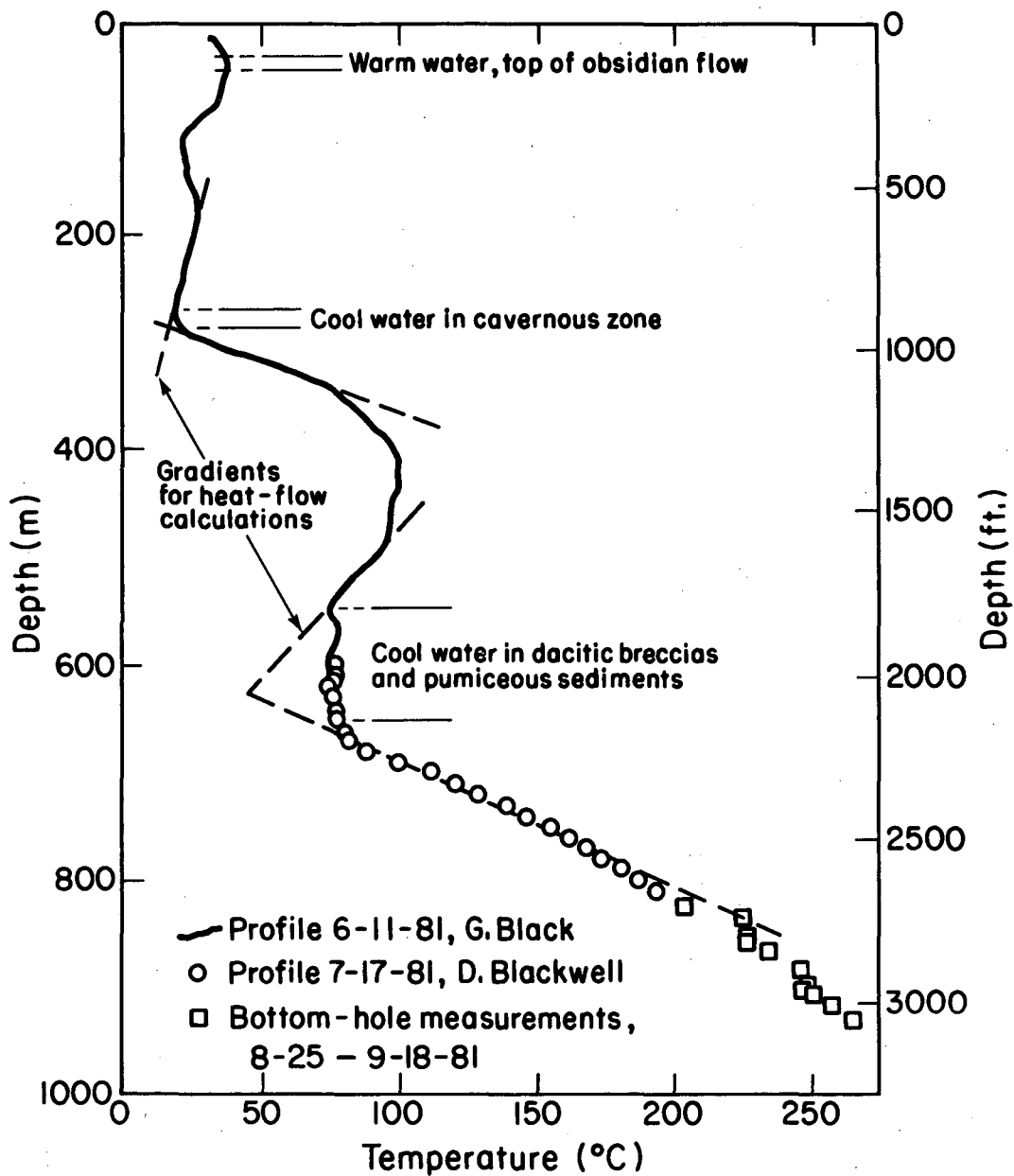


Figure 14. Temperature profile from a USGS test hole at the Newberry Volcano, Oregon. Note the linear temperature gradient below 675 m (from Sammel, 1983). (XBL 841-9518)

the temperature profile recorded in the U.S. Geological Survey test hole, Newberry 2, drilled close to the center of the Newberry Volcano, a Holocene caldera measuring 6 by 8 km (Sammel, 1983). The variations in the temperature profile above 675 m indicate both cold- and hot-water flow in separate volcanic units. Below 675 m the temperature profile, in an impermeable basaltic andesite, shows a conductive (linear) gradient of about 500°C/km. Temperature inversions and positive and negative temperature peaks, such as recorded in the Newberry 2 test well, are fairly typical of conditions observed in wells where hot or cold fluids enter and leave the well bore via permeable stratigraphic units or major fractures.

Thermal-Conductivity and Heat-Flow Measurements

To determine the conductive heat flow (Eq. 2) of a target area, the rock thermal conductivity must be obtained using cores or chips taken from the interval of the hole showing a linear temperature gradient. Thermal conductivity is measured from cores in the laboratory using a divided-bar apparatus (Roy et al., 1968) or from chips in the field using a needle probe (Combs et al., 1977). The divided-bar technique involves comparing the rock against a standard whose thermal conductivity is known (Birch, 1950; Goss and Combs, 1975). The needle-probe method (Von Herzen and Maxwell, 1959) depends on the transient rate of radial flow away from a linear heat source.

Because of the time and cost involved in the laboratory measurement and because the needle probe works best on a fine-grained mineral aggregate, there has been considerable interest in finding simpler, rapid, and practical field techniques for determining thermal conductivity. For example, Goss and Combs (1975) examined in detail the possibility of estimating thermal conductivities of Imperial Valley rocks from common geophysical borehole logging parameters (e.g., porosity, density, compressional-wave velocity). A digitized suite of borehole logs was studied in relation to divided-bar thermal conductivities or cores from the same well, and an empirical relationship was found that predicted thermal conductivity reasonably well. Poppendiek et al. (1982) reported the development of two experimental transducers that were designed to measure heat flux and thermal conductivity directly in shallow holes.

Sass et al. (1981) discuss a real-time method for determining temperature, thermal conductivity, and hence heat flow in unconsolidated sediments during rotary drilling. After drilling to the depth of measurement (< 100 m), they drive a probe hydraulically through the bit up to 1.65 m into the formation. The 2-m-long steel probe contains three thermistors 0.5 m apart that provide continuous temperature records during and following emplacement. The passive temperature record is typically run for 1500 seconds, long enough to permit extrapolation to equilibrium temperatures as $1/t \rightarrow 0$. Then a current of about 100 mA is applied to a line-source heater in the probe, and thermal conductivity is calculated from the rate at which temperatures change over 15–20 minutes.

Although heat-flow measurements are valuable for regional evaluations, they may not always be performed in the subregional to detailed exploration stages. Besides the cost of coring and performing the thermal-conductivity measurements in hard-rock environments, many other factors seem to limit the general usefulness of this method. For example, some shallow to moderate-depth holes may not yield a linear temperature gradient; or if they do, the gradient has to be corrected for the effects of drilling, topography, erosion, etc. It is a well-known problem that terrain corrections must be made to measured temperature gradients in areas of perturbing topography (Blackwell et al., 1980). Not only must the actual topography be known to a reasonable approximation, but the numerical solution for the true heat flow also requires knowledge of surface temperatures and the subsurface distribution of

thermal conductivities (Henry and Pollock, 1982). Reader and Fairbank (1983) measured heat flow in 15 diamond drill holes at Meager Creek, British Columbia, and concluded that the effort was mainly useful for refining the raw temperature profile information but did not help in developing distinct exploration targets.

Temperature and thermal-conductivity data from 25 wells within a 20-km radius of Mount Hood, a stratovolcano in north-central Oregon, indicated a very complicated pattern of heat transfer (Steele and Blackwell, 1982). The data elucidated the variable shallow groundwater circulation around the volcano. The holes, which varied in depth from 65 m to 1.8 m, could be drilled no closer than 5 km from the summit, and were unable to detect the presence of a neck-type magma chamber. On the basis of the conductive gradients there was also no suggestion of a large subvolcanic magma chamber with a top 3 km or less from the surface and a radius of 2 to 3 km, as has been proposed for some andesite stratovolcanoes (Steele and Blackwell, 1982).

MAGNETICS

High-level aeromagnetic surveys with a line separation of ~ 2 km, together with low-level surveys flown at constant terrain clearance with a line separation of ~ 0.4 km, have been flown over prospective geothermal areas mainly as a mapping technique. The aeromagnetic surveys are sometimes followed by more detailed ground magnetic traverses over areas where higher resolution data are needed. Aeromagnetic data for the U.S., flown at a 3-mile (4.8-km) line spacing, was collected by the U.S. Department of Energy and can be obtained on tape from the EROS Data Center, Sioux Falls, South Dakota 57198.

Whereas a commercial contractor must be employed to collect and process aeromagnetic data, ground data can be obtained easily by trained field personnel using modern, light-weight proton-precession magnetometers (0.1-nT sensitivity) with digital readouts and the capacity to store data internally. Line and station separations used in ground surveys can vary appreciably depending on the local magnetic field variations. Used as a mapping aid, magnetics may provide some of the following information:

- (a) location and depth of concealed intrusives,
- (b) location and extent of major faults, and
- (c) areas of possible hydrothermal alteration.

Unfortunately, magnetics has tended to be one of the less informative exploration methods in many geothermal areas. The authigenic mineral assemblages of many hydrothermal-geothermal systems studied contain only small amounts of ferrimagnetic minerals (magnetite-titanomagnetite and pyrrhotite) (Cavaretta et al., 1982), and the geologic significance of many magnetic features detected by surveys may not become apparent until surface and subsurface geologic data are compiled and other geophysical data have been collected and interpreted.

On the other hand, there are reported cases where magnetics have provided a direct indication of a geothermal system. The best example of this involves the correlation between magnetic lows to zones of intense hydrothermal alteration. Extensive magnetic lows such as the Broadlands Geothermal Field, New Zealand (Hochstein and Hunt, 1970), and discrete circular lows such as the Coso geothermal field (Fox, 1978; Roquemore, 1984) have been correlated with zones of argillic hydrothermal alteration, a feature attributed to the alteration of magnetite to hematite and ferric hydroxides by oxygenated, acidic

hydrothermal waters. Magnetic lows in geothermal areas also occur as narrow, linear zones of 100-nT change associated with specific faults, such as in the Basin and Range Province, along which faults circulate hydrothermal fluids (Goldstein et al., 1976; Halliday and Cook, 1978). However, magnetic lows may occur for other reasons; e.g., the large magnetic low in the vicinity of Diablo Hot Springs within the Long Valley caldera may be only a roof pendant of nonmagnetic metasediments concealed beneath the Bishop Tuff, which fills the caldera (Williams et al., 1976).

In Iceland, detailed ground-magnetic surveys are used to locate narrow linear features such as dikes and faults where the basement is concealed by a cover of soil or sediments. Flövenz and Georgsson (1982) describe one such successful application of magnetics for siting a well at the intersection of a reversely magnetized dike (i.e., a long narrow magnetic low) and a line of warm springs. They surmised that hot water was moving laterally along the dike boundary before ascending and discharging, 40 °C cooler, along a fracture zone oblique to the dike.

Less numerous are the direct relations between magnetic highs and geothermal systems. Pyrrhotite, a weakly magnetic iron sulfide, forms in a reaction between pyrite and hydrogen sulfide gas at temperatures above 300 °C and therefore probably occurs in many high-temperature reservoirs. Fine-grained pyrrhotite and pyrite were reported at Cerro Prieto (Elders et al., 1979), but the volume concentration of pyrrhotite observed was far too low and its occurrence too deep to yield a recognizable magnetic anomaly.

Magnetic highs associated with recent intrusions of igneous rocks can be a diagnostic feature in some locales. In Iceland, magnetic highs due to specific dikes were used as a guide to high-permeability fault and fracture zones bounding the dikes (Pálmasson, 1976). Magnetic highs in the Salton trough are caused by the intrusion of basaltic dikes and larger, deeper plutons concealed within the thick, nonmagnetic pile of Pliocene-Recent marine, deltaic, and lacustrine sediments. There is good evidence at both the Salton Sea (Griscom and Muffler, 1971) and Cerro Prieto geothermal fields (Goldstein et al., 1984) that the magnetic anomalies are associated with magma emplacement into an extensional basin or spreading center (Elders et al., 1972). Results of deep development drilling and geophysics at the eastern part of the Cerro Prieto geothermal field have shown that the large (300-nT) circular high located east of the original production area is probably due to a magnetite-rich portion of basalts and gabbros emplaced into the shallow crust as a consequence of still-active tectonic stresses, and that continued magma injection is the heat source (Goldstein et al., 1984). The distance between the Cerro Prieto magnetic anomaly and the hydrothermal surface manifestations is large enough that the anomaly could have been missed or ascribed to basement structure if earlier subregional aeromagnetic and ground surveys had not been conducted.

Although only rarely used at the subregional stage of exploration, the Curie isotherm analysis has been used to identify broad areas of hotter crust. This Curie isotherm occurs at a depth where the temperature exceeds the Curie point for the ferrimagnetic minerals present; i.e., the temperature at which ferrimagnetic minerals become paramagnetic. A standard assumption in Curie-point depth analysis is that magnetite (Fe_3O_4), with a Curie temperature of approximately 575 °C, is the main ferrimagnetic mineral (Bhattacharyya and Leu, 1975). The occurrence of impurities, mainly titanium in the magnetite lattice, reduces the Curie temperature to 520–560 °C in most continental rocks (Buddington and Lindsley, 1964). However, where the main ferrimagnetic mineral is titanomagnetite, particularly a variety with a high percentage of the ulvöspinel (Fe_2TiO_4) end member of the $x\text{Fe}_2\text{TiO}_4 \cdot (1-x)\text{Fe}_3\text{O}_4$ solid-solution series (Nagata, 1961; Irving, 1964), then the Curie

temperatures might be 200 °C or less. Oceanic tholeiite basalts have low Curie temperatures, and this type of rock was encountered at Cerro Prieto (Goldstein et al., 1984). Numerous geologic situations may occur where a magnetic anomaly is produced by rocks containing both titanomagnetite and magnetite or where the ferrimagnetic minerals are zoned with depth, e.g., titanomagnetite-rich basalts grading downward into magnetite-rich gabbros. In the latter case, two distinct Curie-point isotherm depths could be present.

Several techniques for determining the Curie-point isotherm depth from aeromagnetic data have been tried with varying degrees of success. The simplest technique involves fitting of an individual anomaly or its Fourier transform to a depth-limited body, such as a magnetized vertical prism or cylinder (Bhattacharyya and Leu, 1975; Byerly and Stolt, 1977; Shuey et al., 1977; Goldstein et al., 1984). The problem with this numerical approach, as well as others, are that the depth sought, the depth-to-bottom estimate d , is the least-well resolved parameter and the one most likely to be influenced by how the data are de-trended to account for regional effects. Other complications involve interference effects from neighboring anomalous bodies and the usually specious assumption that the source has a specified, simple geometry.

Where multiple interfering anomalies occur, two-dimensional (2-D) statistical techniques have been used. This approach assumes that the anomalies within a large area are due to a randomly distributed ensemble of vertical prisms embedded in a nonmagnetic host; each prism has arbitrary dimensions but is magnetized in a direction close to the present field (Spector and Grant, 1970). A radially averaged wavenumber spectrum for the gridded map area is computed, and the slopes of best-fit, straight-line segments to the spectrum are found. These indicate the depths of sources beneath the plane of observation. If the map area is large enough so that the very low frequency components from the prism bottoms contribute to the spectrum, the spectrum may show a low-frequency peak (Spector and Grant, 1970). Boler (1978) showed that the frequency $f_{\max}$ of this spectral peak is related to the mean depth d to the source bottoms by

$$f_{\max} = \frac{1}{2\pi(d-h)} \ln \frac{d}{h},$$

where h is the mean depth to the ensemble of source tops.

In order for the spectral technique to succeed, the area covered by the magnetic survey must be sufficiently large. For example, to resolve a source bottom at depth d , the length L of the survey area must be such that

$$L > 2\pi d$$

(Shuey et al., 1977). Because typical values of d lie in the range of 10 to 20 km, spectral techniques must be applied to fairly large data sets, as was done by Connard et al. (1983), who studied a $1^\circ \times 1^\circ$ area in the High Cascade Province of central Oregon.

There are conflicting reports on the efficacy of Curie-isotherm analysis for geothermal exploration. Using the same aeromagnetic data taken over Yellowstone National Park, Wyoming, Bhattacharyya and Leu (1975) and Won and Son (1983) independently calculated the Curie depths by spectral analysis of the 2-D filtered data and direct inversion of filtered aeromagnetic profiles, respectively. While both approaches indicated a shallowing of the Curie isotherm near the ring-fracture zone of the caldera, the two sets of results were sufficiently different to raise questions on the accuracy of either approach. In another case, Kam (1980) applied a 2-D spectral analysis to aeromagnetic data over the Imperial Valley

and found surprisingly large (20-km) Curie depths for the Salton Sea and Brawley geothermal areas, where much shallower depths (< 10 km) are expected on the basis of known temperature gradients. It is hard to say why the results contradict those that seem intuitively more reasonable. Errors in Curie depths can arise from a variety of causes, such as terrain effects, overprocessing or improper processing of the raw data, too few data, and assuming the wrong model for the magnetized region. It should also be mentioned that the Curie depth may not be a well-defined physical boundary; the thermal demagnetization of the ferromagnetic minerals may occur over a depth range of more than 1 km, depending on the thermal gradient and the mineralogy.

GRAVITY

Gravity surveys, often conducted with magnetic surveys, provide information that helps reduce the ambiguities of magnetic interpretation and information on subsurface density variations. Gravity surveys are far more difficult to perform properly than magnetic surveys, and the data are more difficult to reduce, but the work can be worth the effort.

Gravimeter readings are first converted to accelerations, using the appropriate conversion for the meter. These values are then corrected for instrument drift by distributing changes in meter readings that occur between reoccupations of a base station. Finally, the Bouguer gravity anomaly is calculated for each station; viz,

$$g_B = g_{obs} \pm dg_L + dg_{FA} - dg_B + dg_T,$$

where

g_{obs} is the corrected gravity from the meter reading,

dg_L is the latitude correction,

dg_{FA} is the correction for distance above sea level (free-air correction),

dg_B is the correction for the excess mass between the station and sea level (Bouguer correction), and

dg_T is the correction for local terrain variations near the station.

If precise gravimetry is done, one also needs to correct for the tidal effects of sun and moon. Worden and LaCoste and Romberg gravimeters are used for exploration. Both measure the force required to restore a mechanical beam to a horizontal position. A discussion on gravimeters and surveying techniques can be found in most textbooks on applied or exploration geophysics, such as Telford et al. (1976).

The usual objectives of gravity surveys in geothermal exploration are to discern structural features (normal faults and grabens), to obtain depth-to-bedrock estimates in basinal areas, and to delineate concealed 3-D density inhomogeneities caused by intrusives and zones of hydrothermal metamorphism. Because many geothermal areas under exploration are in topographically rugged areas (e.g., Quaternary volcanic belts), meter readings must be carefully reduced to avoid meaningless and misleading anomalies. A significant source of error may be introduced from the Bouguer and topographic corrections if the wrong density is assumed for the near-surface rocks. There has been a tendency to use 2.6 to 2.65 g/cm³ as the average rock density, but this may be too large a value (Finn and Williams, 1982, 1983). The average density of surface rocks should be verified from rock and core samples or from "density profiles" run over selected topographic features (Nettleton, 1976).

Gravity profiles may be fitted to 2-D density models (Webring, 1985), and gravity map

data may be gridded and fitted to 3-D density models (Cordell and Henderson, 1968) using automatic inversion routines. However, before this is done the Bouguer gravity values, g_B , at each station should be corrected for the regional gradient, the long wavelength trends due to density variations in the deep crust-upper mantle. The objective is to obtain a "residual Bouguer" gravity map that only contains upper-crustal information. If a regional gravity survey covers the area of the detailed survey, the regional effects may be removed with some degree of confidence. If no regional data exist, the regional effects may be estimated and corrected using one of the various numerical techniques described by Nettleton (1976). The two commonest techniques are (1) to use grid operators in the spatial domain to achieve a high-pass filtered gravity map and (2) to find and remove a trend surface (a potential surface described by a polynomial fitted in a least-squares sense to the data). Regardless of the operation, the separation of regional from local effects is complicated by the continuous and overlapping nature of their spectra. As a result, removing "regional" effects may introduce an unknown element of error into the residual anomalies. Couch and Gemperle (1979) describe numerical procedures and results of their attempts to extract local structural information from gravity data taken around Mt. Hood, Oregon.

Gravity surveys have provided insight into rock densification due to water-rock reactions, a phenomenon that explains the relation between gravity highs and convecting thermal waters in the Salton trough (Meidav, 1970; Goldstein and Carle, 1986) and at hot-spring sites within the Basin and Range Province (Goldstein and Paulsson, 1978). These gravity highs are usually associated with sedimentary or volcanoclastic reservoir rocks, which, because of their higher initial porosity and permeability than, say, fractured granite, are more likely to manifest density increases due to the precipitation of hydrothermal silica, calcite, and/or zeolites in pores and fractures. In the higher-temperature parts of a geothermal system ($> 250^\circ\text{C}$), densification can also proceed through chemical reactions that yield a calc-aluminum silicate assemblage (greenschist facies) denser than the original minerals (Elders et al., 1979).

Where small gravity changes due to hydrothermal densification are sought, a La Coste and Romberg Model G geodetic meter with a precision of $1\ \mu\text{Gal}$ (1 ppb of the Earth's gravitational attraction) is used. In practice, the realizable accuracy of the best raw data after proper looping techniques and tidal corrections is normally $\pm 15\ \mu\text{Gal}$ because of instrument tares (Grannell et al., 1981). Accuracy also depends on good station elevations. In flat areas elevations may be estimated from topographic maps to an accuracy of $\pm 1\ \text{m}$, but this accuracy level could introduce an uncertainty of $\pm 0.2\ \text{mGal}$ at any one station (Sawyer and Cook, 1977).

