GL01275

LBL-18224
Preprint

Lawrence Berkeley Laboratory
UNIVERSITY OF CALIFORNIA

EARTH SCIENCES DIVISION

To be published in Geothermal Science and Technology

SUBREGIONAL AND DETAILED EXPLORATION FOR GEOTHERMAL-HYDROTHERMAL RESOURCES

N.E. Goldstein

May 1986

Prepared for the U.S. Department of Energy under Contract DE-AC03-76SF00098

LEGAL NOTICE

LBL-18224

SUBREGIONAL AND DETAILED EXPLORATION FOR GEOTHERMAL-HYDROTHERMAL RESOURCES

Norman E. Goldstein

Earth Sciences Division
Lawrence Berkeley Laboratory
University of California
Berkeley, California 94720

May 1986

This work was supported by the Assistant Secretary for Conservation and Renewable Energy, Office of Renewable Energy Technologies, Division of Geothermal Technology of the U.S. Department of Energy under Contract No. DE-AC03-76SF00098.

TABLE OF CONTENTS

INTRODUCTION 1
- Exploration Technique Overview 1

GEOLOGICAL TECHNIQUES 5
- Igneous-Volcanic Rock Associations 5
- Stratigraphic and Structural Interpretation 10
- Airborne Remote Sensing 11
  - Photographic Techniques 11
  - Thermal Infrared Imagery 13
  - Landsat and Airborne Multispectral Scanning 13
- Field Investigations 14
  - Hydrothermal Alteration as a Guide to Subsurface Conditions 15
    - A Sandstone Reservoir: Cerro Prieto, Baja California 16
    - An Andesite-Rhyolite Reservoir: Los Azufres, Mexico 16

GEOCHEMICAL TECHNIQUES 21
- Classification of Water Discharges 22
  - Sodium Chloride Water 22
  - Acid Sulfate-Chloride Water 22
  - Acid Sulfate Water 22
  - Calcium Bicarbonate Water 22
- Classification of Gas Discharges 23
- Sampling and Analysis of Liquid and Gaseous Discharges 25
  - Helium-Isotope Ratios 25
  - Noble Gases 26
  - Oxygen-Hydrogen Isotopes 26
  - Tritium 29
  - Geochemical Geothermometry 29
- Sampling and Analysis of Surface Soils and Rocks 32
- Sampling and Analysis of Surface Volatile Trace Elements 32
  - Radon 32
  - Mercury 34
  - ^{4}Helium 36
- Sampling and Analysis of Subsurface Rocks 38
  - Present Subsurface Temperatures and Field Boundaries 38

Dating Thermal Events 38

GEOPHYSICAL TECHNIQUES 41
Thermal and Temperature Surveys 41
Basic Heat-Flow Equations 41
Shallow Temperature Surveys 43
Temperature-Gradient Measurements 44
Thermal-Conductivity and Heat-Flow Measurements 47
Magnetics 48
Gravity 51
Electrical and Electromagnetic 55
Factors That Influence Formation Resistivity 56
Electrical Resistivity in Geothermal Areas 60
Self-Potential 61
The Telluric Method 65
The In-Line Telluric Method 70
DC Resistivity and Induced Polarization 70
Electromagnetic (EM) Techniques 80
Controlled-Source Electromagnetics 86
Seismological Surveys 92
Passive-Seismic Techniques 92
Active-Seismic Techniques 97

DISCUSSION 101

ACKNOWLEDGEMENTS 101

REFERENCES 103

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 ($SiO_2 < 55\%$) and rhyolitic ($SiO_2 > 70\%$) 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