The effect of hydrothermal alteration on density is illustrated in Figure 15, which shows a density-depth plot for a geothermal well at Cerro Prieto based on well logs. The lithology consists of a deltaic sequence of sandstones and shales. The A/B contact, which crosscuts lithologies, is a gradational contact between relatively fresh (A) and hydrothermally altered (B) sediments (lower greenschist). Below the contact, shale densities exhibit a marked increase from 2.15 to $2.5\ \text{g/cm}^3$; a smaller increase is noted for the sandstones. Numerical analysis of gravity anomalies in the Salton trough may reveal interesting and important features of the hydrothermal system. A direct 3-D numerical inversion of the Bouguer gravity anomaly over the East Mesa geothermal field (Biehler, 1971) reveals densification effects that can be attributed to hydrothermal fluids ascending a complex set of faults (Goldstein and Carle, 1986). Figure 16 shows the subsurface topography of the densified zone. The shallow, ridge-like features correlate with self-potential and thermal anomalies and are consistent with drilling results.

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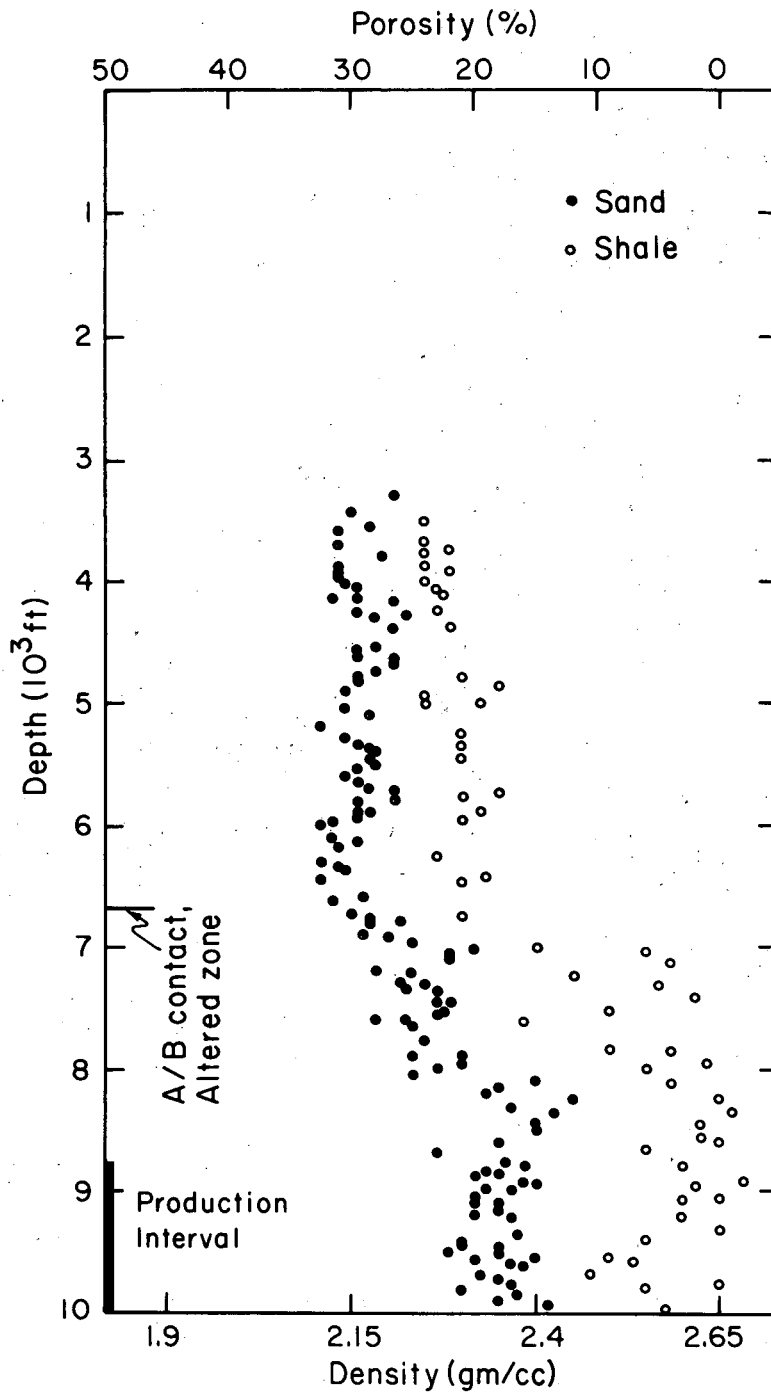


Figure 15. Density versus depth for a geothermal well at the Cerro Prieto geothermal field, Baja California. Note that the density values of both shale and sand are higher in the hydrothermally altered sediments (B) below the A/B contact than they are in the relatively fresh sediments (A) above the A/B contact. (XBL 802-6772)

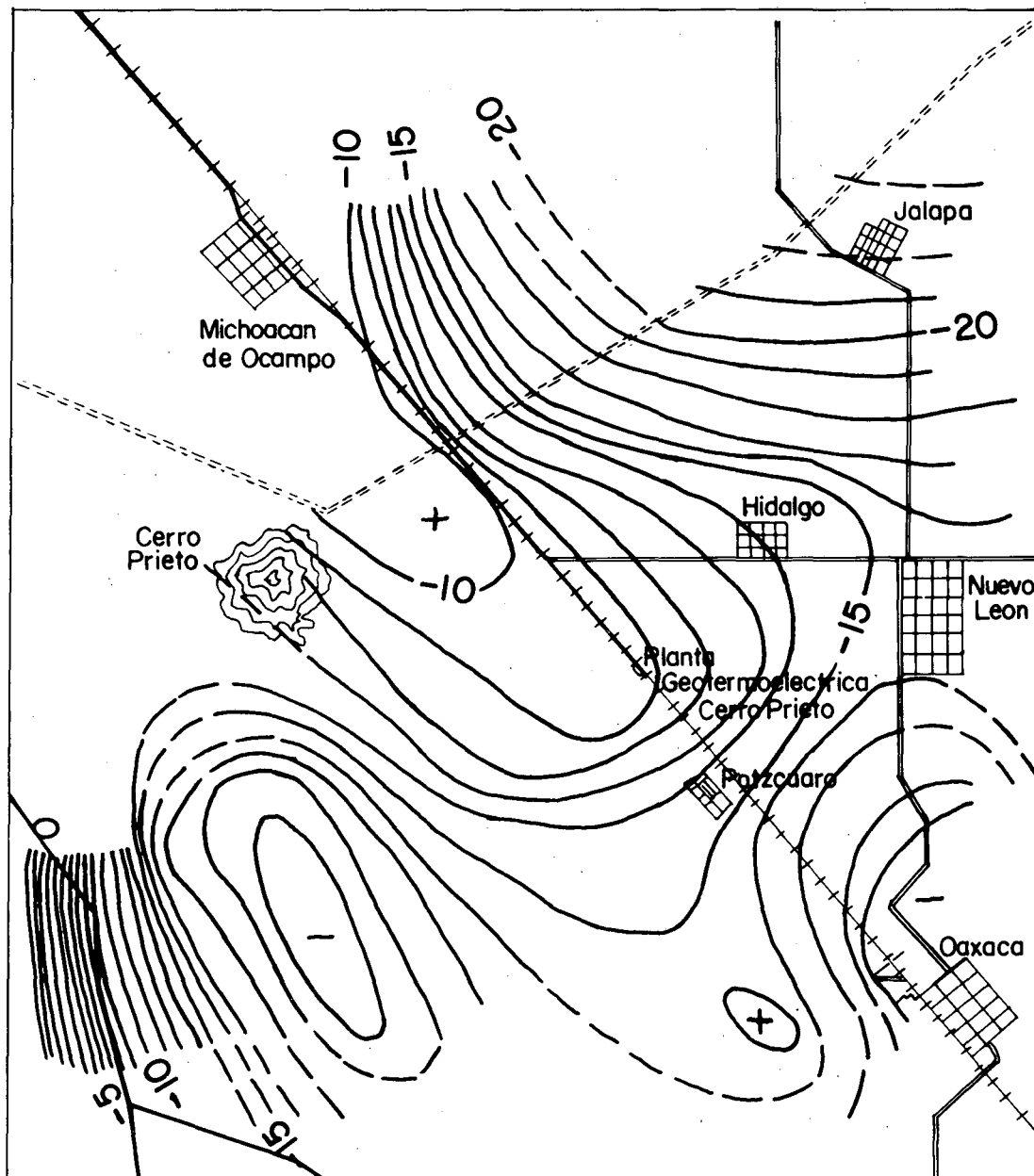


Figure 16. Depth (meters) to the top of the densified zone beneath the East Mesa area, Imperial Valley, California. The mean density contrast of 0.14 g/cm^3 between the anomalous source and the host rocks is attributed to authigenic quartz, sulfides, and calcite deposited as a result of thermal waters ascending along subvertical faults.
(XBL 865-10807)

Rymer and Brown (1986) reviewed the association of gravity features, both highs and lows, with volcanic structures. They reported that, to a first approximation, the gravity anomalies appear concentric with and are caused by sources within the edifices. Positive anomalies characterize many basaltic volcanoes; the sources are generally believed to be intrusions of shallow mafic dikes and associated igneous bodies into less-dense sediments and volcanics. This is the general explanation for a 27-mGal residual anomaly centered on the Medicine Lake Volcano (Finn and Williams, 1982). Supporting evidence for a young subvolcanic intrusive complex comes from electrical measurements that show a resistive body (Stanley, 1982).

Negative anomalies with amplitudes of up to -60 mGal and long wavelengths are associated with silicic calderas (Rymer and Brown, 1986). These anomalies are generally believed due to the large thicknesses of low density pyroclastic ash and pumice that fell back into the collapsing calderas together with later sediments that formed in the calderas. Close inspection of the gravity anomalies over the Long Valley caldera, California (Jachens and Roberts, 1985) and the Valles Caldera, New Mexico (Wilt and Vonder Haar, 1986) reveals that there are perturbations in the gravity that could be due to post-caldera intrusives, such as the resurgent dome.

There has been considerable interest in the use of gravimetry for locating large zones of magma or partially molten rock. Magma density is approximately 20% less than that of the solid rock of the same composition. This application was discussed by Chapman (1975) and Isherwood (1975), who independently interpreted regional gravity data from the Geysers-Clear Lake area. Isherwood (1975) reported that the 30-mGal residual gravity low near Mt. Hannah, 10 km north-east of The Geysers geothermal field, could be explained as a partially molten, silicic intrusive mass 10 to 13.5 km deep. He fitted the anomaly to a spherical body. Chapman (1975) processed the data differently and modeled the gravity source as a truncated cylinder, 3 km to its top and plunging to the southwest. Seismic-wave attenuation results by Iyer et al. (1979) seem to favor the shallower-source model, but the seismic results are somewhat ambiguous. The Geysers-Clear Lake gravity interpretations have prompted other scientists to conduct gravity surveys and/or to reinterpret existing data in search of large, low-density chambers. To date, no other cases have been reported in which the effects of a melt zone are believed to be discernable. The popular view of a magma chamber beneath The Geysers has not been borne out by deep electromagnetic soundings (Keller and Jacobson, 1983a). In general, the gravity low from a deep silicic magma chamber would not only be small but could also be difficult to differentiate from shallow lateral variations in rock densities.

ELECTRICAL AND ELECTROMAGNETIC

Although electrical and electromagnetic methods have been used widely in geothermal exploration for many years, the techniques suffered initially from a lack of well-documented case histories and problems in interpreting field data in terms of 2-D and 3-D structures. These problems and other controversies surrounding the application of electrical and electromagnetic techniques were brought out in a workshop held in 1976 on "Evaluation of Electrical Methods in the Geothermal Environment" (Ward, 1977). In the last few years considerable progress has been made to better understand the functional relationships between electrical measurements and subsurface thermophysical parameters through laboratory and field studies. Data-acquisition and interpretation techniques have also improved, but interpretation, both numerical and in terms of hydrogeologic models, remains a problem. Before discussing these techniques it is appropriate to review the principal factors that affect

rock resistivities and contribute to electrical-resistivity anomalies associated with geothermal reservoirs.

Factors That Influence Formation Resistivity

Water-saturated rocks may be pictured as two-phase systems consisting mostly of a high-resistivity ($> 10^4$ ohm·m) phase of silicate minerals (Telford et al., 1976) and a smaller volume of low-resistivity (< 0.1 to nearly 10 ohm·m) electrolyte. Because of this contrast, rock resistivity is mainly controlled by those factors which impede ionic conduction in the fluid phase; i.e., the degree of water saturation, the number and mobility of ions, and the connectivity of flow paths through the rock matrix.

In the absence of conductive minerals (e.g., most sulfides) or significant surface-conduction effects due to clays, resistivities of liquid- (electrolyte) saturated rocks are caused by ionic conduction through pores and are adequately described by an empirical relationship (Archie, 1942):

$$\rho_0 = F \rho_w,$$

where

- ρ_0 = formation resistivity,
- ρ_w = pore water resistivity, and
- F = the "formation resistivity factor."

This relationship is often expressed in terms of rock porosity as:

$$\rho_0 = a \rho_w \phi^{-m}, \quad (3)$$

where

- ϕ = porosity (expressed as a fraction),
- a = a number near unity for most porous rocks, and
- m = a number near 2 for porous rocks, and somewhat less than 2 for limestones, granites, or other fractured rocks with low-matrix porosity.

Archie's law (Eq. 3) was formulated for "clean" porous sandstones; i.e., an idealized rock. Various workers have attempted to extend Archie's law to the more realistic models where ion-exchange/surface-conduction effects exist or may predominate, where most ion conduction occurs in a fracture network, or where temperature effects and two-phase conditions may be present (Keller and Frischknecht, 1966; Worthington, 1982; Bussain, 1983). Considering a shaly-sand formation with surface conduction, Bussain (1983) provides a good review of physical models and proposes a rock model that at low frequencies (i.e., ignoring frequency-dependent conductivity) predicts a formation resistivity ρ_0 of

$$\rho_0 = \rho_w \phi^{-m} \left[\frac{1 - \rho_w / \rho_r}{1 - \rho_0 / \rho_r} \right]^{-m}, \quad (4)$$

where ρ_r is the effective resistivity of the rock matrix. In this model the empirical Archie "a" factor is given a more meaningful physical basis in terms of m and ρ_r . If matrix conduction is depicted as a volume cation-exchange phenomenon, one can express ρ_r as

$$\rho_r = (\Lambda_+ Q_v)^{-1},$$

where Λ_+ is the equivalent cation conductance ($S \cdot cm^2/g$ - equivalent), and Q_v is the concentration of cations (g - equiv/). Table 6 shows experimental values of ρ_r for rocks containing the common hydrothermal clay smectite (montmorillonite). Olhoeft (1983) has found that zeolites, a common class of secondary minerals that form from the devitrification of glass in volcanic rocks and the hydrothermal alteration of feldspars, can decrease bulk rock resistivity by a significant amount; the effect increases with temperature.

Of particular interest is the effect of temperature on the ionic conductivity of pore fluids. Ucock et al. (1980) determined the electrical resistivities of aqueous solutions with salinities and temperatures representative of waters found in geothermal areas (brine temperatures between 22 and 375 °C and concentrations of 3 to 26 wt%). Their data provide a more complete physical basis for the formation resistivities that may be detected or anticipated in geothermal areas. They found that, for the temperature range studied, water resistivity is related to temperature through a polynomial expression,

$$\rho_w = b_0 + b_1 T^{-1} + b_2 T + b_3 T^2 + b_4 T^3,$$

where T is temperature in °C and the b terms are temperature coefficients for a particular dissolved salt. The water resistivity also depends on the salt concentration. This is expressed by the formula

$$\rho_w = 10/\Lambda C,$$

where $\Lambda = B_0 - B_1 C^{1/2} + B_2 C \ln C + \text{higher-order terms}$, and C is the molar concentration. The B coefficients depend on solution chemistry. Figure 17 shows resistivities of NaCl solutions with concentrations typical of geothermal brine calculated from a regression analysis that fits both temperature and concentration parameters. These data show that the temperature effect is greatest below 150 °C, where increased ion mobility (lower viscosity) is the dominant factor. As temperature rises, changes in water density begin to offset ion mobility. The lower-density water results in a decrease in the dielectric permittivity of the solution and a decrease in the number of dissociated ions in solution. Above 300 °C the dielectric property of water decreases, causing increased association among ions and an increase in fluid resistivity (Quist and Marshall, 1968).

Figure 17 also shows the resistivities of low-temperature/less-saline ground or recharge waters (A) compared to high-temperature/moderately saline waters of a geothermal reservoir (B) such as Cerro Prieto. In the vicinity of a geothermal reservoir, water resistivities might be represented as a continuum of values, e.g., the area between the dashed lines. It is clear from Figure 17 why the resistivity methods are used widely in the search for liquid-dominated reservoirs. It also shows that water resistivities are not a particularly good indicator of reservoir temperature. A high-temperature reservoir with 3 wt% NaCl (area B) will have the same water resistivity as a low-temperature reservoir with 5 wt% NaCl (point C).

It is also clear from Eqs. (3) or (4) that porosity has a strong influence on formation resistivity. In nongeothermal areas resistivities of water-saturated rocks in the shallow crust (~ 3 km) typically vary over several orders of magnitude, roughly from 10 to over 1000 ohm·m: younger unconsolidated sediments are the least resistive, whereas crystalline rocks, with their lower porosities, have the highest resistivities. Furthermore, resistivities below a few hundred meters tend to increase monotonically with depth because of (a) normal compaction, diagenetic alteration and cementation of sediments and (b) increased confining stress,

TABLE 6 Effect of Smectite on Matrix Resistivity			
Smectite, %	Average ρ_r , ohm·m	Measured m (Eq. 4)	Number of samples
0	0.83	1.81 $\pm$ 0.25	16
20	0.17	1.05 $\pm$ 0.11	3
80	0.15	2.30 $\pm$ 0.26	4
100	0.10	2.56 $\pm$ 0.08	3

Data from Waxman and Smits (1968).

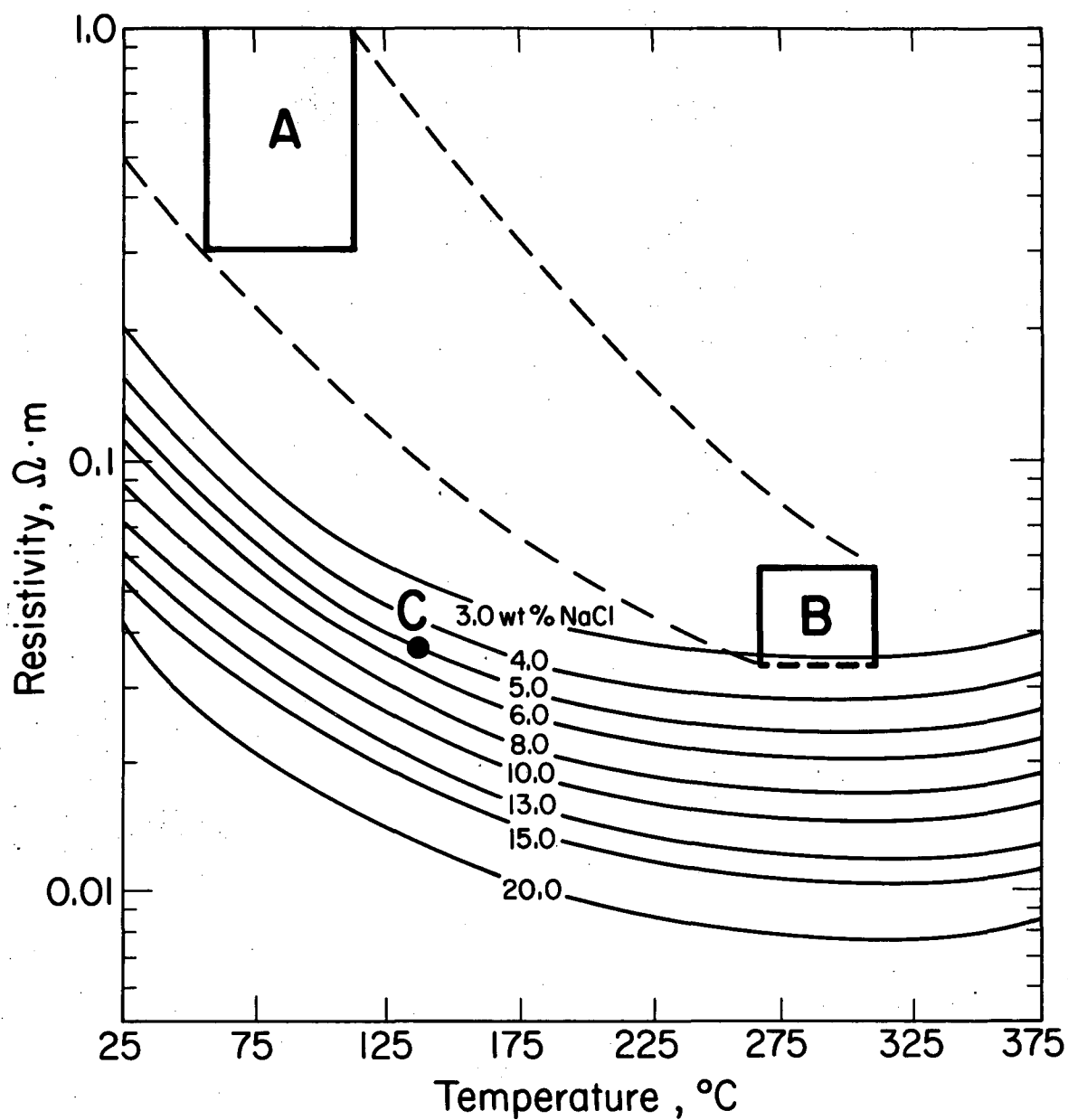


Figure 17. Resistivities of NaCl solutions as a function of temperature. Solution concentrations are typical of geothermal brines. See text for the explanation of A, B, and C (from Uçok, et al., 1980). (XBL 8312-2445)

which reduces fracture apertures. In higher temperature regimes the thermal annealing of microfractures may be significant. The model for rock resistivities in areas above a cooling magma can be quite different from this picture. The mechanical energy released during the expulsion of water and other volatiles from the hood zone of a H_2O -saturated melt can significantly increase fracture porosity in a broad region above and around the intrusive (Moskowitz and Norton, 1977). The resulting hydrothermal convection system would be expected to produce a measurable resistivity low above a magmatic body of sufficient size during part of its cooling history due to water-rock reactions that (a) enhance the salinity of fluids by leaching huge volumes of rock, (b) increase porosity by the dissolution of quartz and through a volume reduction of certain secondary minerals, and (c) increase surface-conduction effects by the creation of clays and zeolites on fractures and open pores. On the other hand, the mixing of cold groundwaters with hydrothermal fluids may cause local increases in rock resistivity as a result of mineral clogging along flow channels. Shallow, unfractured igneous dikes and plutons will also introduce a resistive element to the bulk resistivity of the region.