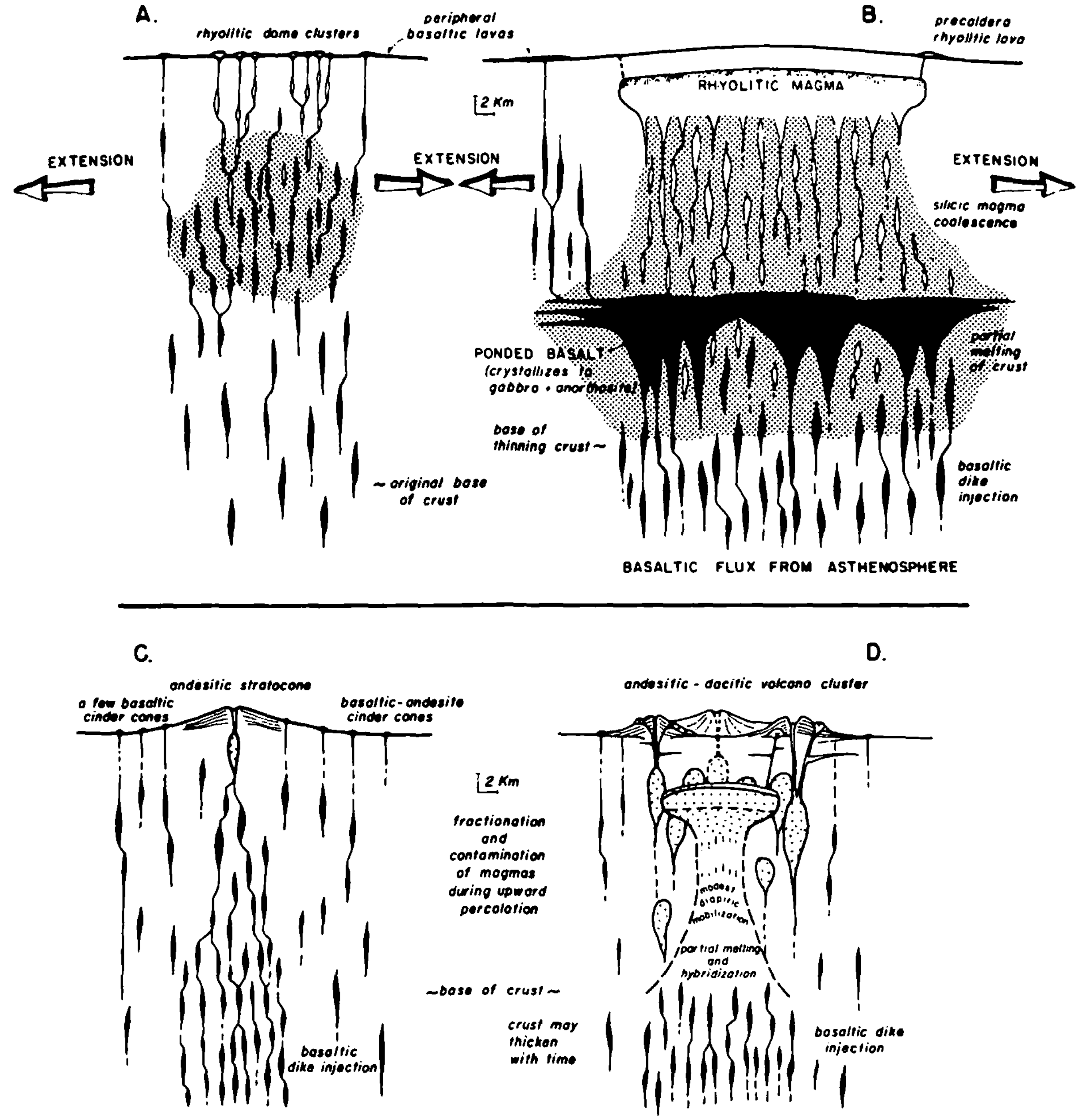


Figure 1. Some contrasting styles of lithospheric magmatism. The two upper panels depict rhyolitic-basaltic magmatism under stress conditions favoring marked crustal extension: (A) modest power input; (B) large power input, advanced stage. Shaded regions indicate partial melting of crustal rocks sufficient 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 km^3) 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.