Electrical Resistivity in Geothermal Areas

Despite the various quantifiable and unknown factors that affect measured resistivities, it remains a generally accepted tenet of geothermal exploration that low resistivity in the proper setting is an indicator of a hot-water geothermal reservoir, such as at the Broadlands Field, New Zealand (Risk et al., 1970), and at the Kríuvík area, Iceland (Arnórsson et al., 1976). In both these areas, the hydrothermally altered and conductive ($\leq 10 \text{ ohm}\cdot\text{m}$) volcanic rocks extend to the surface and serve to outline the main reservoir region. However, as more detailed resistivity data have been made available, and as more attention has been directed to quantitative interpretations and correlations of results with reservoir conditions and processes, we have begun to recognize that a resistivity low may be a necessary condition for a geothermal environment but not a sufficient condition on which to site an exploration hole. Resistivity lows commonly occur near the surface in deeply weathered terrains and sediment-filled basins, particularly where either brackish or alkaline pore fluids or minerals with ion-exchange capacity (zeolites and smectite) exist. At Roosevelt Hot Springs an extensive shallow conductor was mapped that Ward et al. (1978) attribute in large part to secondary kaolinite and smectite clays in fractured granite; surface-conduction effects were estimated to be three times as important as ionic conduction of the brine. The Roosevelt Hot Springs resistivity anomaly is related to an active geothermal system, but fossil systems or concealed conductive, graphitic limestone or metasedimentary units in basement rocks could give similar results. Certain areas of the arid Basin and Range Province and the Salton trough pose special problems for electrical methods because of the large-thickness, low-resistivity rocks that mantle the surface to depths of hundreds of meters in places.

In geothermal areas the lowest resistivities may be associated with the discharge areas, which may be laterally displaced from the main high-temperature reservoir rock. At the Cerro Prieto geothermal field electrical-resistivity data were misinterpreted until a careful 2-D interpretation of the dc-resistivity data was made in conjunction with geophysical well logs and mineralogical studies of drill cuttings. The lowest resistivities ($< 2 \text{ ohm}\cdot\text{m}$) occur west of the field and are attributed to the hydrothermally altered surficial rocks in the discharge area and a thick sequence of marine sediments. Early exploration holes drilled into the low-resistivity region did not encounter high temperatures. On the other hand, the shallow reservoir at a depth of 1.1 to 1.4 km was not resolved as a resistivity low even though the reservoir rocks consist of porous, sandy, deltaic rocks. Instead, the reservoir appears

from surface measurements as a more resistive zone of 4 ohm-m. This is believed to be caused by a combination of (a) a large volume of thermally metamorphosed shales that enclose the reservoir sands and (b) pore plugging (cementation) near the periphery of the thermal dome as a consequence of cold, CO_3^{2-} -saturated waters coming into contact with the heated rock.

In spite of the problems and ambiguities imposed by past and present hydrology and fluid circulation, electrical and electromagnetic techniques rightfully remain important in the exploration for hot-water, porosity-controlled reservoirs. Their value in locating deep fracture-controlled or steam-dominated reservoirs is less certain because fewer case studies are available for comparison. Zhody et al. (1973) reported a resistivity high associated with a vapor-dominated zone in Yellowstone National Park. Onodera (1982) reported indications of more resistive rocks associated with a two-phase zone at the Onikobe Field, Japan. On the other hand, it is not conclusively established that vapor-dominated geothermal systems are distinguishable by the presence of a resistive zone. A steam or two-phase zone may be overlain by a more conductive condensate layer, as at Kawah Kamojang, Indonesia (Hochstein, 1976). Furthermore, the steam zone itself may contain sufficient free water to appear non-resistive. A few monolayers of water molecules adsorbed onto mineral surfaces may be sufficient to cause a vapor-dominated system such as the Larderello steam field to appear moderately conductive (Olhoeft, 1981).

Self-Potential

Among the low-cost geophysical surveying techniques, self-potential (SP) has received renewed interest because of the growing number of well-documented surveys showing correlation between SP and thermal anomalies. The SP method is based on the measurement of natural dc voltages of a few millivolts to over a volt that are generated by a variety of natural phenomena, including thermal gradients and fluid flow. One advantage of the technique lies in its apparent simplicity; one merely measures the potential difference (voltage) between two nonpolarizing electrodes with a small battery-operated digital voltmeter. Equidistant stations along a line may be surveyed using a leapfrog technique (fixed electrode separation), or an arbitrary station array may be measured using the total field survey, in which a base electrode remains fixed and a roving electrode is moved to each successive station. In spite of its simplicity, the SP voltage can be contaminated by various sources of noise, and this has often cast doubts over the technique. Corwin and Hoover (1979) and Ward and Sill (1982) discuss sources of noise and signal and specify proper field procedures for dealing with the noise problems.

The two SP sources related to geothermal activity are temperature gradients and pressure gradients (Corwin and Hoover, 1979). If a temperature gradient (ΔT) occurs across a volume of rock, a phenomenon explained as the differential thermal diffusion of ions in the pore fluid and of electrons and donor ions in the rock matrix will produce a voltage (ΔV), a process called the thermoelectric Soret effect.

Fluid flow through pores and fractures can also generate a voltage in the direction of flow from the interaction of the fluid with the Helmholtz double layer at mineral grains. The electrokinetic or streaming potential, as this effect is called, can be written in explicit form for flow through a capillary tube:

$$\Delta V = \frac{\rho \epsilon \zeta}{4 \pi \eta} \Delta P,$$

where ρ , ϵ , and η are electrical resistivity, dielectric constant, and viscosity of the fluid, respectively; ΔP is the pressure drop; and ζ is the zeta potential across the double layer. Streaming potentials arise because there is a separation of charge at the solid-liquid boundary. Negatively charged ions (anions) tend to remain more tightly bound to unbalanced charges at the mineral surfaces while the positive ions (cations) are freer to move with the fluid. At some distance from the boundary there exists a region where cations become free to move with the fluid; the electric potential at this point is the zeta potential.

In a series of laboratory experiments using crushed pure minerals and rocks to simulate porous media, Ishido and Mizutani (1981) confirmed the linear relation between streaming potentials and driving pressure, and investigated the effects of mineralogy, fluid pH and temperature on ζ . They found that ζ tends to be larger for quartz and quartz-rich rocks, increases with pH to pH of about 8, and then may exhibit variable behavior, depending on rock mineralogy, and that ζ increases with temperature due to H^+ desorption from the solid. They concluded that the streaming potential coefficient $\rho\epsilon\zeta/\eta$ around a geothermal/hydrothermal convection cell would be sufficiently inhomogeneous due to temperature gradients to cause measurable SP anomalies of the order of 100^+ mV at the surface. Experimental work and theory on the relationship between streaming potential and two-phase fluid flow seem to be scant and inconclusive.

Examples of SP anomalies observed over geothermal fields are shown in Figures 18 and 19. In general, the SP anomalies found in geothermal areas appear as elongate negatives and as dipolar anomalies with roughly equal-amplitude positive and negative segments. Figure 18 shows the elongate zone of negative anomalies associated with Leach Hot Springs and the Hot Springs Fault, Grass Valley, Nevada (Corwin, 1976). After data smoothing, there is a -50-mV anomaly around the hot spring and an elongate negative close to the surface expression of a fault believed to be a conduit for the thermal waters (Corwin and Hoover, 1979). Figure 19 is a smoothed SP anomaly over the East Mesa geothermal field (Corwin et al., 1981). The A zone, the axis of the dipole anomaly, coincides with the location of the hotter wells and a set of northwest-trending faults defined by well-log and seismic interpretations. In their mathematical model of the field, Goyal and Kassoy (1981) calculate an effective width of 230 m for a vertical fault feeding the reservoir. Corwin et al. (1981) estimate a conduit 300 to 600 m wide to explain the gradient of the dipolar SP anomaly. A similarly shaped but larger SP anomaly was found to coincide with the production zone of the Cerro Prieto geothermal field (Corwin et al., 1980; Fitterman and Corwin, 1982). Subsequent analysis of well logs revealed that the SP-anomaly axis correlates reasonably well with a concealed secondary fault, one of the north- to north-northeast-trending faults that may be important to the hydrothermal circulation system.

Nongeothermal electrochemical conditions may also produce anomalies similar to those described above. For example, conductive sulfide deposits and graphite produce negative anomalies of a few tens to hundreds of millivolts. These are usually explained in terms of two half-cell reactions: one in the zone of oxidation above the water table, the other in the reducing zone. The Woodlawn Orebody, Australia, produces a narrow negative with a maximum of -330 mV, while pyritiferous black shales in the same area cause asymmetric negatives of up to -250 mV (Cifali and Whiteley, 1981). Pyritiferous black shales are also believed to be responsible for a broad negative traceable for many miles in Buena Vista Valley, Nevada (Goldstein et al., 1976). Formational SP anomalies from conductive shales can often be recognized by their long strike length. The weathering of the hydrothermal mineral alunite to sulfuric acid is believed to be responsible for large SP anomalies at several known localities, including the Dome fault zone at Roosevelt Hot Springs (Sill and Johng, 1979).

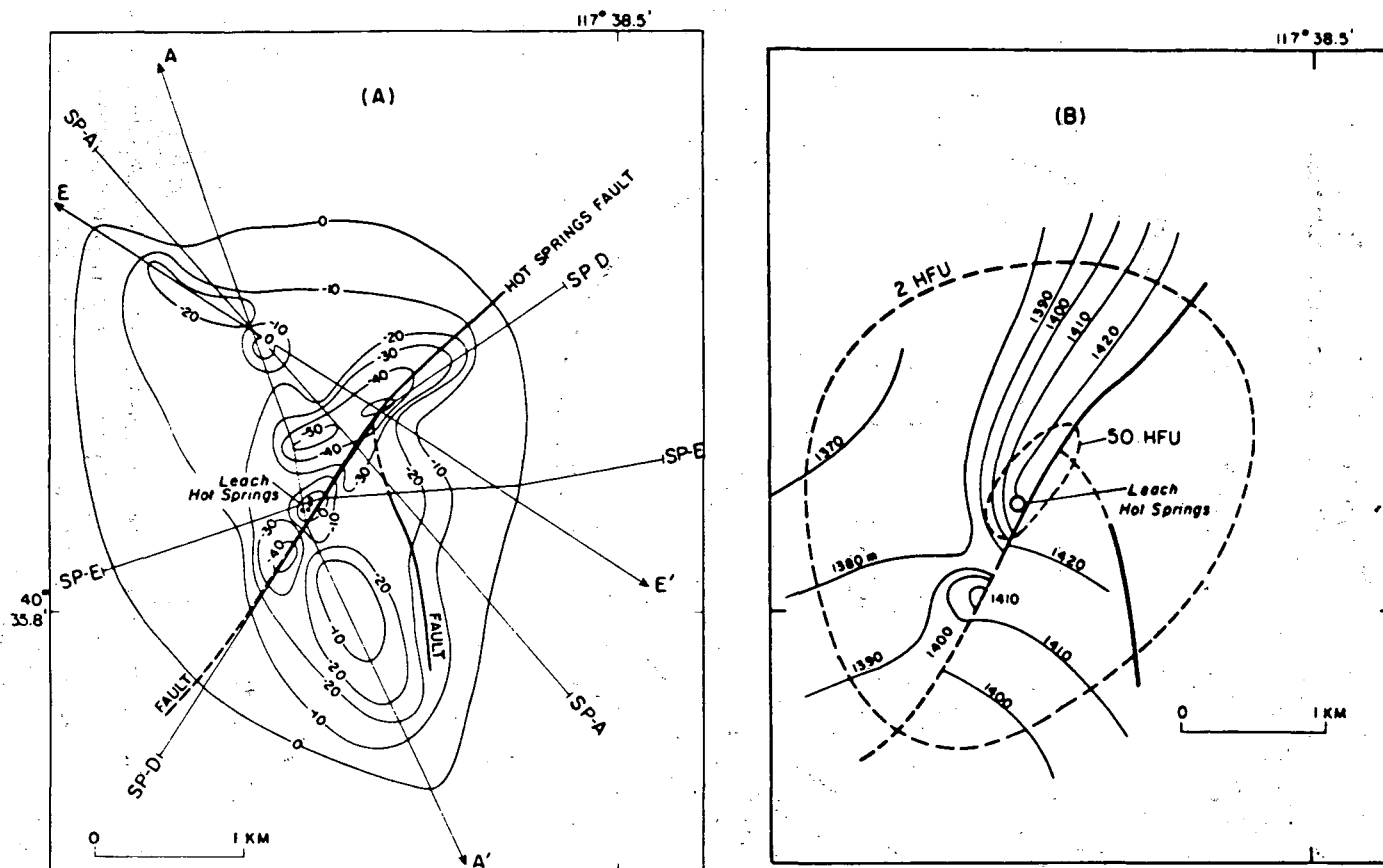


Figure 18. (a) Self-potential distribution in Leach Hot Springs area, Grass Valley, Nevada, based on smoothed data taken in September 1975. SP-A, SP-D, SP-E, A-A', and E-E' are traverse lines along which measurements were made. Electrode spacing was usually 100 m; contour interval is 10 mV. Faults dashed where inferred. (b) Heat-flow contours (dashed) and altitude of water table above sea level (solid lines) in the Leach Hot Springs area (from Olmsted et al., 1975). (XBL 833-8698)

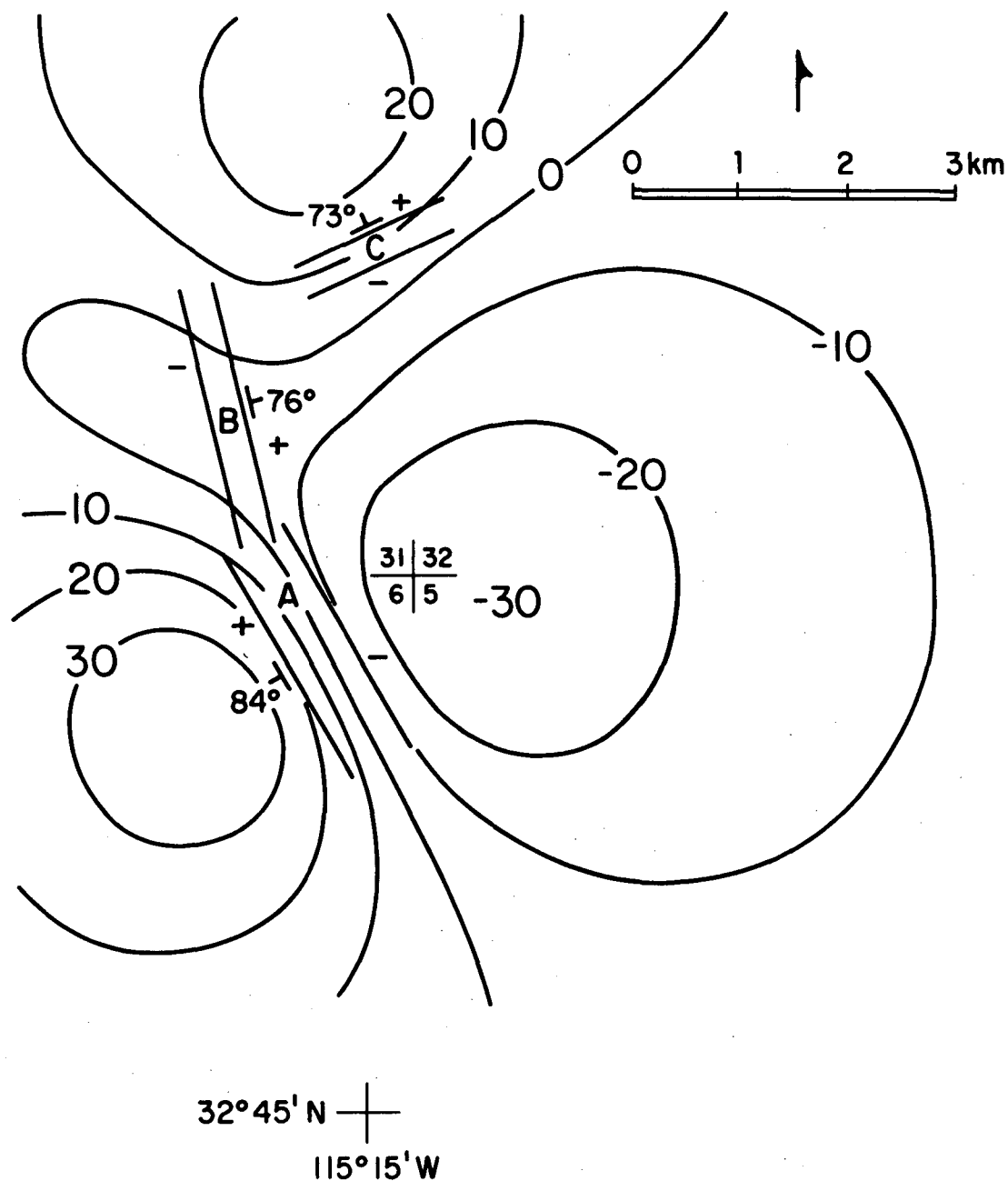


Figure 19. Smoothed self-potential anomaly over the East Mesa geothermal field, California. The axis of the dipole anomaly (A) coincides with the location of the hotter wells and a set of northwest-trending faults (from Corwin et al., 1981). (XBL 8312-2448)

Nonthermal groundwater flow in areas of high rainfall and steep topography may also cause an SP anomaly whose shape is an inverse of the topography and whose amplitude can be extremely large. Corwin and Hoover (1979) reported a -2700 mV anomaly over Mount Adagkak (645 m high) in the Aleutians.

In addition to the ambiguity of SP-anomaly shape/amplitude information and the multiplicity of SP sources, researchers have not until now developed effective techniques for modeling SP anomalies. Most early modeling was based on charge separations or current loops that, although they may have given results approximating observed anomalies, gave no physical insight into the source mechanisms. Fitterman (1979) and Sill (1983), expanding on the work of Marshall and Madden (1959) and Nourbehecht (1963), have developed modeling techniques based on coupled flows. Sample SP surface-voltage curves based on Sill's (1983) formulation are shown in Figures 20 and 21 for point sources, where ρ is the electrical resistivity, ρ_T is the thermal resistivity, ρ_P is the hydraulic resistivity or impermeability, and C is the voltage-coupling coefficient. This coefficient, which can be determined experimentally for laboratory models, relates forces (e.g., gradients of electric potential, pressure, and temperature) to current flow. The magnitudes of these voltage coefficients needed to fit field data are usually much larger than magnitudes found experimentally (Fitterman and Corwin, 1982; Sill, 1983) and may be larger because of a lack of good experimental data for geothermal-rock/fluid conditions. Figure 20 shows a symmetric negative anomaly caused by a point pressure source at a vertical resistivity boundary. Figure 21 shows a dipolar anomaly for a similar model but one in which the primary force is a point temperature source.

Corwin (personal communication, 1983) reports that not every known geothermal resource area has an associated SP anomaly. According to the coupled-flow formulation given by Sill (1983), the only nontrivial set of conditions giving zero surface voltage is a point thermal source in a homogeneous half-space where $C \neq 0$. All point-pressure sources will result in a surface voltage, as will the introduction of any vertical or horizontal boundary separating regions of different ρ , ρ_T , ρ_P , or C .

Sill (1982) also reformulated the cross-coupled fluid-flow problem in terms of fluid velocities and showed that a positive SP anomaly, as large as hundreds of millivolts, would occur over a plume of ascending hot water. The magnitude of such an anomaly would depend on, among other factors, the upward fluid velocity at the depth of maximum temperature gradient and the rock permeability. Such anomalies have been reported to occur in Japan, such as in the crater of the Yaakeyama volcano (Japan's Sunshine Project, 1984).

The Telluric Method

The telluric method determines subsurface resistivity structures by measuring the electric fields associated with flow of natural (telluric) currents in the earth. These currents result from the interaction between ionized gases from the sun and the earth's magnetic field. The phenomenon produces a broad spectrum of ultralow-frequency electromagnetic waves. The method has been used mainly as a low-cost method for exploring the structure of sedimentary basins (Kunetz, 1958; Yungul, 1977), and most of the pioneering work has been done in France, Germany, and the USSR since the mid-1930s.

The physical basis for the method and the techniques for data acquisition, processing, and interpretation will not be described in detail here. As the telluric method is not widely used in geothermal exploration, in contrast to the related magnetotelluric method discussed later, we shall limit discussion to a brief outline and examples of results.