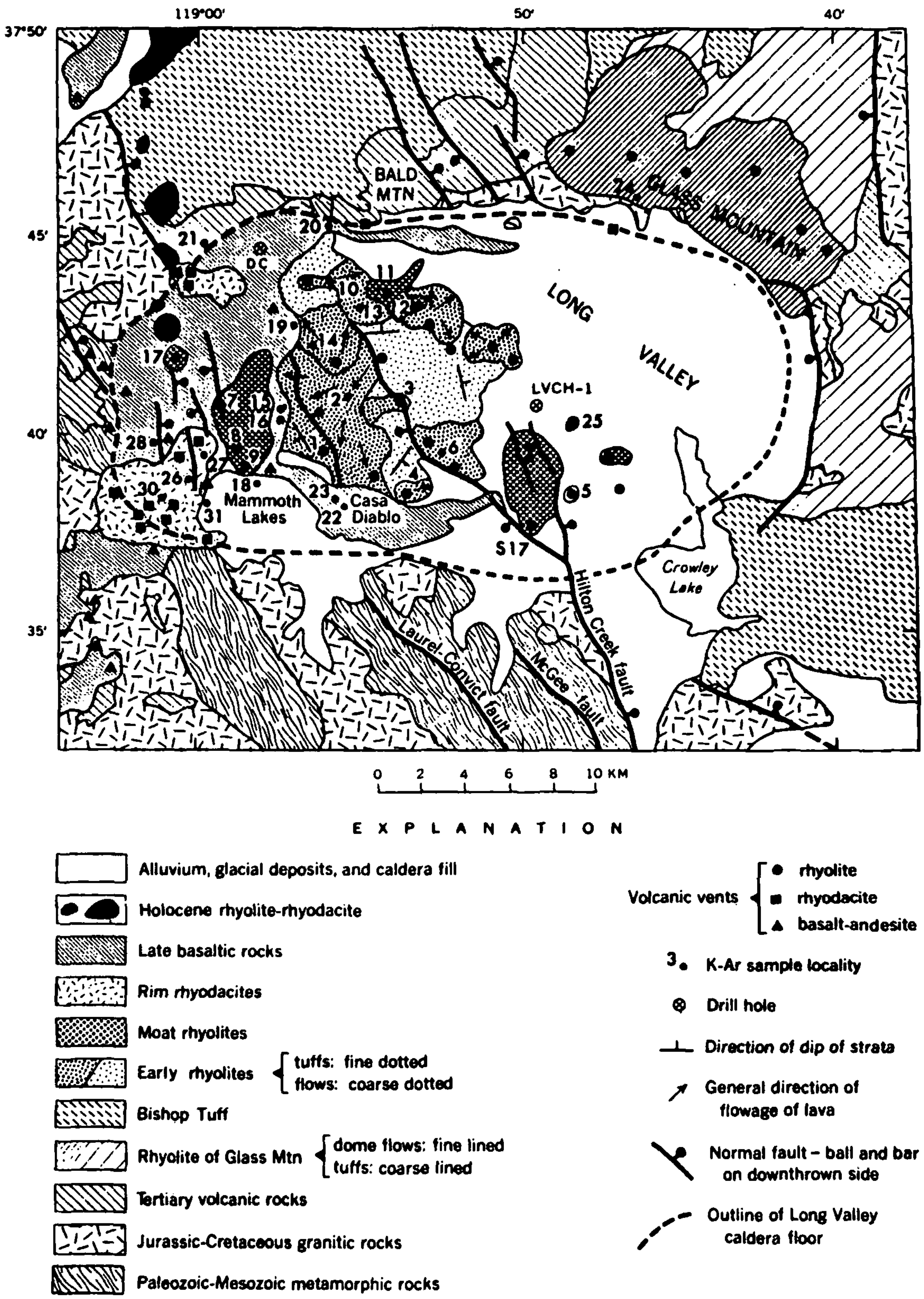


Figure 2. Generalized geologic map of Long Valley caldera (from Bailey et al., 1976). (XBL 849-3862)