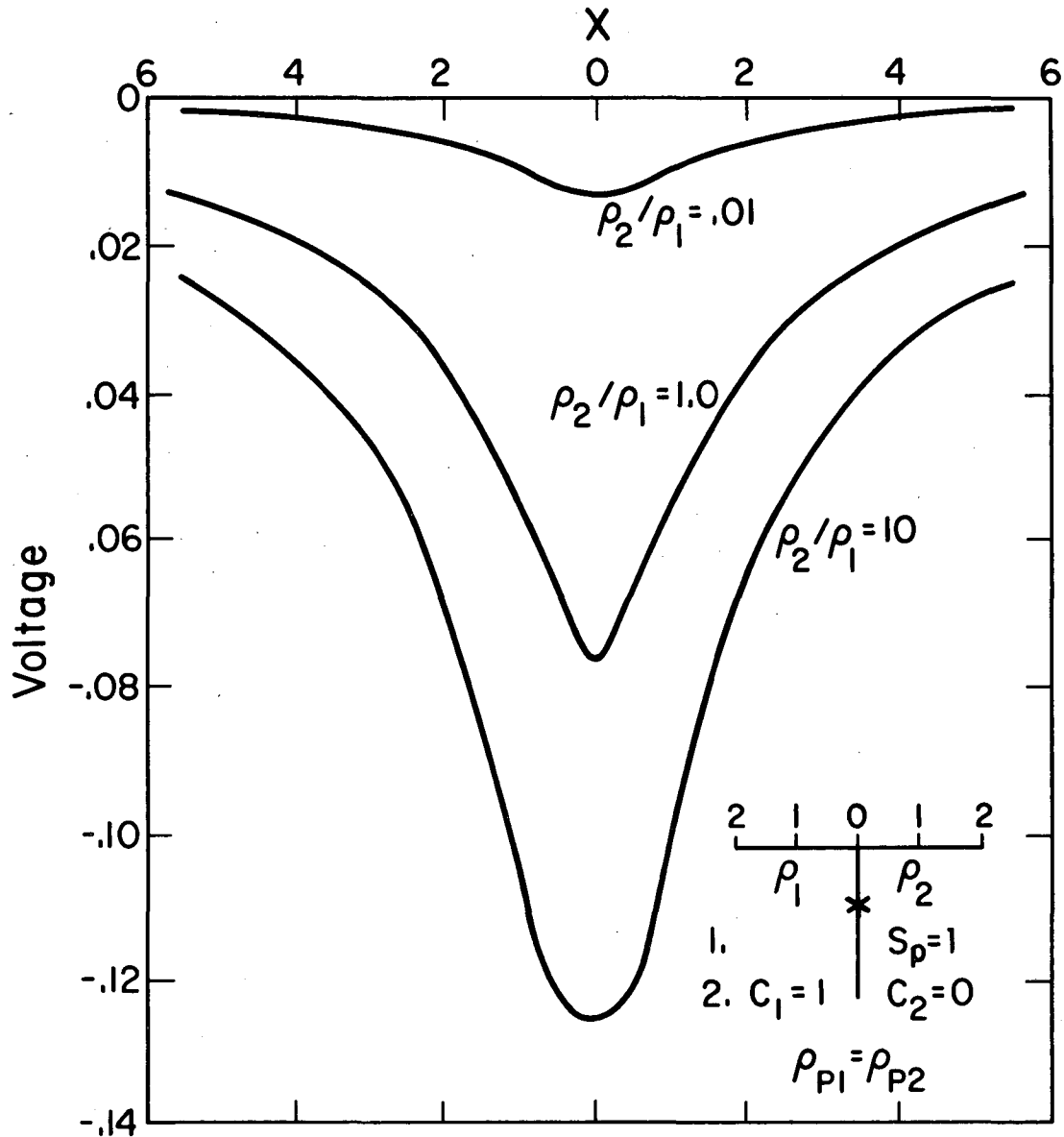


Figure 20.

SP surface-voltage curves of a symmetric negative anomaly caused by a point pressure source at a vertical resistivity boundary. ρ = electrical resistivity, ρ_P = hydraulic resistivity or impermeability, and C = the voltage-coupling coefficient, which relates forces to current flow and can be determined experimentally for laboratory models (from Sill, 1983).
(XBL 841-9514)

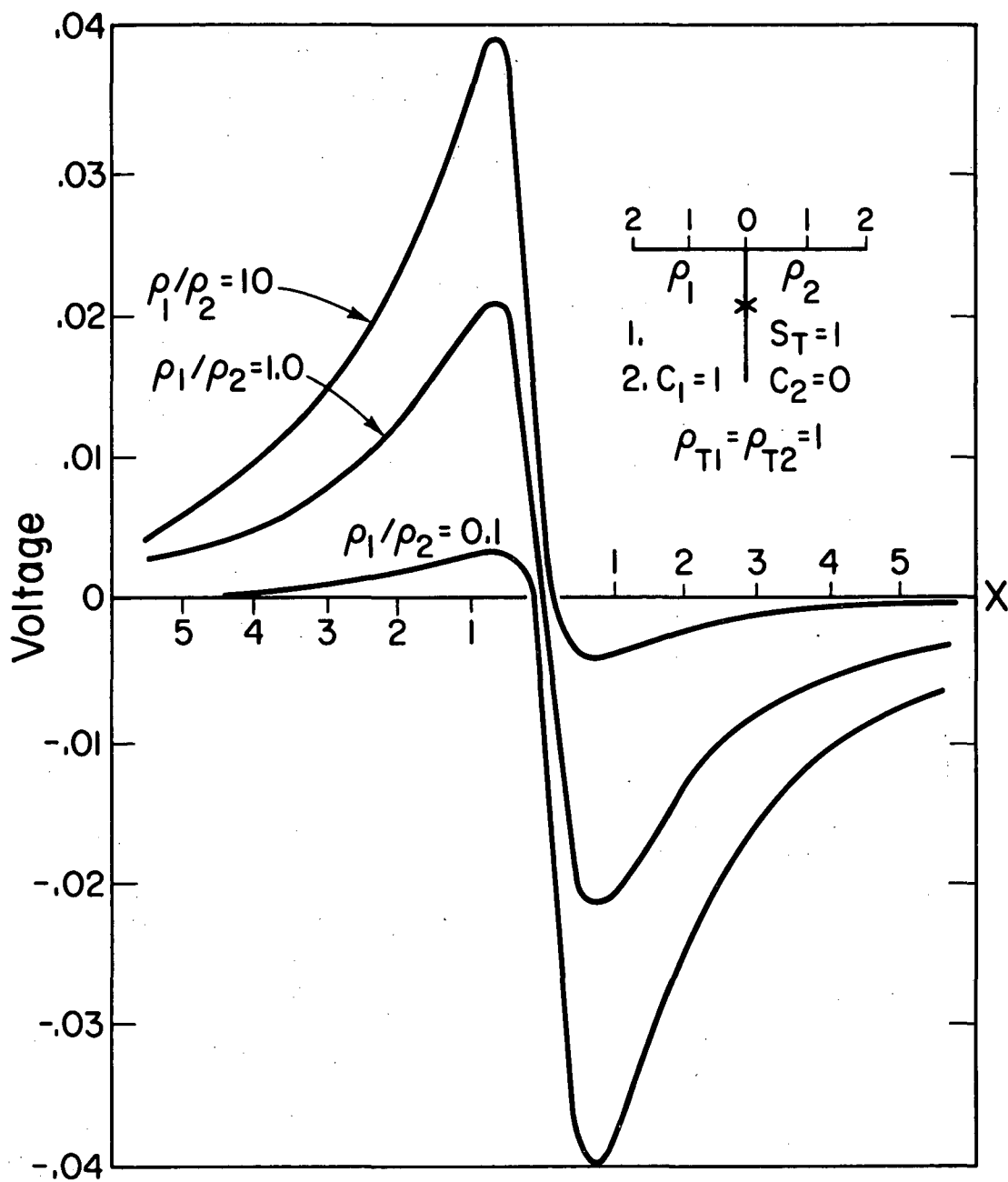


Figure 21. SP surface-voltage curves of a dipolar anomaly caused by a point temperature source at a vertical resistivity boundary. ρ = electrical resistivity, ρ_T = thermal resistivity, and C = the voltage-coupling coefficient, which relates forces to current flow and can be determined experimentally for laboratory models (from Sill, 1983). (XBL 841-9515)

Traditionally, one measures natural electric fields by grounded electric dipoles and appropriate filters and amplifiers at a minimum of two stations simultaneously. At a base station B, electric fields are measured along orthogonal directions x and y ; at the roving stations R, measurements are made along other directions u and v . If the earth can be approximated by an arbitrary n -layered rock sequence with layer thickness h_i and layer resistivities ρ_i beneath each station and underlain by a resistive basement, then the electric fields at the base and roving stations are linearly related by

$$E_u(t) = aE_x(t) + bE_y(t),$$

$$E_v(t) = cE_x(t) + dE_y(t).$$

Under certain conditions, a , b , c , and d , called the transformation coefficients, are real valued (i.e., there is no phase difference in the electric field at B and R); the coefficients depend only on the direction of the measuring directions and the differences in conductance S , where

$$S = \sum_i^n (h_i / \rho_i) \quad (5)$$

below the stations. Berdichevskiy (1960) showed that the transformation coefficients are real and independent of frequency of the electromagnetic wave so long as measurements are made within a particular range of frequencies called the telluric band or "S" zone. Depending on local conditions, this band usually ranges from 0.002 to 0.03 Hz, frequencies at which the variation from a purely dc response is small.

The transformation coefficients can be obtained from chart records, $x-y$ plots, or digital data in various ways (Yungul, 1977; Humphreys, 1978). The key feature of these coefficients is that the determinant of the Jacobian matrix, $J = |ad-bc|$, is related to the ratio of conductances S_B and S_R below the base and roving stations. Results may then be displayed by contouring J over the survey area. This provides a relative picture of how resistivity above the resistive basement varies over the area. If depth to basement is known from drill-hole logs and other geophysical data (e.g., seismic), then it is possible to convert the J map to a map of resistivity ellipses at each station, R, relative to a unit circle electric field at B. The ellipse orientation indicates the preferred current-flow direction relative to the base station; the ellipticity indicates the strength of this preference; and the ellipse area represents the J value at the location (Humphreys, 1978). To obtain some indication of how resistivities vary with depth as well as laterally around the base station, the analysis is made at several frequencies. The lower frequencies relate to more deeply penetrating currents. This is accomplished by bandpass filtering the electric-field signals around several peaks in the telluric spectrum.

Figure 22 shows a resistivity map derived from telluric J values over the Salton Sea geothermal field (Humphreys, 1978). In this example, an approximate relation between J and resistivity ($\rho = 1.23J$) was determined on the basis of known base-station resistivities. The resistivity low over the Salton Sea field, located near the top of the figure, correlates extremely well with a heat-flow anomaly (Lee and Cohen, 1979) that occurs around five small rhyolite domes extruded approximately 16,000 years ago (Muffler and White, 1969).

Despite some success with the telluric method, practitioners have found it to be somewhat slow and labor-intensive. For this reason, the "in-line" (also known as the electric-field ratio) telluric technique has been used for rapid electrical-resistivity reconnaissance.

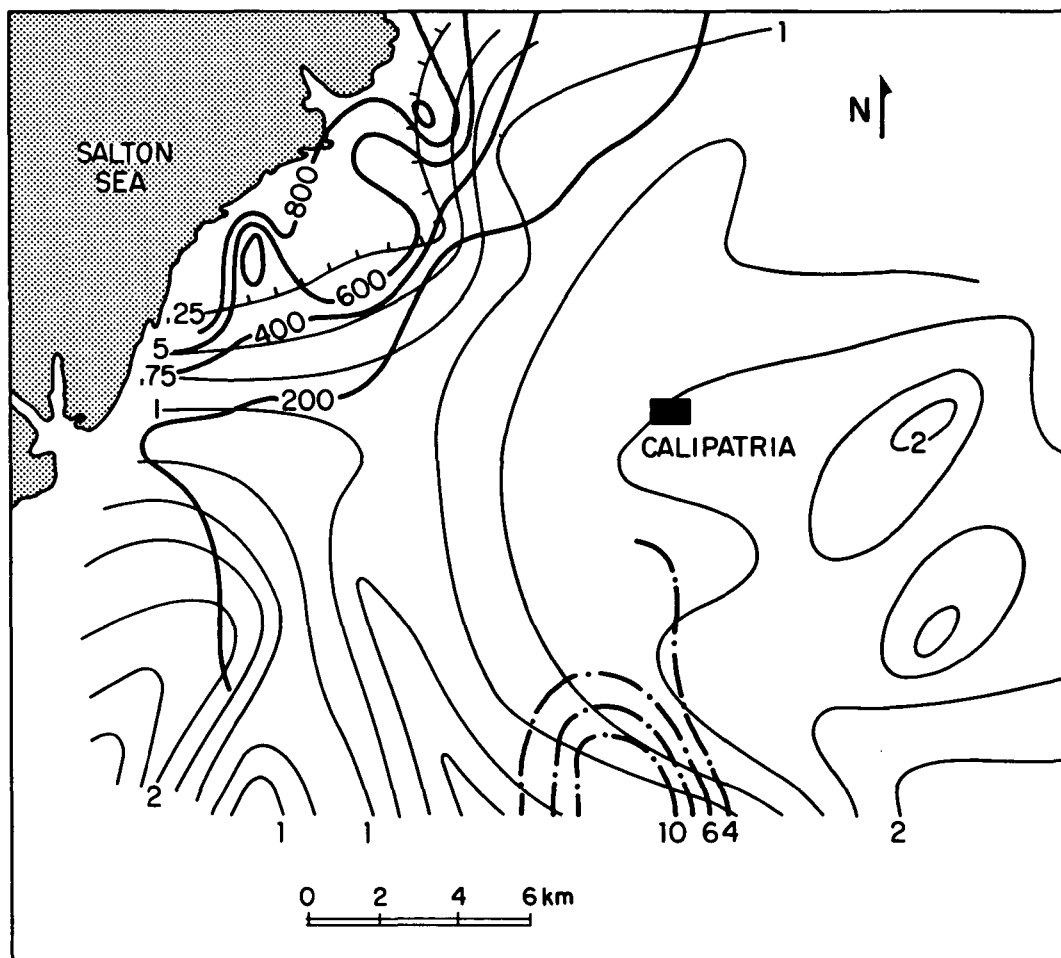


Figure 22. A plot of telluric J values at 67-second period (light solid lines) versus heat flow at the south shore of the Salton Sea geothermal field (after Humphreys, 1978). The heavy solid contours indicate heat flow in mW/m^2 (from Lee and Cohen, 1979). The dash-dot lines are temperature gradients in $^{\circ}\text{F}/100\text{ ft}$ (from Combs, 1971). (XBL 845-9785)

The In-Line Telluric Method

This method, devised by Neuenschwander and Metcalf (1942) and later elaborated by Dahlberg (1945), Yungul (1965), Yungul et al. (1973), and Beyer (1977a,b), involves measuring natural low-frequency telluric fields using an array of three collinear, grounded electrodes 250 to 500 m apart. The electrodes are placed along a traverse line to form two adjacent electrode dipoles, with the middle electrode serving as the common electrode. The signals detected by each dipole are amplified, narrow-bandpass filtered, and either fed to channels of an $x-y$ plotter for graphical display and later interpretation or fed into a digital signal processor for immediate numerical processing and display. Using either method, one derives the amplitude ratio of the electric fields, the phase difference between the electric fields, or the parameters of the ellipse created by the time-varying electric-field vector. The collinear array is leapfrogged along the survey line to give a continuous set of relative electric-field values. In the simplest form of data analysis, the successive ratios, each given by the slope of the quasi-linear curve on the $x-y$ plotter, are multiplied, as shown in Figure 23, to yield a relative amplitude profile of the electric-field component in the direction of the array. Maxima and minima indicate changing subsurface resistivities within a subsurface region. The depth of investigation is an inverse function of the frequency of the wave recorded. In practice, two frequencies that contain consistently high signal levels are studied; center frequencies of the bandpass are commonly set at 0.05 Hz for deeper probing waves and at the 8-Hz Schumann resonance for shallower probing waves. This approach provides a means for depth discrimination.

Carlston (1982) studied the technique at Roosevelt Hot Springs, where she found a broad but distinct resistivity low near the Opal Dome Fault. The relative amplitude curves at 0.05 and 8 Hz along one line studied in detail was consistent with the subsurface-resistivity model derived from detailed analysis of Schlumberger resistivity data. However, she also noted that, without prior knowledge, the electric-field telluric-ratio data would have been useful in only a qualitative way.

The technique does not readily permit quantitative interpretation but offers a low-cost way to find a buried conductor such as the 2-dimensional one shown in Figure 24. Difficulties in interpretation and anomaly resolution arise because small conductive inhomogeneities on the surface produce strong anomalies at all frequencies, and variations in overburden thickness and conductivity also produce strong anomalies, primarily at the higher frequencies (Beyer, 1977a,b). Telluric methods have been generally abandoned in favor of the magnetotelluric method described in a later section.

DC Resistivity and Induced Polarization

Detailed dc-resistivity surveys are usually conducted with one of the collinear, four-electrode methods: Schlumberger and dipole-dipole arrays (Figure 25) are most often used. An ultralow-frequency alternating current or a commutated direct current (I) is applied to the earth via a pair of current electrodes (C_1, C_2), and the voltage is measured between a pair of potential electrodes (P_1, P_2). The apparent resistivity ρ_a , that is, the resistivity that would be measured if the earth were a homogeneous half-space, is given by

$$\rho_A = \frac{V}{I} K,$$

where V is the voltage across the potential electrodes and K is a geometric factor that depends on the array. In resistivity profiling the separations between the electrodes (hence

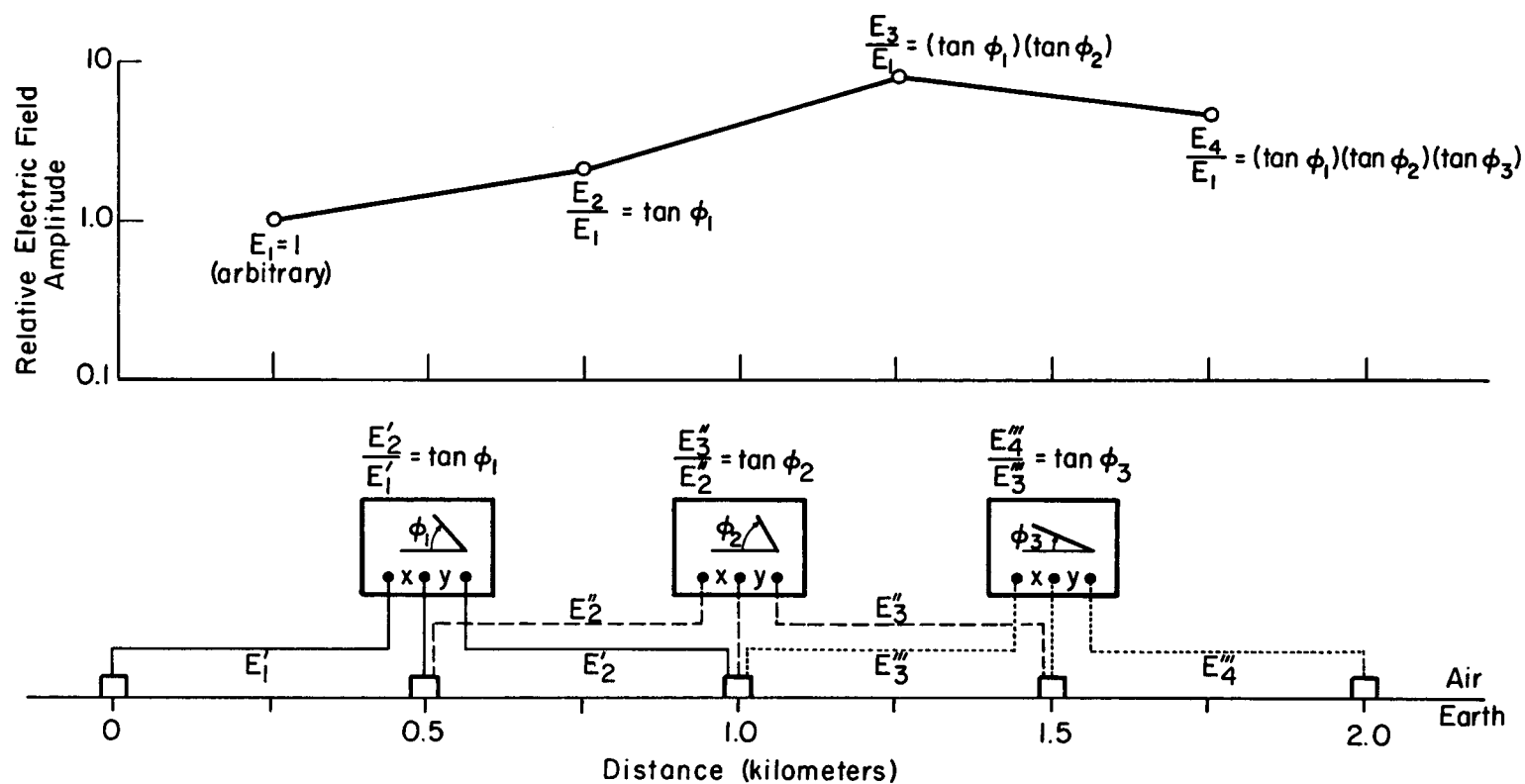


Figure 23. The in-line telluric method. Slopes indicate relative amplitude differences between points on the array. Maxima and minima indicate changing subsurface resistivities within the subsurface region being studied. The depth of the region is an inverse function of the frequency of the wave recorded (from Beyer, 1977a,b). (XBL 773-5229)

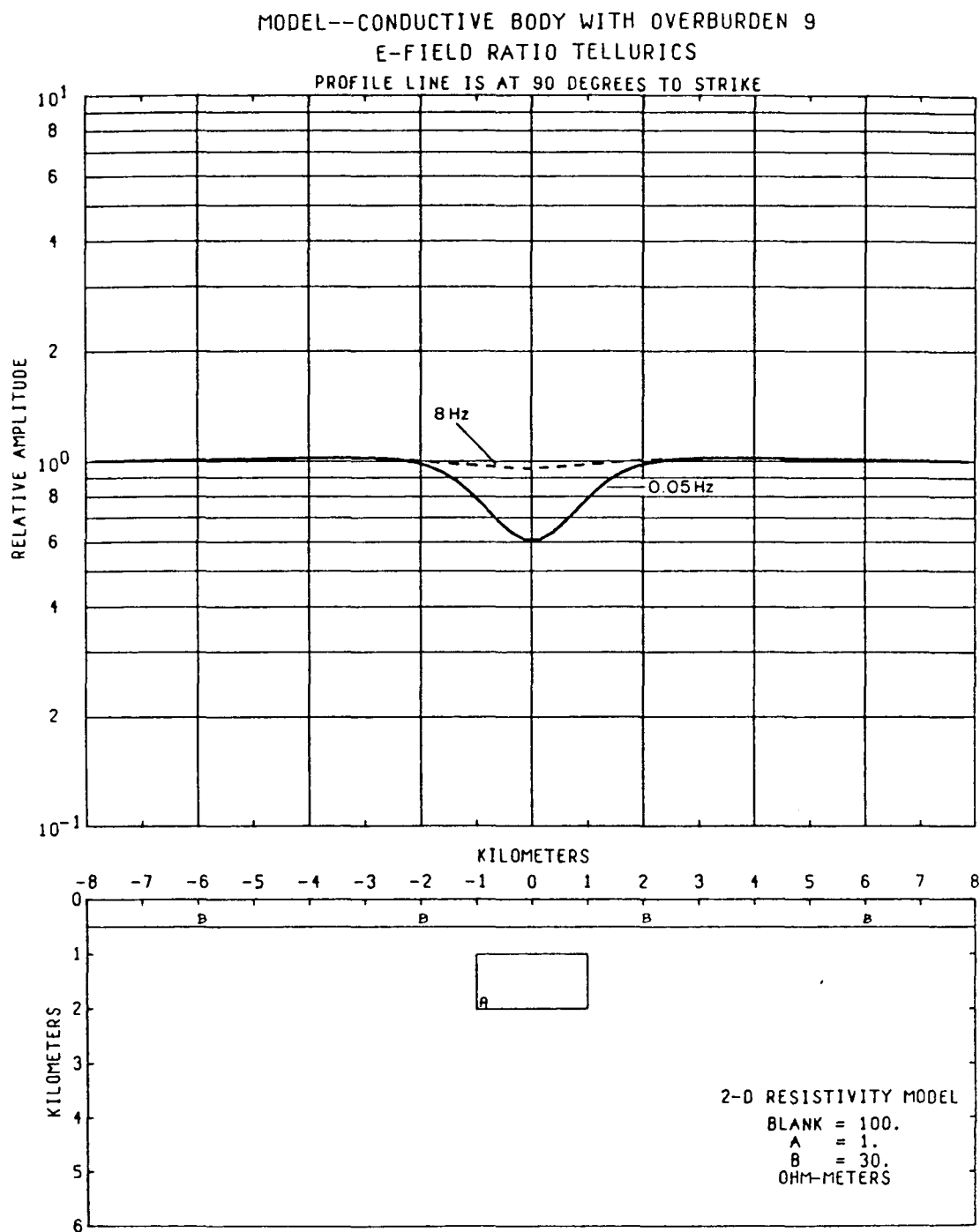
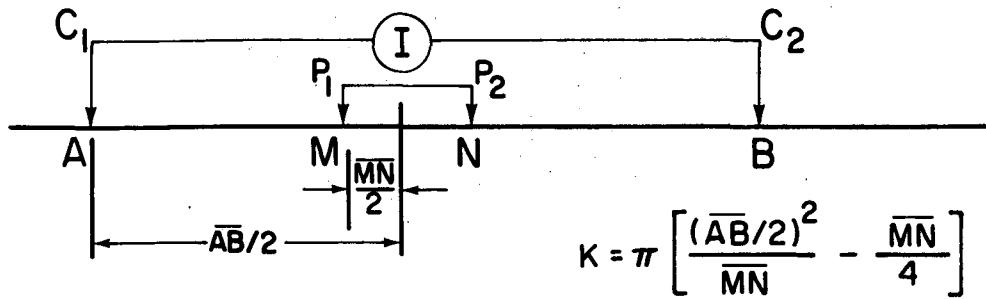
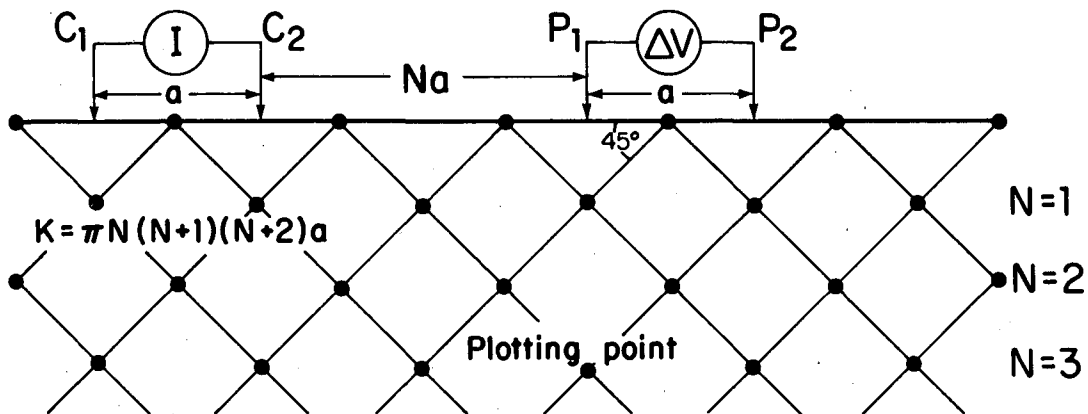


Figure 24. Calculated in-line telluric anomalies at 0.05 and 8 Hz caused by the concealed 2-D conductor beneath an overburden.
(XBL 833-8695)

SCHLUMBERGER ARRAY



DIPOLE - DIPOLE ARRAY



ROVING DIPOLE y (strike)

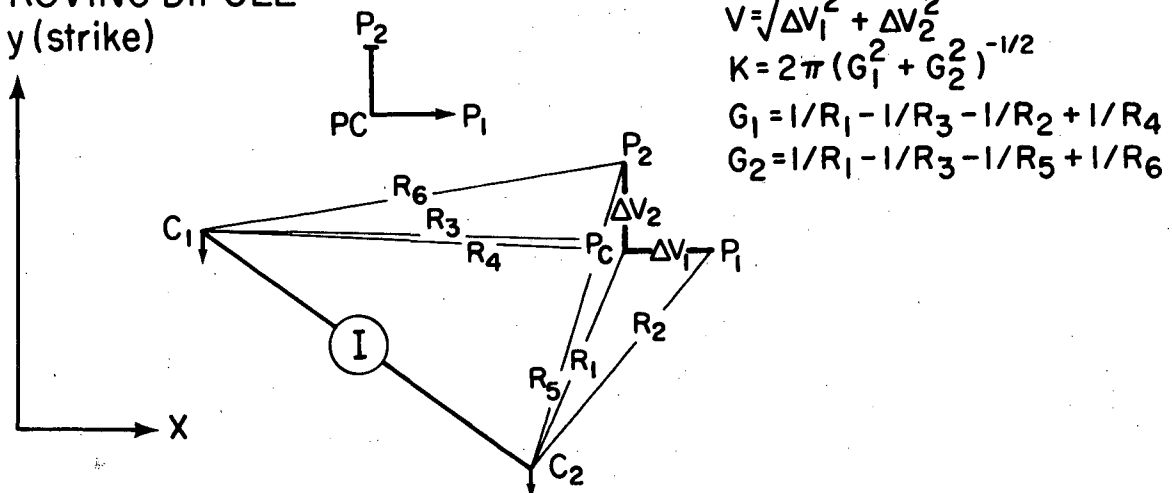


Figure 25. Diagrams of three common methods for conducting dc-resistivity surveys. (XBL 833-1740)

K) are kept constant, and the entire array is moved along the survey line in fixed increments. Profiling is used mainly as a rapid way to detect anomalies, and it seems to be a standard procedure in Iceland for detecting near-vertical, fluid-filled fractures (Flövenz and Georgsson, 1982). More commonly, the electrode separations are varied, and a set of apparent resistivity measurements is obtained at each station as a function of current-electrode separation, such as with Schlumberger expanders where $\overline{AB}$ is increased. Vertical electric soundings (VES) using the Schlumberger array are basic for obtaining a 2-D picture of subsurface resistivity.

Because the depth of current penetration, hence depth of exploration, is related to the $\overline{AB}$ distance (Roy and Apparao, 1971), current-electrode separations are expanded to as much as 10 or 20 km where deep exploration is required. Consequently, VES surveys in geothermal exploration are logistically more complex than those in shallow mineral and groundwater searches. Larger transmitters are needed and a greater weight of wire must be moved, therefore imposing greater demands on manpower and/or vehicles and imposing constraints on accessibility. Further, to measure the weak voltages accurately when current electrodes are far from the potential electrodes, synchronous detection and signal averaging are normally needed to pull the signal out of telluric noise.

Because of limited access, density of vegetation, rugged terrain, and high contact resistances, dc-resistivity surveys would be difficult to perform on the steep and densely wooded flanks of a Holocene volcano, such as in the High Cascade Range. However, Zhody and Bisdorf (1982) managed to perform a Schlumberger sounding in a forested and mountainous region with high contact resistance by placing electrodes along winding roads and correcting for the geometry of the electrode locations.

Besides impeding survey progress, local terrain may also cause a topographic anomaly that has a distressing resemblance to the geological anomaly sought. Although this undesirable effect has been well known to geophysicists for some time, Fox et al. (1980) systematically analyzed the terrain effect on the dipole-dipole array by a 2-D finite-element solution. They showed that a dipole-dipole line oriented perpendicular to a valley-ridge topographic sequence with uniform resistivity will result in a complex pattern of apparent resistivity highs associated with ridges and apparent resistivity lows associated with valleys. Topographic anomalies become important where slopes exceed 10 degrees for at least one dipole length. The severity of topographic anomalies in resistivity data may not have been properly appreciated in the past, because one could expect ridges to be more resistive than valleys, especially valleys that contain water-saturated fluvial deposits and running streams and those that mark fault traces and zones of hydrothermal alteration.

Schlumberger VES seems to be the favored resistivity method, particularly outside the U.S., because (a) the technique requires less wire handling than the other arrays and (b) interpretation can be done easily and without computers by using various sets of master curves and auxiliary point diagrams that provide a layered-earth (1-D) interpretation beneath the midpoint of the array (e.g., Compagnie Générale de Géophysique, 1963; Orellana and Mooney, 1966). While good results from this approach have been reported, there are several problems and limitations with interpreting multilayer sounding curves. First, there is the well-known problem of nonuniqueness: a sounding curve can correspond to a number of different combinations of layer thicknesses and resistivities. Nonuniqueness is related to the problem of equivalence: for certain type curves it is impossible to resolve separately the resistivity (ρ_i) and thickness (h_i) of the i th layer. Depending on the sounding curve, there can exist a range of values ρ_i and h_i such that $\rho_i h_i = \text{constant}$ (T equivalence) or $h_i / \rho_i = \text{constant}$ (S equivalence) will produce identical sounding results

(Battacharya and Patra, 1968). Other problems arise because the AB distance must be made large in relation to the depth of exploration desired. Not only does this require more wire handling, but it increases the probability that one of the current electrodes will cross a vertical resistivity boundary. This leads to another type of equivalence in which a vertical boundary can be misinterpreted as a deep horizontal boundary.

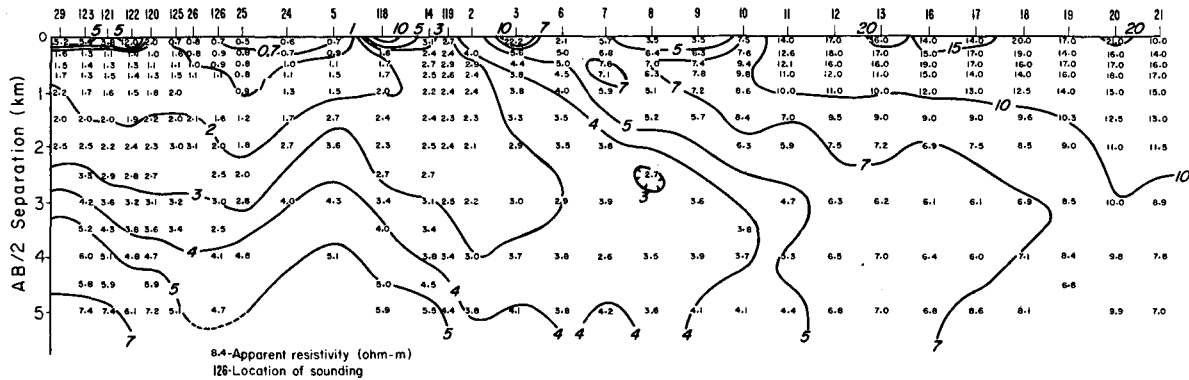
The points made above about the Schlumberger array are shared by all dc-resistivity techniques. Beyer (1977a) compared the Schlumberger and dipole-dipole techniques in terms of resolution, rejection of surface noise due to local inhomogeneities, and ease of interpretation. He concluded that the choice might depend on the target in relation to the geological environment. For locating an extensive, almost flat-lying reservoir, such as Ahuachapán, El Salvador, the Schlumberger technique was useful because the necessary information could be obtained with a low station density. However, in the Basin and Range Province, where narrow, near-vertical structures are sought, dipole-dipole may be more diagnostic.

Layered (1-D) interpretations are obtained efficiently today using numerical inversion codes that run on desktop computers (Zhody, 1973). When more accurate results are needed, 2-D modeling codes for mainframe computers are used. Wilt et al. (1978) used a finite-difference program described by Dey and Morrison (1977) to interpret data over the Cerro Prieto geothermal field (Figure 26). Tripp et al. (1978) used a 2-D transmission surface program to interpret data over Roosevelt Hot Springs geothermal field. In both cases, the use of the 2-D codes was essential to obtain better information on subsurface resistivities and to understand the hydrogeology and the hydrothermal alteration. However, the use of 2-D forward-modeling codes to fit the field data requires considerable effort and experience on the part of the interpreter. For this reason a number of iterative inversion programs have been written to find a 2-D resistivity model that fits the data in a least-squares sense (Sasaki, 1982; Tripp et al., 1984).

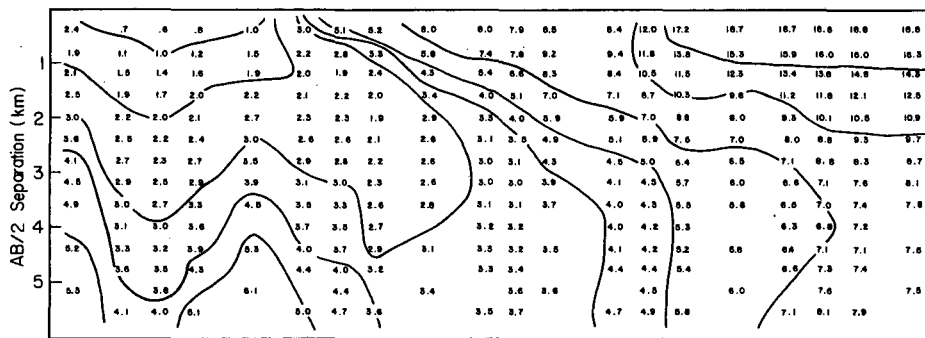
The roving dipole or bipole-dipole is another technique that enjoyed a period of popularity as a reconnaissance technique (Figure 24). One to three very long (2- to 5-km) grounded-current dipoles, called bipoles, are laid out over the area of interest. Potential differences are measured by the short receiver dipoles, often orthogonally paired, which are moved rapidly from station to station (Keller et al., 1975). By convention, the calculated apparent resistivities are plotted at the receiver station, and the data are contoured. The technique was useful in rough terrain and for mapping near-surface resistivity discontinuities, such as the Broadlands geothermal field (Risk et al., 1970). However, it was discovered that the data from one bipole source were not easy to interpret. Dey and Morrison (1977) analyzed the roving-dipole method and concluded that, to extract useful information in areas of 2-D resistivity distributions, one would have to employ a number of bipoles, thus obtaining an information density approaching that of a dipole-dipole survey. They showed that bipole-dipole results are sensitive to bipole orientation and that false anomalies are generated near vertical contacts. They concluded that selective dipole-dipole lines are preferred, even for reconnaissance.

To circumvent the problem of bipole orientation, various methods using two orthogonal bipoles were advocated (Tasci, 1975; Doicin, 1976; Bibby, 1977; Keller and Furgerson, 1977). Harthill (1978) employed a "quadripole" technique over a 2300-km² area of the Imperial Valley, California. Two source bipoles were sequentially energized with a square-wave electric current of between 200 and 400 A. The resulting electric fields were measured by orthogonal wire detectors; the detector sensed resultant voltages ET_1 and ET_2 , which were then combined to produce a resultant voltage ET . By changing the ratio of ET_1 to ET_2 , the resultant ET was caused to rotate through 360 degrees, tracing out a field ellipse. The arithmetic

OBSERVED RESISTIVITIES- SCHLUMBERGER EXPANDERS



CALCULATED RESISTIVITIES



2 D RESISTIVITY MODEL

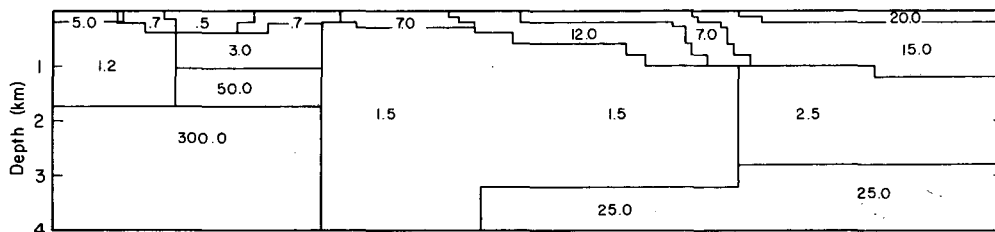


Figure 26. Observed and calculated dc resistivities and 2-D resistivity model of the Cerro Prieto geothermal field, Baja California. The observed resistivities were obtained by Schlumberger expanders. The calculated resistivities were generated by a finite-difference model (from Wilt et al., 1978). (XBL 788-1633)

mean of the maximum and minimum axes of the ellipse gave an apparent resistivity that had the desirable property of being independent of the bipole orientations. However, because of lateral changes occurring in resistivity, apparent resistivity values for a station were usually different, depending on the location of the bipole pairs. To resolve effects of lateral-resistivity inhomogeneities, Harthill (1978) had to perform data averaging so that the results could be contoured. The resulting resistivity map of the Imperial Valley showed a belt of low resistivity running north-south from the Salton Sea field to the Heber field. The results may correlate with groundwater salinities, temperatures, and hydrologic effects of freshwater runoff from adjacent ranges, as well as with underflow of Colorado River water and irrigation waters. However, this complete picture has not been put together.

Related to dc resistivity is the complex resistivity method in which the ground is energized over a frequency range from nearly dc to several hundred Hz, thus permitting one to extract information on the induced polarization (IP) effect. The IP effect manifests itself when a low-frequency alternating current passes through saturated rocks containing metallic mineral grains and certain minerals (particularly clays and zeolites) with unbalanced surface charges. The current flow is blocked by electrochemical or polarization forces at the electrolyte-mineral interfaces and this causes the voltage amplitude and phase measured between potential electrodes to vary with the frequency of the applied current (Figure 27). The resulting amplitude and phase spectra of the earth's impedance are often displayed in an Argand diagram (Figure 28), which has the form of the dispersion proposed by Cole and Cole (1941) to describe the behavior of complex dielectrics. Pelton et al. (1978) showed that IP parameters also fit a modified Cole-Cole dispersion:

$$Z(\omega) = R_0 \left[1 - m \left(1 - \frac{1}{1 + (i\omega\tau)^c} \right) \right], \quad (6)$$

where

$Z(\omega)$ = the complex impedance of the earth,

R_0 = the dc-resistivity value, $Z(0)$,

m = an IP parameter called chargeability (not related to the cementation factor in eqs. 3 and 4),

τ = a time constant,

c = a positive value near 0.25 for most rocks,

ω = angular frequency, and

i = $\sqrt{-1}$.

In this idealized but very useful representation of rocks with polarizable minerals, the chargeability m is as defined by Seigel (1959) and is limited to the range $0 \leq m \leq 1$. Because $|Z(\omega)|$ must decrease monotonically with frequency, the value of the exponent c is limited to the range $0 \leq c \leq 1$.

Complex resistivities measured in the field can often be fitted to a function given by Equation (6) to yield the polarization parameters of the earth (m , c , and τ) and the dc resistivity R_0 . However, complex resistivity dispersion is strongly affected by grounded fences and pipelines and by a physical effect called inductive coupling. Because cultural effects are extremely difficult to separate from geological effects (Sumner, 1976; Nelson, 1977), attempts are made to place survey lines or electrodes either normal to cased drill holes, gas pipelines, fences, etc. or far enough away to minimize their effect. Inductive-coupling (IC), an electromagnetic effect, is due to the frequency-dependent mutual inductive

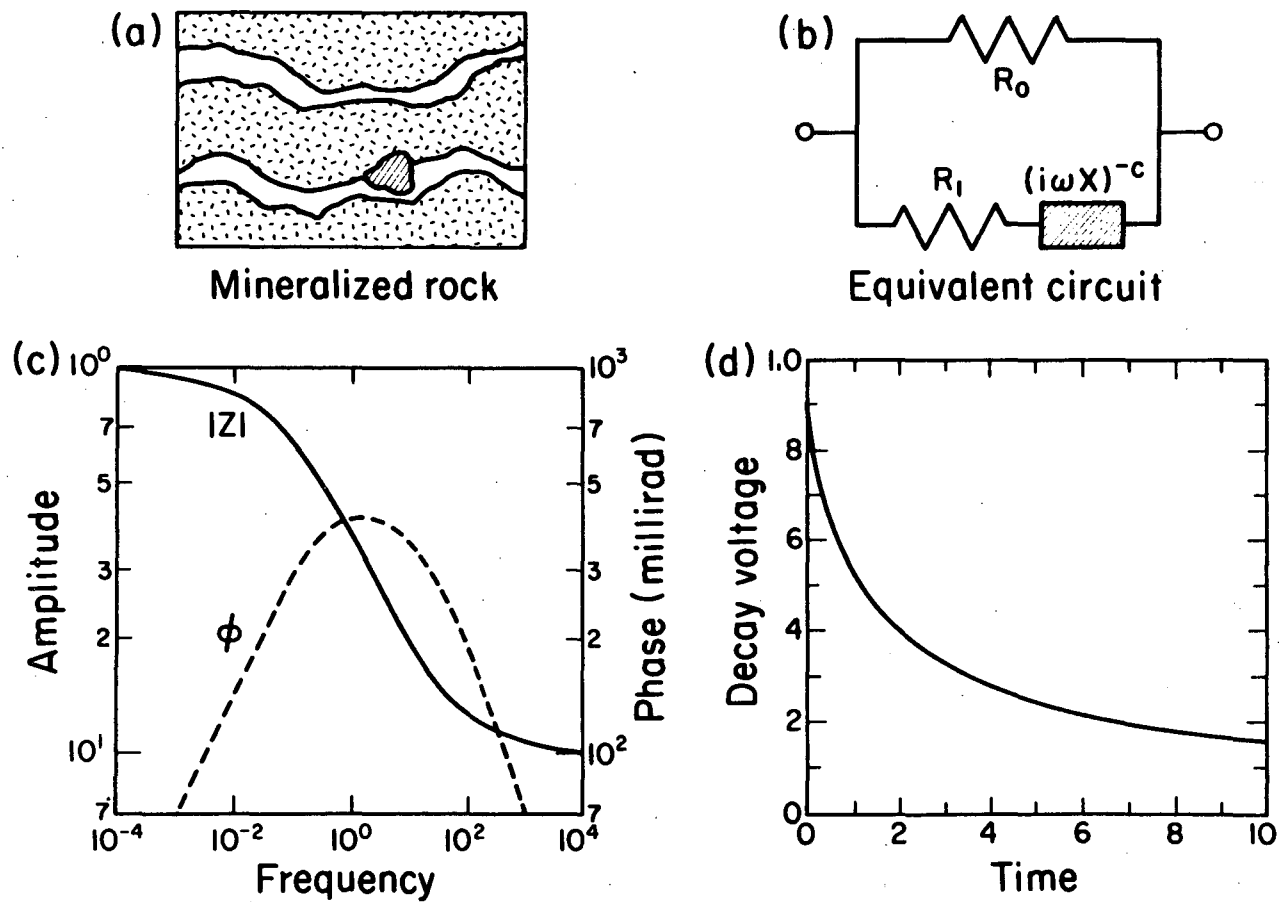


Figure 27. Examples of frequency domain spectra and time domain decay voltages caused by the presence of conductive sulfide grains along ionic flow paths. (XBL 8312-2446)