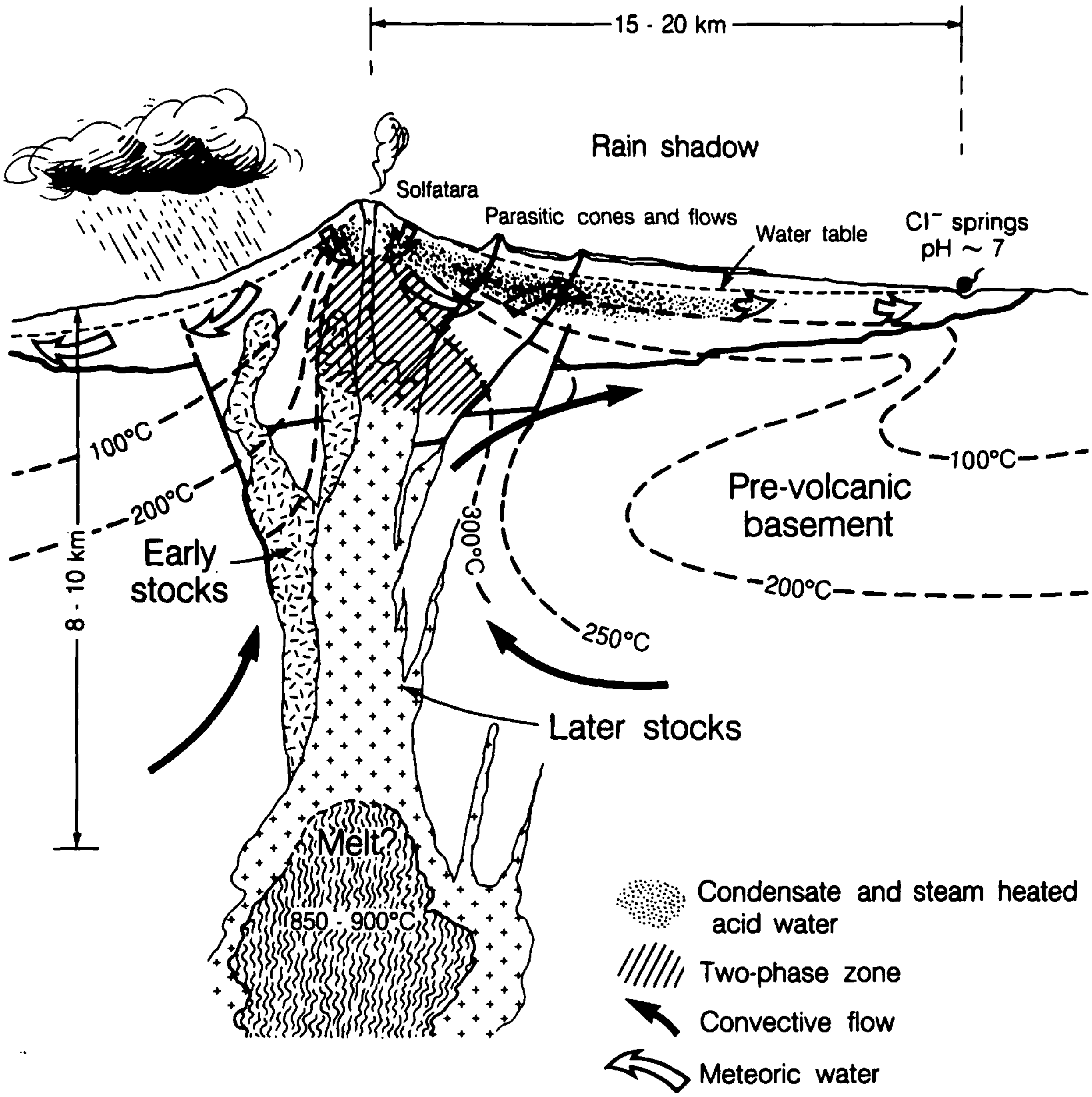


Figure 3. General schema of a possible hydrothermal-geothermal system associated with an andesite-dacite composite volcano typical of active island arcs (after Henley and Ellis, 1983).
(XBL 8411-6164)

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 subvertical, 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 subsurface 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