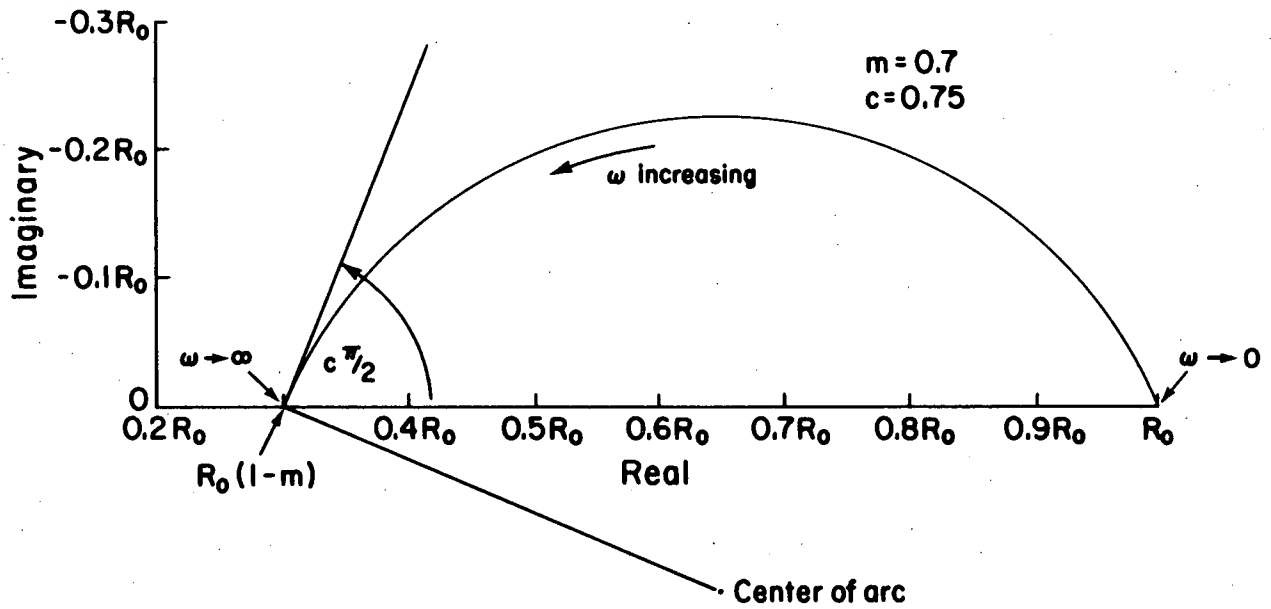


Figure 28.

Argand diagram of the voltage measured in Figure 24c. R_0 = dc resistivity value, $Z(0)$; m = chargeability parameter; c = frequency-dependent term; ω = angular frequency.
(XBL 8312-2447)

impedance between transmitter and receiver terminals. IC is particularly evident where the surface layer is conductive, and it increases with frequency and the separation between electrodes. The IC-dispersion effect can sometimes be separated from true polarization dispersion because of differences in the time constants (Wynn, 1974; Wynn and Zonge, 1975, 1977; Pelton et al., 1978; Major and Silic, 1981).

IP has been used infrequently in geothermal exploration, possibly because IP results have not been encouraging. On first look, IP would seem to be a promising technique, because polarizable minerals such as pyrite, clays, and zeolites all commonly occur in hydrothermal-geothermal systems. In practice, IP surveys have not provided any special exploration information, possibly because pyrite and zeolites are too common and widespread in volcanic rocks and because these polarizable minerals concentrate in the near-surface, lower-temperature fringes of hydrothermal-geothermal systems. Moreover, laboratory studies show (Keovil and Ward, 1962; Olorunfemi and Griffiths, 1985) that saline pore water (e.g., $\text{NaCl} > 10^3$ ppm) in sandstones decreases the IP effect. Rocks saturated with geothermal brines may have only very small IP responses. Chu et al. (1979) found a higher-frequency (> 1 Hz) IP effect at Roosevelt Hot Springs that they attribute to clays, and some lower-frequency effects that could be from pyrite. Risk (1981) investigated an IP response over the Broadlands field that he traced to a near-surface sand layer. Curiously, petrologic examination of cores from this unit did not reveal any minerals capable of producing the observed IP effect.

Electromagnetic (EM) Techniques

All dc-resistivity arrays share common shortcomings: (a) topographic corrections are needed where local relief causes large vertical separations between electrodes, (b) the method is inherently more sensitive to resistive zones than conductive ones, (c) the lengths of wires needed are very long compared to the depth of exploration, and (d) the conductive overburden will greatly suppress the anomaly from a deeper conductor. Because of inherent limitations (b) and (d), there are two geological situations for which dc-resistivity techniques will not always work well enough to give useful information:

- (a) a deep conductor beneath a thick section of resistive surface rocks, and
- (b) a conductor in a resistive basement overlain by a thick conductive surface layer.

To ameliorate these problems, the magnetotelluric (MT) method and, to a lesser extent, the related, higher-frequency variation called audiomagnetotellurics (AMT) have gained wide acceptance for geothermal exploration. MT surveys are often used in subregional-scale reconnaissance, but there are also applications for MT/AMT surveys on the detailed scale with station separations of 0.5 to 1.0 km. Details on the techniques are given by Vozoff (1972) and Kaufman and Keller (1981). For these natural-field techniques, ultralow-frequency electric and magnetic fields arising from ionospheric and magnetospheric currents plus higher-frequency fields from worldwide lightning discharges provide the inducing fields (see the section on the telluric method). By measuring magnetic as well as electric fields, one can obtain an impedance function that is diagnostic of the resistivity distribution near the station.

MT data are typically collected over the frequency range of 0.002 to 200 Hz, and AMT data are collected over the overlapping range of 8 Hz to 10 kHz by a field detection system similar to that discussed in the section on controlled-source electromagnetics. Modern systems employ either three-component superconducting quantum interference-device (SQUID) magnetometers or low-noise induction-coil magnetometers (Stanley and Tinkler, 1982;

Morrison et al., 1984), and data are gathered with an in-field processor and a remote reference for noise cancellation (Gamble et al., 1979).

As originally described by Cagniard (1953), MT surveys require the measurement of a single component of the horizontal magnetic field (H_y) and the orthogonal component of the electric field (E_x), which are assumed to be components of a plane wave and which are related by the earth impedance ($Z(\omega)$) at the measuring site:

$$E_x(\omega) = Z(\omega) H_y(\omega).$$

The measuring directions x and y are arbitrarily chosen. For practical purposes, this expression can be recast in terms of a frequency-dependent scalar apparent resistivity, ρ_a ,

$$\rho_a = \frac{1}{\omega\mu} \frac{|E_x|^2}{|H_y|^2} = 0.2T |Z|^2 \text{ ohm}\cdot\text{m},$$

where E_x is in mV/km, H_y is in nT, $f = \omega/2\pi = T^{-1}$, and the magnetic permeability μ is that of free space ($4\pi \times 10^{-7}$ H/m). Depth of exploration is usually expressed in terms of the "skin depth" of the EM wave; i.e., the depth at which a plane EM wave attenuates to $1/e$ of its amplitude at the surface of a homogeneous half-space with resistivity ρ (in km):

$$\delta = 0.5T\rho. \quad (7)$$

From the above expression it is clear that AMT frequencies have a limited depth of exploration in conductive terrains, such as many geothermal areas, where normal surface resistivities are only 2 to 10 ohm·m (Hoover et al., 1978). The AMT signal in the midrange (about 2 kHz) becomes weak in the northern hemisphere during the winter, when thunderstorms are farther away. Hoover et al. (1978) report that AMT can be conducted in all seasons up to 200 Hz, an acceptable limit because the higher frequencies have little depth of exploration. MT signals are usually very low in the 0.1- to 1.0-Hz band, a serious drawback to MT before the remote-reference technique for signal enhancement was introduced (Gamble et al., 1979).

Except within broad sedimentary basins, lateral resistivity inhomogeneities occur, and the simple scalar relationship between E_x and H_y is invalid. That is, Z usually depends on the measuring directions. To extend the MT method to more realistic geological conditions, Cantwell (1960) demonstrated that one should measure two orthogonal components of both horizontal fields (E and H) and to describe the impedance as

$$E_x = Z_{xx} H_x + Z_{xy} H_y, \quad (8)$$

$$E_y = Z_{yx} H_x + Z_{yy} H_y.$$

Equation (8) is usually shown in tensor notation as

$$\vec{E} = \underset{\approx}{Z} \vec{H},$$

where the impedance tensor

$$\underset{\approx}{Z} = \begin{vmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{vmatrix}$$

is composed of four complex elements. Equation (8) is applicable for natural fields at the

earth's surface over a broad frequency range, independent of wave polarization and point of measurement (Madden and Nelson, 1964; Word et al., 1970). The impedance tensor is rotated mathematically to find the principal resistivity directions, i.e., the directions for which the diagonal terms of Z are minimized ($Z_{xx}, Z_{yy} \approx 0$). If the earth were truly two-dimensional, a rotation angle would exist for which the diagonal elements go to zero, and a pair of frequency-dependent apparent-resistivity functions would be sufficient to describe the earth:

$$\rho'_{xy} = \frac{0.2T}{f} |Z'_{xy}|^2 \quad (9a)$$

and

$$\rho'_{yx} = \frac{0.2T}{f} |Z'_{yx}|^2, \quad (9b)$$

where the primes designate rotated parameters. The corresponding phase spectra are

$$\Phi'_{xy} = \tan^{-1} \left(\frac{\text{Im}(Z'_{xy})}{\text{Re}(Z'_{xy})} \right), \quad (10a)$$

$$\Phi'_{yx} = \tan^{-1} \left(\frac{\text{Im}(Z'_{yx})}{\text{Re}(Z'_{yx})} \right). \quad (10b)$$

The electrical strike found from the rotation process has a 90-degree ambiguity that can be removed on the basis of known geology or by adding a vertical-component magnetic measurement and performing a "tipper" analysis of the data (Word et al., 1970; Vozoff, 1972) to obtain a tipper strike. The apparent resistivity, ρ'_{ij} , which aligns with the tipper-strike direction, is called the E -parallel-to-strike (or TE mode) resistivity. The other apparent resistivity, ρ'_{ji} , is called the E -perpendicular-to-strike (or TM mode) resistivity. This terminology is commonly used even though we rarely encounter 2-D conditions. At most points of measurement Z_{xx} and Z_{yy} can be minimized by rotation, but no direction exists for which they are zero. A measure of the 3-D nature of the earth is expressed as

$$\frac{Z'_{xx} + Z'_{yy}}{Z'_{xy} - Z'_{yx}},$$

called the "skewness" (Swift, 1967). A common 3-D effect observed in the rotated MT resistivity curves is caused by local, near-surface conductivity inhomogeneities. These cause a parallel splitting of E -parallel and E -perpendicular resistivity curves, and bias both curves upward or downward, depending on whether the site is inside or outside the inhomogeneity and on the size of the inhomogeneity in relation to the skin depth of the wave in the inhomogeneity (Berdichevsky and Dimitriev, 1976; Hermance, 1982; Madden and Park, 1982; Park et al., 1983). Because small inhomogeneities do not affect the phase of the rotated resistivities, phase data can be used to obtain an initial 1-D model. This approach may seem simpler than correcting for the bias effects in the resistivity curves, but as the use of phase data requires an independent measure of surface resistivity (by means of dc resistivity, for example), phase data alone are not generally inverted.

The computational problems of interpreting MT data for fully realistic (i.e., 3-D) earth models is so immense (Mozley, 1982) that interpretations are usually based on 1-D approximations of the E -parallel resistivity or phase data followed sometimes by 2-D modeling. There exist a number of 1-D inversion routines (Oldenburg, 1979) that, despite their limitations, may yield useful insights when applied to the ρ_{xy} and ρ_{yx} separately. The 1-D models may then suggest 2-D starting models so long as the station separations are sufficiently small to avoid the undersampling problem. Even in situations where the sounding curves may appear amenable to 1-D inversions, the interpretation process is not without its pitfalls. In an example of 1-D MT interpretation, Rigby and McEuen (1982) showed that a free inversion (i.e., no imposed constraints on layer parameters) gave a resistivity-depth section that poorly matched the smoothed electric log from a nearby well at the East Mesa geothermal field. The problem was not with the inversion method, but with the nonuniqueness of the inversion of a finite data set containing errors. Using the smoothed well-log resistivities as constraints on inversion, they achieved a good match to the logged interval and extended their interpretation to depths below the bottom of the well more accurately.

Beyond 1-D inversions numerical analysis is limited to forward calculations using finite-element methods for the 2-D problem (Stodt, 1979) coupled to an automatic least-squares optimization-of-parameters program (Jupp and Vozoff, 1977). For 3-D forward modeling there are an integral-equation approach (Ting and Hohmann, 1981) and a hybrid of integral-equation and finite-element approaches (Lee et al., 1981). The 3-D numerical codes are limited to relatively simple models, e.g., a tabular inhomogeneity in a uniform or layered half-space (Wannamaker and Hohmann, 1982).

There are several well-documented examples of MT interpretations, based on 2-D and some 3-D modeling, that resolve subsurface resistivities well enough for drill-hole targeting. Seeking evidence for a deep, fracture-dominated reservoir and/or a deeper partial melt zone, Wannamaker et al. (1980, 1983) interpreted 93 MT soundings around Roosevelt Hot Springs, Utah. Figure 29 shows the apparent resistivity and phase pseudosections (ρ_{yx} and Φ_{yx}) for the E -perpendicular-to-strike (TM mode) components observed along an east-west line extending from Tertiary volcanics on the west, across the broad Milford graben and thence over the Mineral Mountains (a Tertiary granite pluton). For MT data that appear strongly 2-D, as these do, Wannamaker and Hohmann (1982) showed that only the TM mode results accurately reflect the true electrical structure. Trial-and-error 2-D fitting of the pseudosections was carried out using a finite-element code (Wannamaker and Hohmann, 1982) to produce the resistivity cross section, shown in Figure 30, which fits the data extremely well (Wannamaker et al., 1983). The asymmetric Milford graben appears as a thick sequence of relatively conductive elements. Dips of the graben-bounding faults are not accurately resolved, but the normal fault on the eastern margin of the graben is probably steeply dipping, judging from the high gradients in both the apparent resistivity pseudosection and the gravity profile. Although the resistivities for the granite pluton are generally high, the lower value of 100 ohm-m below the thermal anomaly correlates with the fractured reservoir rocks. There is no evidence for a deep conductor that one might associate with a partial melt.

Gamble et al. (1981) performed a detailed MT survey around the Cerro Prieto geothermal field. Among their findings was evidence for a narrow resistive zone at 500 m, plunging southeastward, which they associated with a region of hydrothermal metamorphism that had been outlined earlier by drilling, geophysical logging, and dc resistivity. The MT survey also located a deep conductor beneath the Mexicali Valley, at a depth of 2 to 3 km, that subsequent drilling and geological interpretation showed was associated with part of the deep brine circulation system that recharges the reservoir (Halfman et al., 1984).

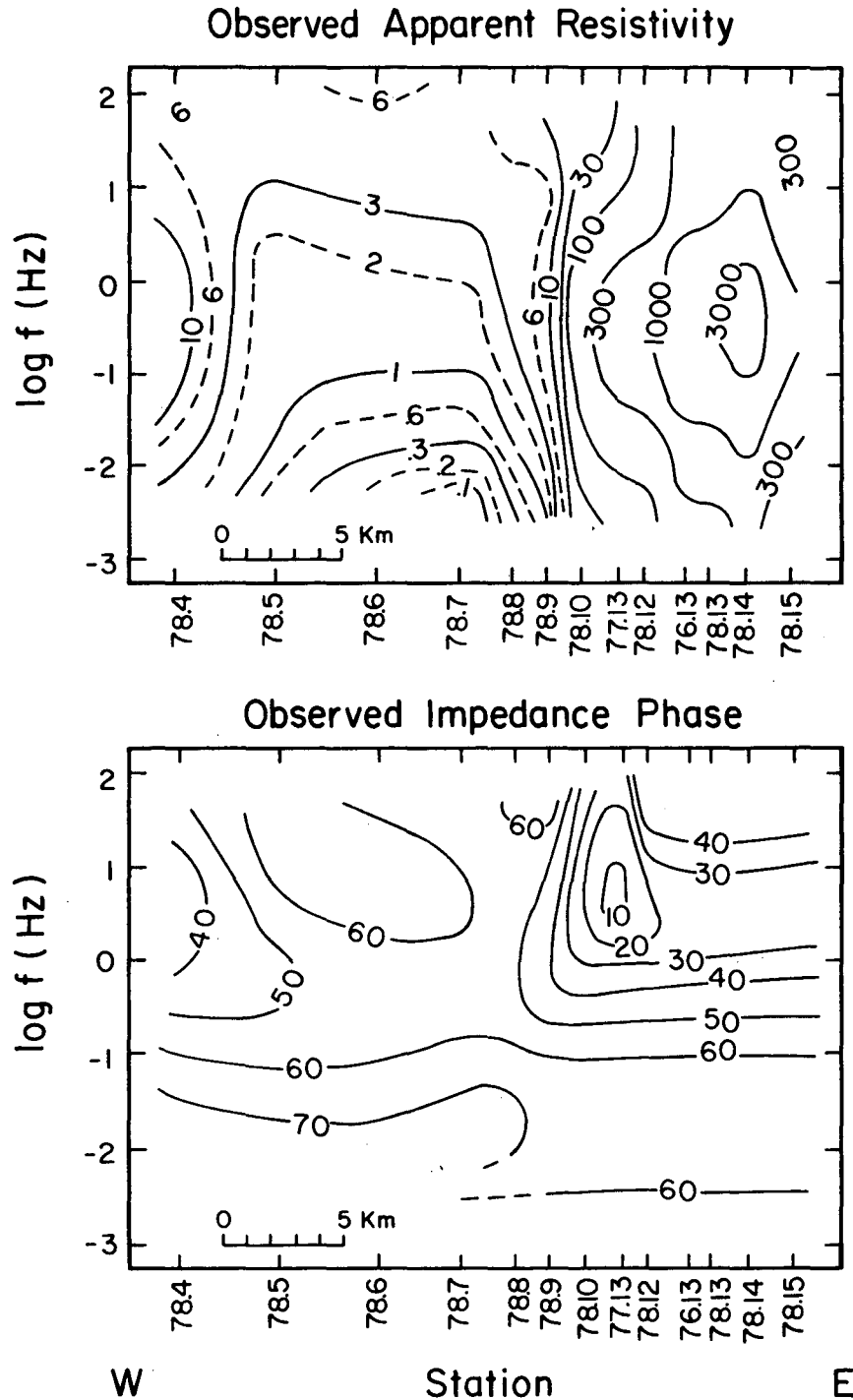


Figure 29.

Observed pseudosections of apparent resistivity (ρ_{yx}) and impedance phase (Φ_{yx}), identified as the *E*-perpendicular-to-strike mode, along a west to east profile crossing the Milford graben and the Mineral Mountains of southwestern Utah (after Wannamaker et al., 1983). Contours of ρ_{yx} are in ohm-m, and those of Φ_{yx} are in degrees. (XBL 856-10596)

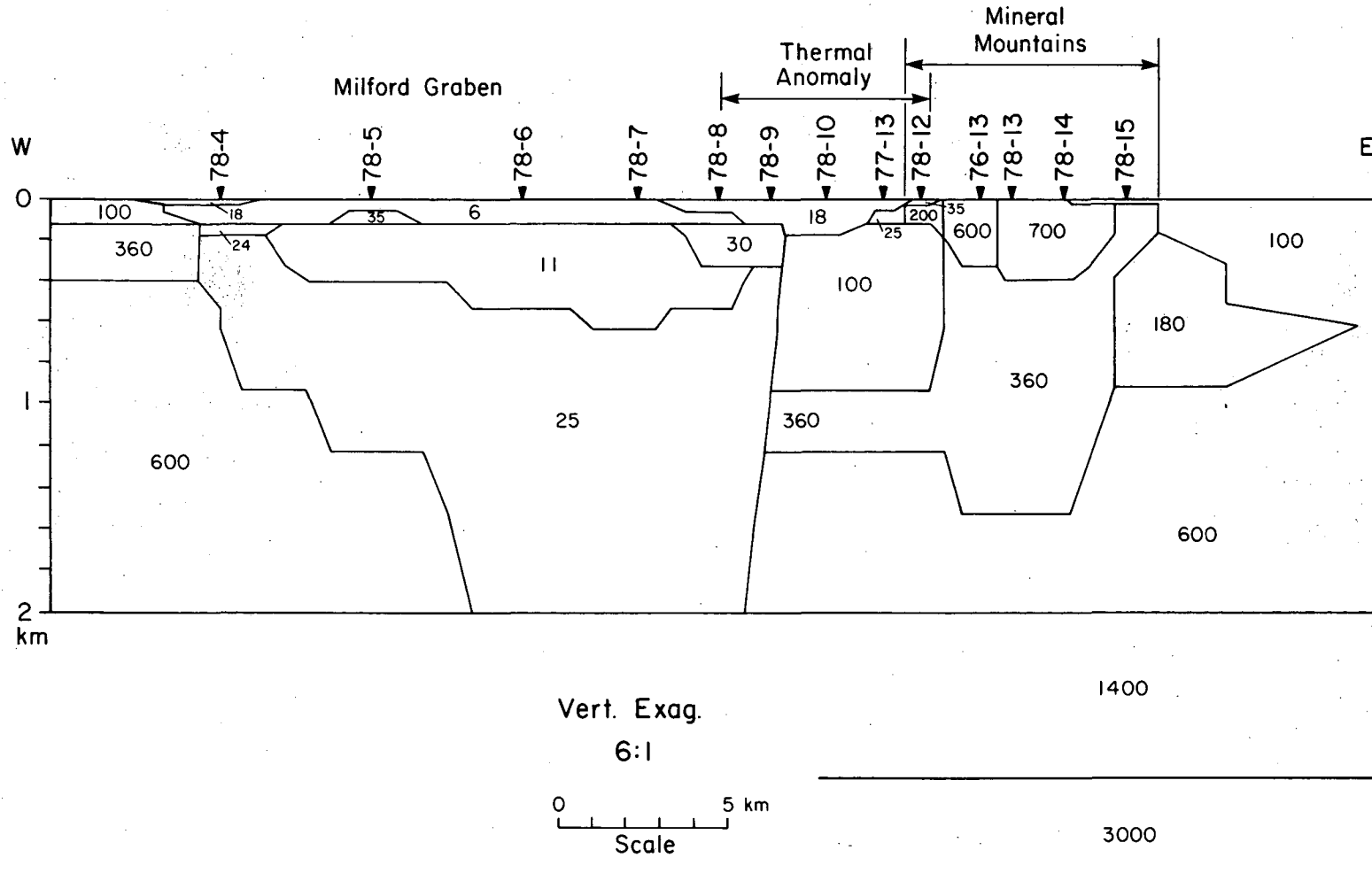


Figure 30. Resistivity cross section for the pseudosections in Figure 26, based on trial-and-error, two-dimensional, finite-element calculations (after Wannamaker et al., 1983). Resistivities are in ohm-meters, and the vertical exaggeration is 6:1. The location of the thermal anomaly associated with the Roosevelt Hot Springs geothermal system is indicated. (XBL 856-10597)

Mozley (1982) interpreted MT data collected around Mt. Hood, Oregon, correcting the data where necessary for local topographic effects. A shallow conductor on the flank of the volcano is the only feature thus far confirmed by drilling to be a bona fide geothermal aquifer (Goldstein et al., 1982); but, using parameters derived from the phase spectra of the impedance (Eqs. 10a and 10b), Mozley resolved a 2-D conductor at 12 km that may be a partial melt zone.

MT and AMT anomalies have been observed in two Pleistocene silicic collapse calderas, Valles Caldera (Wilt and Vonder Haar, 1986) and Long Valley (Hoover et al., 1976). In the Redondo Canyon area, the location of the medial graben and resurgent dome of the Valles Caldera, a 1-D interpretation of MT data indicated a shallow conductor (0.5 km deep) within the Bandelier welded tuff. Although a 1-D interpretation should be suspect in this environment, the conductor correlates reasonably well with known fracture zones that produce hot water in excess of 232 °C.

Hoover et al. (1976) found an AMT conductor near Whitmore Hot Springs in the center of the Long Valley caldera. Nine narrow-frequency bands spanning the 8 to 18,000 Hz range were measured at 25 stations. Subsequent deep drilling revealed that the AMT anomaly was caused by low-resistivity early post-caldera tuffs and lake sediments near the surface. As the 6000-ft-deep well intersected water of only 71 °C, the well was considered noncommercial.

Despite the complexity of MT data acquisition, processing, and interpretation, MT has gained wide acceptance and use in geothermal exploration because of general technological improvements, such as in-field processing and the remote-reference technique for noise cancellation (Gamble et al., 1979). Furthermore, the natural-field technique has certain inherent advantages over dc resistivity:

- (a) there is no need for heavy signal-generating equipment,
- (b) the depth of exploration is increased without the need to move wires and cables, and
- (c) inductive techniques are inherently better able to resolve conductive zones than are galvanic techniques (i.e., dc resistivity).

Controlled-Source Electromagnetics

Recently, controlled-source electromagnetic (CSEM) sounding systems have been developed, tested, and used effectively in several geothermal areas. In contrast to the widely used MT method, CSEM has had limited use, mainly in the U.S. (Goldstein et al., 1982; Keller et al., 1982; Anderson et al., 1983; Kaufman and Keller, 1983; Keller and Jacobson, 1983a, 1983b; Morrison et al., 1983; Wilt et al., 1983; Keller et al., 1984). In principle, the systems are similar to those used for many years in the search for massive sulfide orebodies, but some have been scaled up in power to provide depths of exploration to several km. The transmitters produce a large moment electromagnetic field by supplying 10 to 1000 A to either a long grounded wire (an electric dipole) or a large-area horizontal loop (a magnetic dipole). Signals are detected at the center of the loop or at varying distances from the source using a sensitive magnetometer. Modern signal receivers are capable of performing a wide range of signal-conditioning and signal-processing operations that greatly simplify operations.

Because a natural field method like MT and controlled-source methods should give comparable information under ideal conditions, one might question the use of CSEM over MT, since CSEM requires the additional cost and complexity of a transmitter. While it is not our purpose to argue for one method or another, there are practical reasons why CSEM

might be used with or in place of MT. First, there are several CSEM techniques that do not require E -field measurements, which can be difficult to make where extremely high contact resistances exist. Second, calculations show that the received signals are more specific to electrical conditions between transmitter and receiver, and thus the data are more suited to 1-D interpretation than MT, which usually requires 2-D and 3-D interpretation efforts. Finally, CSEM is inherently more sensitive to the total conductance (Eq. 5) of rock over a resistive basement than is the impedance function obtained in MT.

CSEM systems are categorized as frequency-domain (FDEM) or time-domain (TDEM), according to the shape of the current waveform and the detection intervals. In FDEM work a continuous EM wave is transmitted (e.g., a sine or square wave of variable frequency or a pseudorandom binary waveform), and the resulting primary and secondary fields are recorded continuously at discrete frequencies in the presence of the primary field. Except for the nature of the primary field, FDEM is similar to MT. In TDEM work a transient wave is transmitted (a dipolar square wave with an off period between positive and negative pulses), and readings are made only during portions of the current-off periods, when the induced current in the earth decays.

Wilt et al. (1983) describe a large-moment (10^6 to 10^8 A·m²) horizontal loop FDEM system developed at LBL and UC Berkeley (Figure 31). In-field data processing provides averaged spectra of the amplitude and phase of horizontal (radial and tangential) and vertical magnetic fields plus the complementary ellipticity and wavetilt spectra from 0.02 to 200 Hz. The low-frequency limit is controlled by natural geomagnetic noise whose amplitude at low frequencies increases roughly as $1/f^2$. The high-frequency limit is controlled by a combination of magnetometer response, radio-telemetry bandwidth, and primary signal strength. To extend the low-frequency end of the bandwidth, a distant-reference magnetometer monitors the horizontal components of the natural geomagnetic field (i.e., noise), assumed to be constant over the area. The telemetered and appropriately scaled reference signal is then electronically subtracted from the local signal prior to processing.

Automatic 1-D inversions of FDEM spectra are done by an iterative program that uses the Marquardt least-squares algorithm to fit amplitude-phase and/or ellipse-polarization parameters jointly or separately to layered models (Inman, 1975). Observed data are weighted by the calculated error of field measurements. Experience indicates that 1-D interpretations give results that compare well with 2-D interpretations of dc-resistivity data (Wilt et al., 1983). Because of the rapid fall-off in field strength with distance, dipole fields seem to be much less affected by nearby lateral discontinuities and current channeling, which, for example, impair 1-D MT interpretations. CSEM interpretations are currently limited to 1-D models, but 2-D forward modeling of dipole EM data may be done by a finite-element method (Lee, 1978) and a hybrid (finite-element/integral-equation) method (Lee et al., 1981). However, costs of 2-D modeling are high, and the model considered must be fairly simple to yield an accurate solution.

Rugged terrain presents a problem in interpretation because of topographic variations between transmitter and receiver and a transmitter that is usually not horizontal. To overcome a major part of this problem, one can treat the transmitted (primary) signal as arising from three mutually orthogonal dipoles located above or below the receiver. To interpret field data properly for this case, Haught et al. (1981) developed a computer program that combines layered-model solutions for vertical and horizontal dipoles.

Figures 32 and 33 show the results of a CSEM survey conducted along the western edge of Dixie Valley, Nevada, in a location where several geothermal wells intersect 260 °C

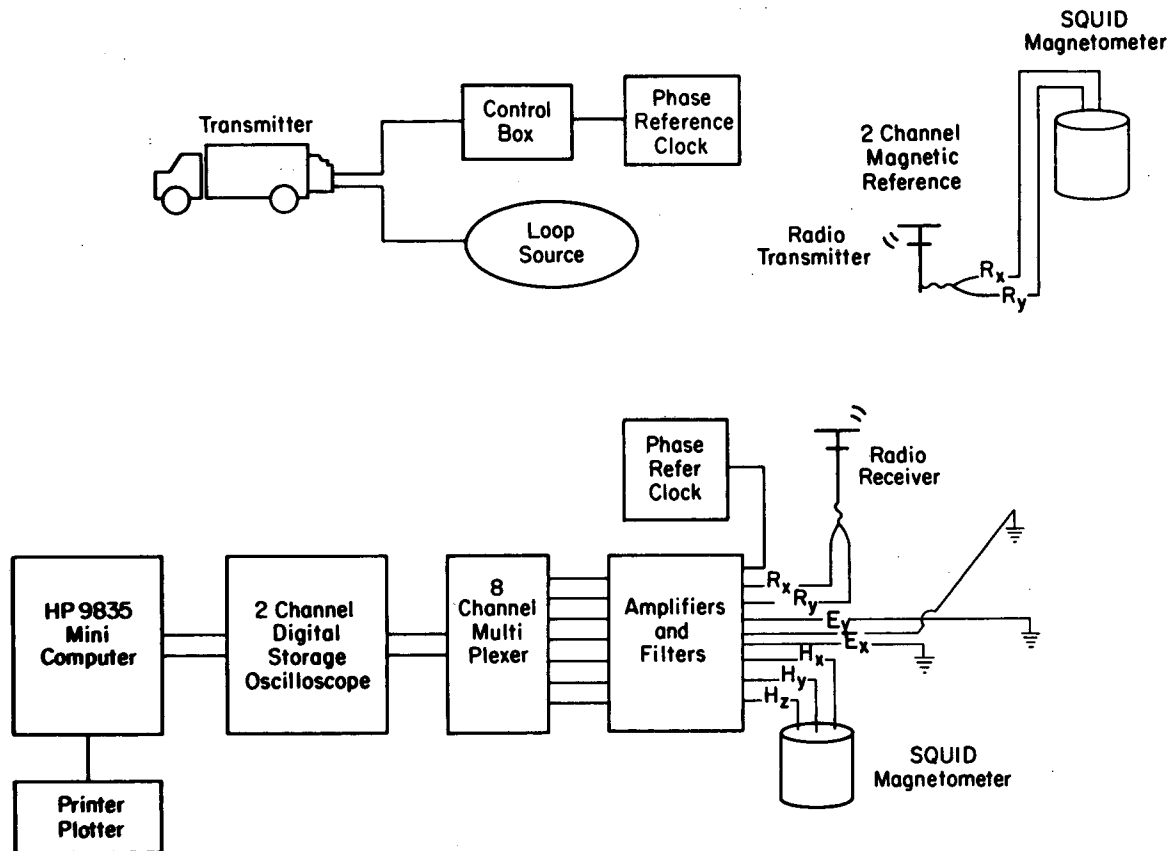


Figure 31. Block diagram of frequency-domain electromagnetic transmitter and in-field processing system developed at Lawrence Berkeley Laboratory and UC Berkeley (from Wilt et al., 1983). (XBL 818-3383A)

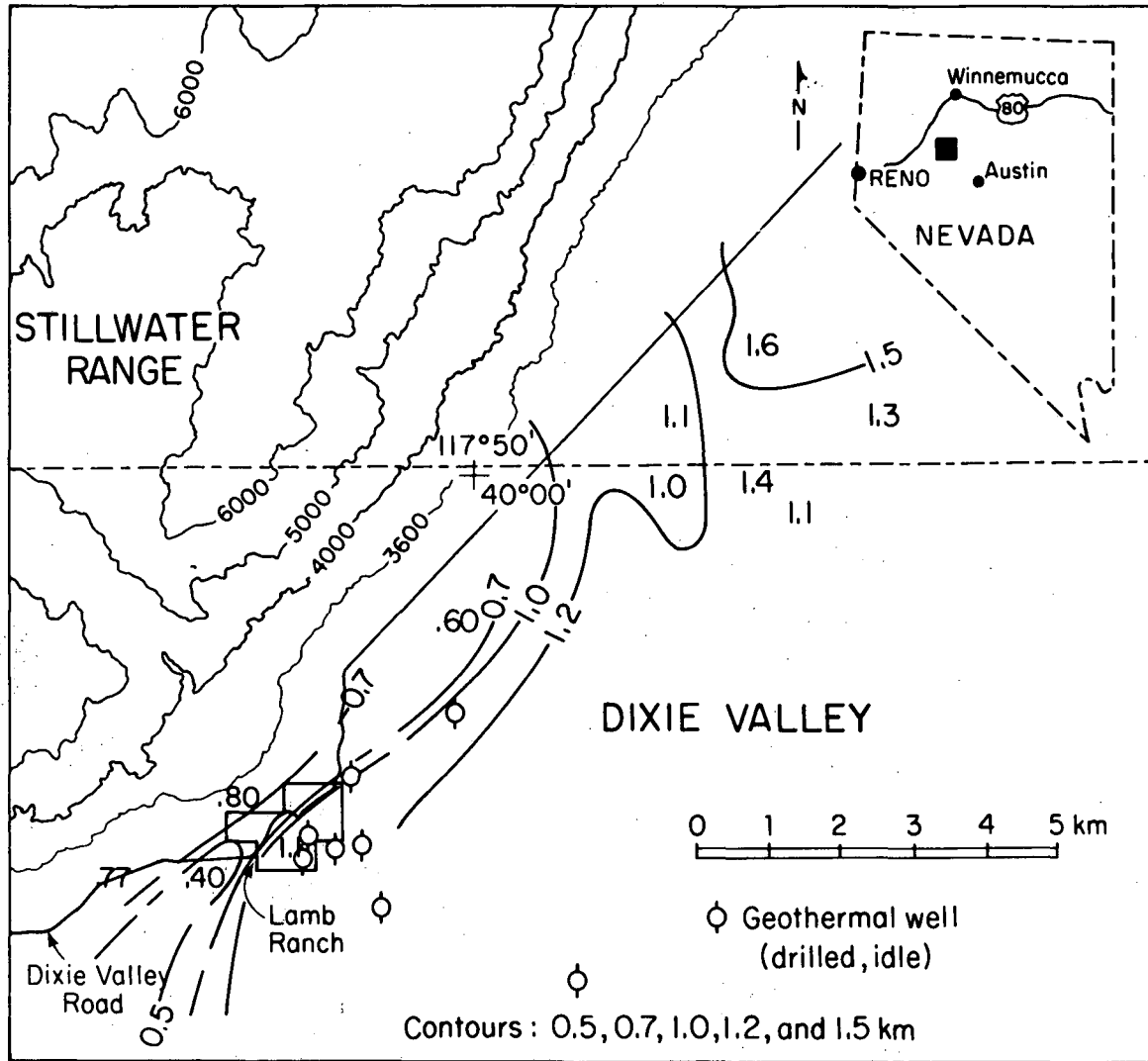


Figure 32. Depth (km) to resistive basement along the east flank of the Stillwater Range, Nevada, as discerned from EM soundings (from Wilt and Goldstein, 1983).
(XBL 8312-2450)

water at a depth of around 2 to 3 km (Wilt and Goldstein, 1983). The water is presumed to have ascended along a range-bounding fault and migrated laterally northeastward. The CSEM discerned a three-layer earth. Depth to the resistive basal layer is plotted in Figure 32, which shows that the basal layer drops off steeply from the Stillwater Range eastward into the valley. This depth agrees well in some places with basement depths from gravity and magnetic interpretation. A long northeastward-trending conductive zone was found within the basal layer (Figure 33). At the western end of the zone there is good agreement between the location of the conductor and producing wells. However, the conductor is shallower than the producing zones and may therefore represent a zone of hydrothermal alteration or warm-water leakage into a shallow aquifer.

Another FDEM technique used in geothermal exploration is the controlled-source audiomagnetotelluric (CSAMT) method. The name is derived from the fact that the controlled source operates in the AMT range of frequencies. Sandberg and Hohmann (1982) conducted a CSAMT survey over the Roosevelt Hot Springs area, Utah, using 600-m-long grounded electric dipoles driven by a commercial transmitter at frequencies from 32 to 9800 Hz. AMT scalar resistivities were obtained using a single component of the magnetic and perpendicular horizontal electric fields. The receiving electric dipole was oriented first parallel, then perpendicular, to the transmitter dipole. Stations were sufficiently far from the transmitting dipole that the EM waves could be treated like plane waves for interpretation. The CSAMT interpretations were consistent with the other geophysical and geological evidence, showing a low-resistivity zone corresponding to one found on the first separation ($n=1$) of a 300-m dipole-dipole resistivity survey. The CSAMT data were collected more rapidly than dc-resistivity data, because stations were not constrained to lines, long wires were not necessary, and only two transmitter sites were necessary. Because of the higher frequencies used the depth of exploration was found to be limited.

TDEM has been more widely used in geothermal exploration than FDEM, in part because of the availability of commercial equipment but also because of certain operational simplicities. Details on the techniques, instrumentation, and results are given by Kaufman and Keller (1983) and by Nabighian (1984). Typically, one detects and records only the vertical magnetic field H_z or its time derivative, which results from induced-current decay in the earth that occurs after the primary current into a loop or a grounded wire is switched off.

Surveys are conducted with the detector in the center of a horizontal loop (central induction sounding) or with detectors at various offsets from a grounded wire source. Configurations that seem to minimize the effects from lateral discontinuities are the central induction sounding and the receiver adjacent to the center of a grounded wire. For the same source moment, deep conductors are more easily detected using large offsets from the grounded wire. Unfortunately, with large offsets one invariably crosses lateral inhomogeneities, such as surface conductors that channel current and distort the decay curve.

The signals from current decay in the earth are weak, and various signal enhancement procedures are needed. Instrument noise and some low-frequency geomagnetic noise may have a long-term zero average and can be removed by signal averaging (stacking). High-frequency noise from power lines and distant lightning strokes are normally removed by filtering before stacking (Stoyer, 1982). It is customary to record many stacked waveforms at each sounding point and edit out the noisy points before taking a final average of the stacked decay curves (Anderson et al., 1983). The final stacked signal is deconvolved to eliminate filter-response effects. The processed decay curve is then numerically transformed into a curve of apparent resistivity versus time, in which form it takes on the appearance of a

very smoothed version of a borehole electric log. These curves can be displayed in cross sections, and the curves may also be fitted in a least-squares sense to layered-earth models, as is done for FDEM sounding curves.

Figure 34 shows a number of resistivity-depth profiles obtained via inversion of TDEM soundings taken 5 to 11 miles northeast of The Geysers steam field (Keller and Jacobson, 1983a; Keller et al., 1984). The grounded wire source, 1.1 km in length and carrying 2000 A, was located approximately 6 miles away, at the south end of Clear Lake. These profiles reveal a conductive zone (3–10 ohm·m) 1–3 km deep, which may be heated water in fractured rocks. A surprising feature of these results is that no conductive rocks were found at great depth (10–12 km) in the area south of Clear Lake, where the gravity interpretations (Chapman, 1975; Isherwood, 1975) indicated a large silicic melt.

SEISMOLOGICAL SURVEYS

Passive-Seismic Techniques

The possible relationship between seismic and geothermal activity was first suggested from work at the Taupo volcanic field, New Zealand (Clacy, 1968), The Geysers (Lange and Westphal, 1969), and at several locations of high-temperature gradient within the Imperial Valley (Douze and Sorrells, 1972). These early studies provided the impetus for passive-seismic methods in geothermal exploration. Because of the initial heavy emphasis on simple ground-noise surveys, seismic exploration became synonymous in the minds of many with that technique. There is far more to passive-seismic exploration than ground noise, and we present here a review and analysis of the various techniques.

All passive-seismic techniques rely on nature to provide a source of seismic energy (local or distant earthquakes or microseisms); the geophysicist provides a suitable array of geophones to detect the vertical (and sometimes horizontal) ground motion and a processing algorithm to convert ground motion into useful information on subsurface parameters and processes. The underlying basis for early passive-seismic surveys was the premise that geothermal mechanisms exist to cause the abnormally high seismic activity observed over the areas of high heat flow, e.g., random pressure variations in a convecting geothermal reservoir (Douze and Sorrells, 1972) and thermal stress cracking. The fact that high noise levels, in the 1 to 10 Hz range, were found at many known geothermal areas (Douze and Sorrells, 1972; Iyer and Hitchcock, 1976) prompted the use of seismic noise as an exploration method. Arrays of geophones were laid out over the area of interest, recording the signals for 24 to 48 hours, and after filtering and spectral analyses the relative signal powers in two or more low-frequency bands were displayed in contour form for the area. Early practitioners, such as Douze and Sorrells (1972), recognized that although some agreement could be found between ground-noise and heat-flow anomalies, one problem was to recognize and eliminate interfering noise sources (wind, traffic, pumps, regional earthquakes, etc). It was later noted that high noise levels were observed in valleys, the sedimentary basins themselves amplifying regional/local noise (Iyer and Hitchcock, 1976; Majer and McEvilly, 1979). The noise levels in some places were found to be proportional to the thickness of the basin fill material (Liaw and McEvilly, 1977), without adequate geological control it would be impossible to discriminate between body waves produced by a "geothermal" source and noise from resonance effects in the alluvium or basin-filling sediments.