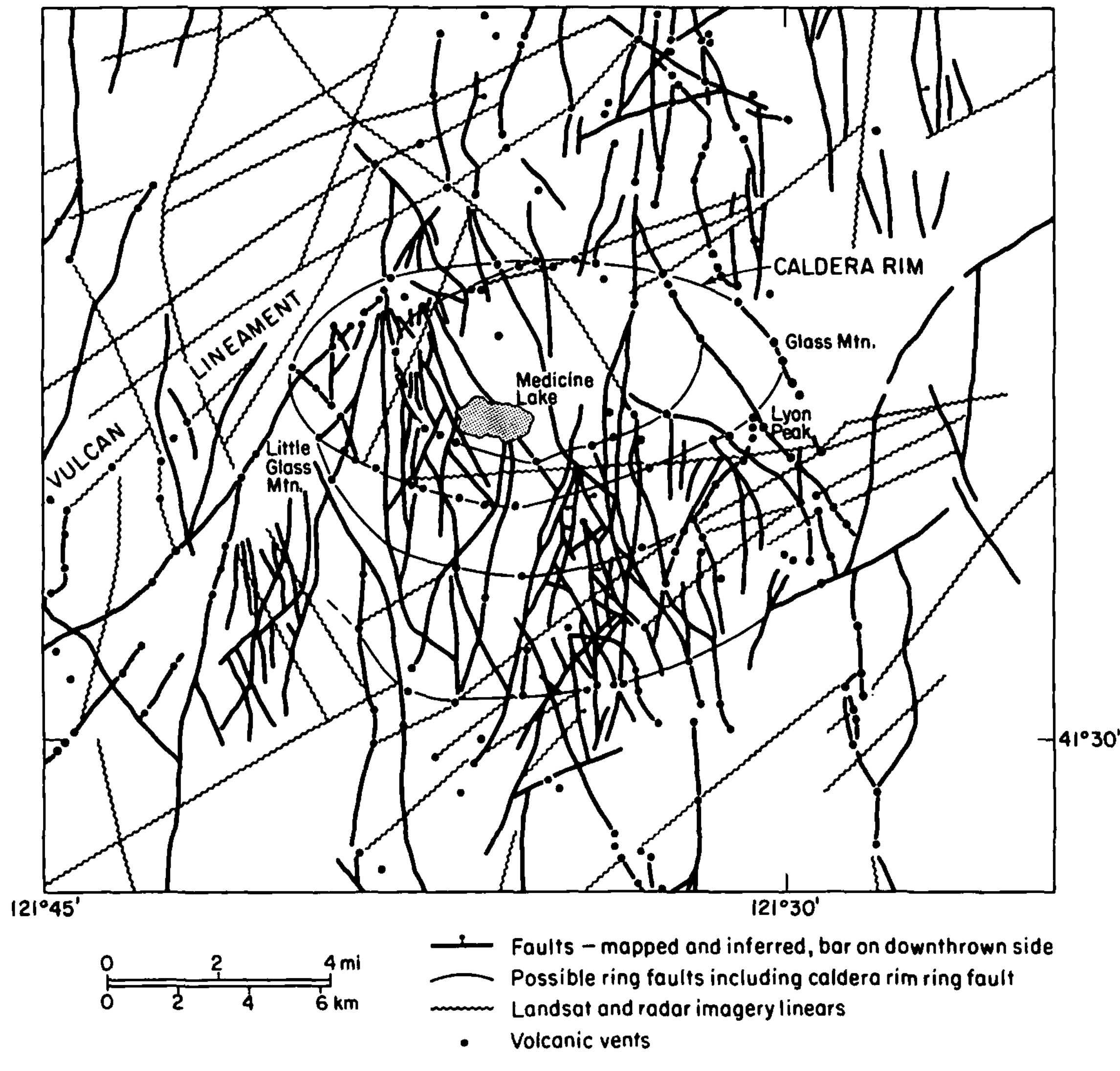


Figure 4. Faults and lineaments at the Medicine Lake volcano, in the Cascade Range, northern California (from Ciancanelli, 1983).
(XBL 841-9516)

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 $\times$ 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.

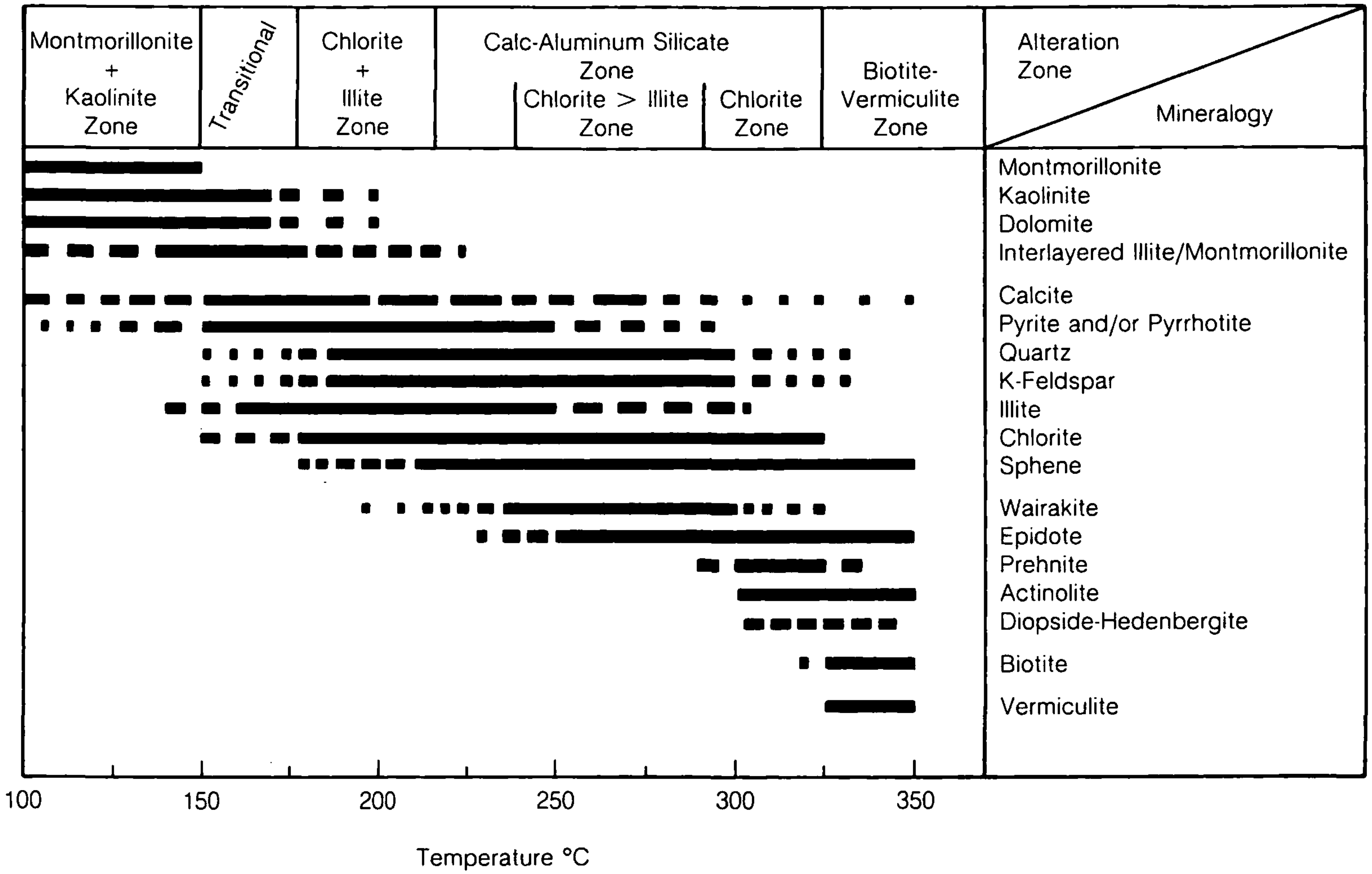


Figure 5. Temperature ranges for the occurrence of hydrothermal minerals in the sandstones of the Cerro Prieto geothermal field, Baja California (after Elders et al., 1979).
(XBL 863-10707)