To overcome this difficulty, Liaw and McEvilly (1977, 1979) tested and demonstrated the use of tight circular arrays of geophones (only 50 m in diameter to avoid spatial aliasing) with frequency-wavenumber ($F-K$) processing to determine the direction and phase velocity

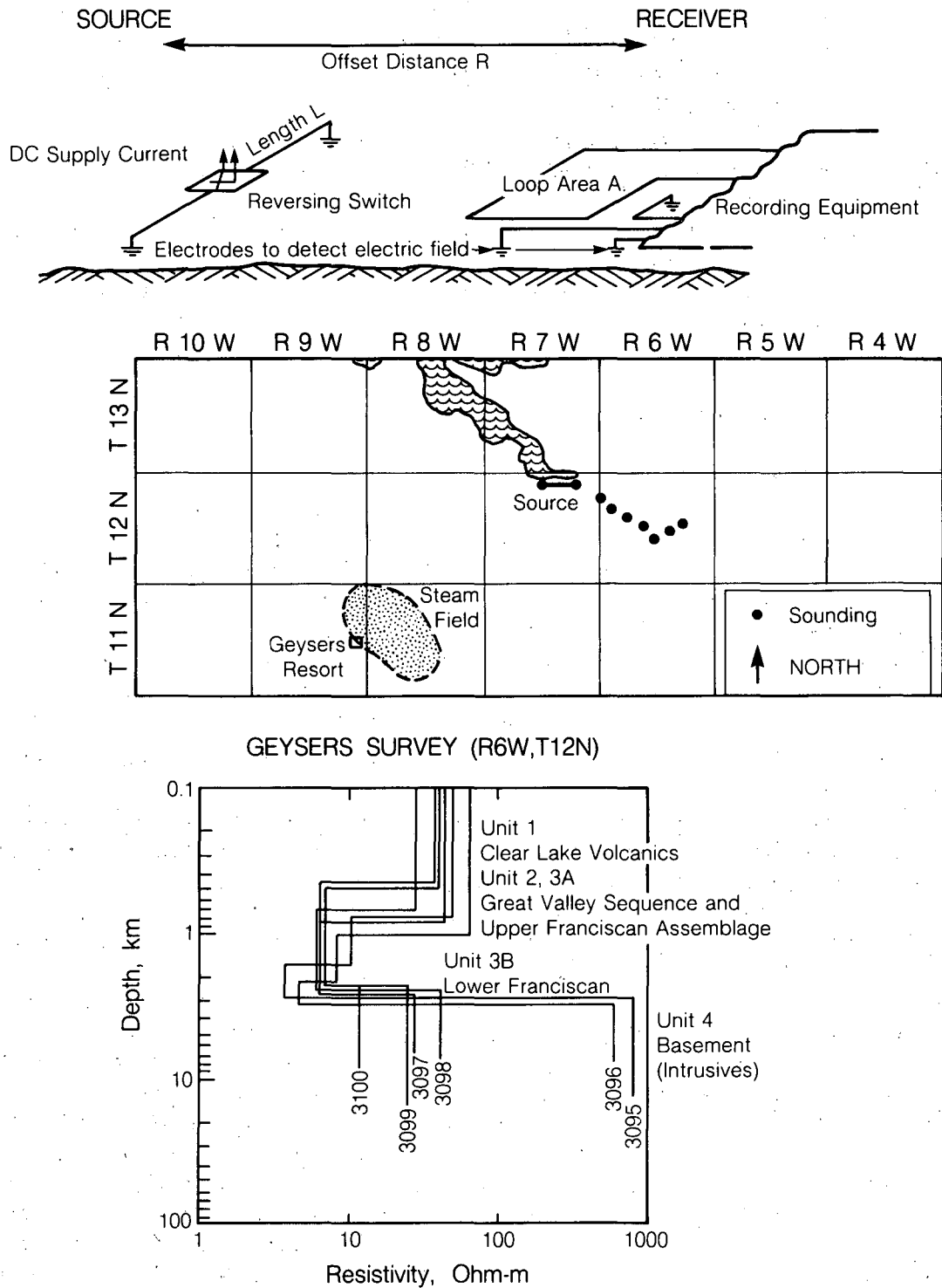


Figure 34. (a) Principal components of the Colorado School of Mines time-domain EM sounding system (from Keller et al., 1984). (b) Composite plot of the resistivity sections obtained by inversion of TDEM soundings in R6W T12N as part of The Geysers survey. The resistive units under the Franciscan assemblage may be serpentinite occurrences ($> 100 \text{ ohm}\cdot\text{m}$). (XBL 849-8844)

very smoothed version of a borehole electric log. These curves can be displayed in cross sections, and the curves may also be fitted in a least-squares sense to layered-earth models, as is done for FDEM sounding curves.

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seismic activity that has been correlated with geothermal fields. Most of the activity detected has been correlated either with tectonic activity (e.g., aftershock sequences) or with fluid reinjection or withdrawal (e.g., The Geysers). However, the data base may still be too small for any general conclusions. Seismic monitoring is usually done for only brief periods (four to six weeks), and there are usually no detailed records of seismic activity prior to field development.

One of the criticisms of the microearthquake technique is that, even with modern triggered digital event recording, data analysis tends to be labor-intensive and slow. Over a six-week observation period nearly a thousand events might be recorded, most of which require processing. To overcome this difficulty, an automated (in-field) seismic processing system has been developed (McEvilly and Majer, 1982), which is now being manufactured commercially. This device not only provides magnitude and hypocenter data, but also performs fault-plane solutions to show the type of strain release (focal mechanism) and performs a Wadati diagram analysis to obtain Poisson's ratio, a parameter of the rocks along the ray path. Because the Poisson's ratio is related to pore fluids and possible geothermal conditions, this parameter is often derived from microearthquake survey data. Poisson's ratio (σ) may be expressed in terms of elastic-wave velocities: $\sigma = (k^2 - 2)/2(k^2 - 1)$, where $k = V_P/V_S$ and V_P and V_S are the compressional shear-wave velocities, respectively. The ratio V_P/V_S may be estimated using a Wadati analysis, which involves plotting the time difference between S and P arrivals ($S-P$ time) against the P arrival time at many different stations for a single event, assuming V_P/V_S is constant along all ray paths to the seismometers. The slope of the line is $k-1$. Whereas σ is typically in the range of 0.25 to 0.30 for normal saturated rocks, anomalously low values of 0.15 were observed at the Coso geothermal area (Combs and Rotstein, 1976) and over the steam-production area at The Geysers (Majer and McEvilly, 1979). The low values of σ have been observed in the laboratory and explained by partial saturation (Nur and Simmons, 1969). However, it is not altogether clear that a low σ indicates steam-filled or two-phase voids. Gupta et al. (1982) noted that seismic stations with a low σ are associated, in general, with large teleseismic P -wave delays, while stations with normal σ do not show significant delays. This implies that a reduction in V_P with a lesser change in V_S could account for the low σ . Abnormally high values of σ , around 0.4, have been observed over two geothermal reservoirs in the Salton trough, the East Mesa (McEvilly and Schechter, 1978) and Cerro Prieto (Albores et al., 1980) reservoirs. Both these reservoirs are in a sandstone-siltstone-claystone sequence, but the σ observed is much higher than for ordinary porous sedimentary rocks, which tend to have the elastic behavior of a Poisson solid, $\sigma = 0.25$. Consequently, both these reservoirs must be exhibiting low rigidity (low shear modulus) due to a combination of high porosity and high temperature.

A third passive-seismic survey technique examines variations in the arrival times of P -waves (S -waves too, if horizontal geophones are also used) from regional or distant earthquakes (teleseisms). For detailed surveys, one or more linear geophone arrays are laid out at 500-m to 1-km geophone separations, and five to seven weeks of monitoring is normally required to detect teleseisms from different azimuths. Differences in arrival times (P -wave advances or delays relative to a reference station or to a crustal reference velocity) across the array are interpreted in terms of differences in V_P beneath the array, hence local differences in elastic parameters of the crustal and upper-mantle rocks. Because the first arriving waves are refracted waves traveling laterally along velocity discontinuities, some measure of depth discrimination can be achieved by using quarry, mine, or otherwise planned explosive blasts at varying distances from the array and with known origin times.

Teleseismic P -wave travel-time variations have been found at several geothermal areas where subregional-scale investigations have been made: Yellowstone National Park, Wyoming (Iyer, 1979); Long Valley, California (Steeple and Iyer, 1976); Coso, California (Reasenber et al., 1980); and Roosevelt Hot Springs-Mineral Mountains, Utah (Robinson and Iyer, 1981). In all these areas P -wave delays are observed, and the low-velocity region is interpreted as evidence for an abnormally high temperature and a small ($< 15\%$) fraction of partial melt. Beneath Roosevelt Hot Springs, for example, Robinson and Iyer (1981) postulate a pipe-like volume, 5 km in diameter, extending from a 5-km depth to the upper mantle, within which pipe there is a 5-7 percent decrease in velocity. If temperature alone were the cause of the velocity change, a 600 to 850°C temperature increase over the surrounding rock would be needed. This implies partial melting, especially in view of the high regional geothermal gradient. Unfortunately, the MT data do not give evidence for a conductivity anomaly that we might expect from a region of partial melt (Wannamaker et al., 1980). Either the melt is not observable by MT (e.g., the melt pockets are too deep and/or electrically discontinuous) or the P -wave anomaly is due partially to nonthermal effects. For example, on studies of the effects of fractures on V_P and V_S , Moos (1983) reports that V_P is decreased by both macro- and microcracks, and that chemical alteration of the rock adjacent to macroscopic fractures also appears to play an important role in reducing velocities.

As an adjunct to P -wave velocity studies, the amplitude and waveform of the P -wave are studied in terms of a variable attenuation along different ray paths. Majer and McEvilly (1979) examined P -wave attenuation at The Geysers by taking the ratio of the spectrum of the P -wave at each station to an arbitrary reference station to obtain a differential attenuation. Assuming that the attenuation factor Q is frequency independent over the bandwidth studied, an attenuation operator can be expressed as:

$$\exp \left[-\pi f \int_S \frac{ds}{Q V_P} \right],$$

where

- ds = incremental distance along ray path S ,
- Q^{-1} = intrinsic attenuation = $2\pi \Delta E / E$,
- $\Delta E / E$ = fraction of strain energy dissipated per cycle,
- V_P = P -wave velocity, and
- f = frequency.

For constant Q , the log of the ratio of P -wave spectra for two stations will be a linear function of frequency with slope $-\pi \delta t / Q$, where δt is the travel-time difference between the stations. The Q obtained in this fashion applies to the "differential ray path." An important assumption of this method is that the path to both stations is the same except for the last fraction, which passes through the zone of investigation. Majer and McEvilly (1979) found a shallow high- Q (low-attenuation) zone overlying a deeper low- Q (high-attenuation) zone at The Geysers. Although there are insufficient laboratory data on how Q behaves under the thermophysical-thermochemical conditions of a geothermal reservoir, the preliminary conclusion is that the high- Q zone is due to the vapor-dominated (i.e., undersaturated) state of the reservoir rocks. Majer et al. (1980) found anomalously high attenuations directly over the Cerro Prieto field, a liquid-dominated reservoir in relatively high-porosity sandstones. Recently, laboratory studies have been made to examine V_P / V_S and both P -wave and S -wave attenuations (e.g., Toksöz et al., 1979; Winkler and Nur, 1982) in dry and brine-saturated sandstones. Winkler and Nur (1982) found that pore fluids

dominate attenuation in the upper parts of the crust and that attenuation is a more sensitive indicator of the degree of saturation than velocity ratios. They found that the attenuation ratio Q_S/Q_P is a sensitive indicator of partial saturation and therefore might find practical applications in the exploration for vapor-dominated geothermal areas.

Active-Seismic Techniques

Modern reflection seismology has been used sparingly and mainly as a research technique at geothermal areas. The high cost of data acquisition and processing plus the complex geologic structures normally found in geothermal areas have limited the extent to which reflection seismology has been applied. Hayakawa (1970) reported on some limited seismic-reflection work at the Matsukawa Field, Japan, which led him to conclude that fissure zones could be located by a rapid decrease of seismic-wave energy in a narrow band and also by a change in wave phase. This work encouraged Denlinger and Kovach (1977, 1981) to investigate whether seismic profiling could be used to map fractures at The Geysers. They reasoned that steam-filled fractures would lower V_P of the medium, the decrease proportional to fracture size, fracture density, and the total thickness of the fracture zone. Theoretically, a low-velocity fracture zone thicker than one-eighth of the wavelength will act seismically like a thin, highly reflecting zone. Using four Vibroseis[®] sources producing a 58- to 12-Hz downsweep, and lines laid out split spread, 12-fold, with a 33-m group interval and 880-m cable length, Denlinger and Kovach (1981) found a good reflector at between 3 and 4 km that may indicate a major tectonic boundary in the Franciscan assemblage; the reflected seismic energy was enhanced because of increased fracturing or the presence of imbricate layers. However, depth resolution of the reflecting horizon was poor because short lines (1.6 km) were surveyed to maximize signal to noise at depths of 1 to 3 km. A better, but more expensive, approach would have been to run the survey 48-fold on lines 3 to 5 km in length.

As seismic-profiling methods are best suited for detecting near-horizontal discontinuities, steeply dipping fracture zones could escape direct detection. This is particularly significant in parts of The Geysers area, where exploratory drilling has encountered steeply dipping fractures.

Working in a slightly less complicated geologic environment, Goupillaud and Cherry (1977) used both conventional Vibroseis[®], which produce compressional waves, and experimental horizontal vibrators, which produce shear waves, to determine whether reflection seismology could locate the fault/fracture zones believed to control fluid circulation into the East Mesa reservoir. Although the amount of data was limited and much of the data wasn't of the best quality, they found evidence for a loss of reflections that correlated with the reservoir and reached the preliminary conclusion that the scattering effect was a direct result of fracturing.

For another Salton trough geothermal field, Blakeslee (1984) examined several lines of Vibroseis[®] data acquired by the Comisión Federal de Electricidad at their Cerro Prieto field. Figure 35 shows a processed seismic section for a line passing over the reservoir in a northwest-southeast direction. The data were migrated using a finite-difference algorithm that allowed the interpreter to account for vertical and horizontal velocity gradients determined from a comprehensive velocity analysis. Abrupt velocity changes closely follow a seismic reflection attenuation zone (RAZ) that is associated with the reservoir. Wells M-130 and M-10A indicate the main production area; the reservoir is at a depth of 1.1 to 1.4 km. The cause of the RAZ is not understood, but it appears on all lines over the reservoir. The reflections may be washed out due to hydrothermal alteration, but the interface region between the unaltered, unconsolidated sediments and the underlying altered sediments is too

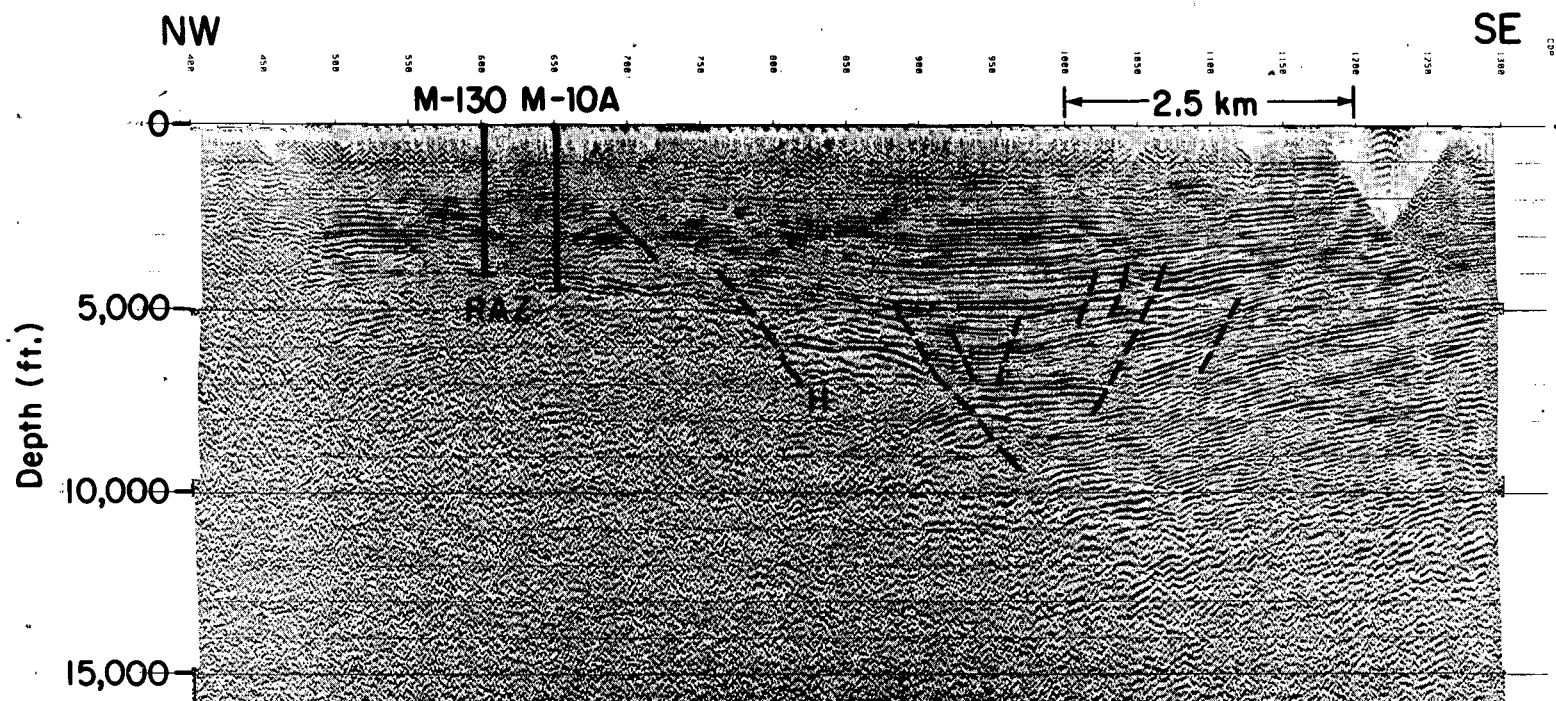


Figure 35. Processed seismic section over the Cerro Prieto field. Over 60,000 traces comprise the data set that was obtained with a Vibroseis[®] source and linear 14-48 Hz, 16-56 Hz, and 18-60 Hz sweeps. There were 256 channels per shot, an 80-ft receiver spacing, and a 4-ms sampling rate. A number of faults run through the section. A seismic-reflection attenuation zone coincides with the reservoir. The lack of reflections may be due to extreme seismic absorption due to thermal causes, fracturing, and/or hydrothermal alteration (from Blakeslee, 1984). (XBB 845-3555A)

gradual to explain the sharp velocity changes. Reasons put forth to explain the absence of reflections are intense fracturing or extensive seismic absorption due to thermal effects. However, neither of these causes has been verified. It is suggested, however, that the RAZ in conjunction with its high-velocity lid can be used as a discriminant of the geothermal system.

Seismic observation in wells have been used for many years as a way of obtaining a sonic velocity log and for improving data quality (Gal'perin, 1974). The approach has evolved into what is now referred to as Vertical Seismic Profiling (VSP) which has proven to be helpful in resolving geologic features not possible by using surface seismic data (Balch and Lee, 1984). VSP surveys have been useful for obtaining the acoustic properties of individual lithologic units, for detecting reflections from near-vertical discontinuities such as faults, and for characterizing a fractured rock mass. All of this information is extremely important in geothermal exploration and geothermal reservoir studies because the fluid circulation system in so many geothermal reservoirs is controlled by either major faults or by dominant sets of fractures.

Conventional VSP uses a seismic detector clamped against the wellbore wall at intervals ranging from 10 to 100 feet. A surface seismic source, located near the hole and then moved to varying offsets, provides the seismic energy. To realize the full potential of VSP, particularly for fracture detection, Crampin has argued in a series of papers (Crampin, 1978, 1984, 1985, among others) that one must use a 3-component detector. Coupled with the use of *P*- and *S*-wave sources at the surface, the 3-component data allows one to examine the phenomenon of shear wave splitting due to anisotropy effects (Leary and Henyey, 1985). There is laboratory (Myer et al., 1985) and theoretical work (Schoenberg, 1980, 1983) that relates the shear wave anisotropy to fracture parameters such as fracture direction, spacing, and "stiffness," a measure of the discontinuity in displacement due to a seismic wave crossing a fracture.

Using the Seismographic Services Corporation (SSC) high-temperature, high-pressure, 3-component, hydraulic wall-lock geophone (K-tool), Majer et al. (1986) conducted a fracture detection experiment using *P*- and *S*-wave sources at The Geysers geothermal field. The tool designed to operate at temperatures up to 225 °C in a water-filled borehole experienced seal failures because the O-rings were not designed for steam immersion. Nevertheless, sufficient data were collected and analyzed to show that shear wave anisotropy was evident in an 11% velocity difference and in apparent splitting of SH- and SV-polarized waves. Moreover, the direction of anisotropy is consistent, to a first order, with the direction of the dominant fracture set in the greenstone caprock.

DISCUSSION

Geothermal exploration is a relatively new area of applied science that has evolved rapidly over the last 15 years or so in response to the challenge of developing alternative energy sources. Prior to 1965 most exploration was done by the pioneering countries in the field of geothermal development: New Zealand, Iceland, France, and Italy. The techniques applied were derived mainly from those used for studying active volcanoes and for mineral exploration. The need for improved techniques gained impetus as the more obvious and accessible areas were drilled and developed, and attention focused on exploring areas where greater exploration risks and uncertainties existed. Aided by the Geothermal Energy Act of 1974, the U.S. Department of Energy has been able to foster much needed research and development that, coupled with case studies, has provided improved tools and techniques to the private sector for data acquisition and interpretation. The case-study investigations supported by DOE in Nevada, the Cascade Range, Cerro Prieto, Roosevelt Hot Springs, and the Valles Caldera, among other places, and the detailed scientific investigations by the U.S. Geological Survey at The Geysers, Long Valley caldera, and Yellowstone National Park, to name but a few systems studied, have provided a greater depth of understanding of the thermophysical, thermochemical, and hydrogeologic processes and parameters associated with a variety of hydrothermal-geothermal systems. Together with corroborating subsurface information gained from many wells, the exploration-technology research has provided substantial benefits to the geothermal industry mainly by putting existing tools and techniques on a sounder scientific basis, by providing better interpretative methods, by assembling a vast amount of basic data, and by improving cost-effectiveness of certain methods.

Geothermal-exploration technology may not change or improve dramatically in the near term, but any change will be largely a function of energy-market forces. We can expect improvements in geochemical geothermometry and geochemical field techniques and in sampling, remote-sensing applications, and interpretation of seismic and electromagnetic data, for example. But the greatest improvements may arise indirectly as a result of careful geochemical and geophysical observations (i.e., monitoring) of detailed geohydrological and reservoir-simulation studies at producing geothermal fields and from fundamental studies related to the evolution of magmatic-hydrothermal systems.

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