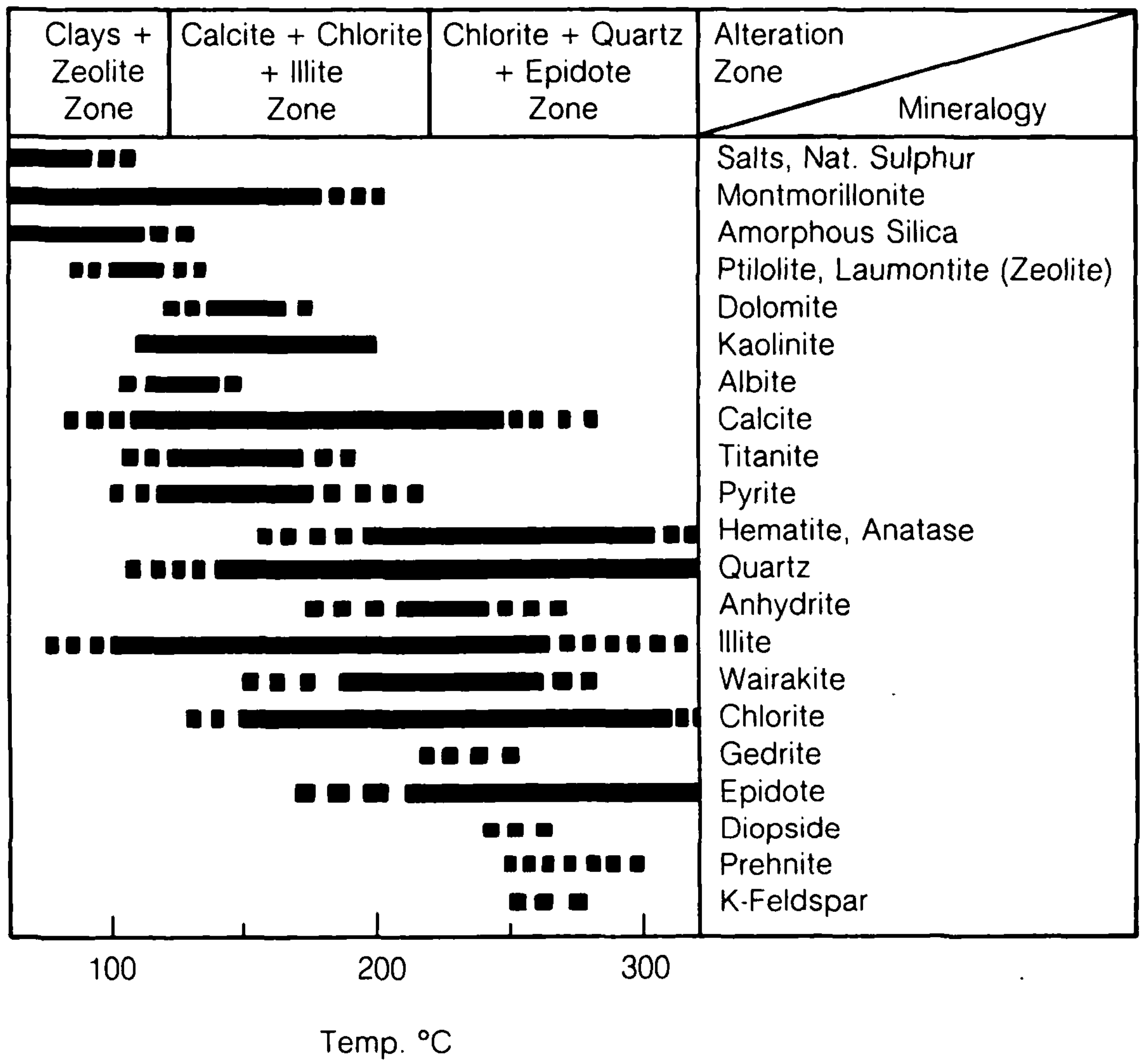


Figure 6. Temperature ranges for the occurrence of hydrothermal minerals in the andesites of the Los Azufres geothermal field, Michoacan (after Cathlineau et al., 1985).
(XBL 863-10706)

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

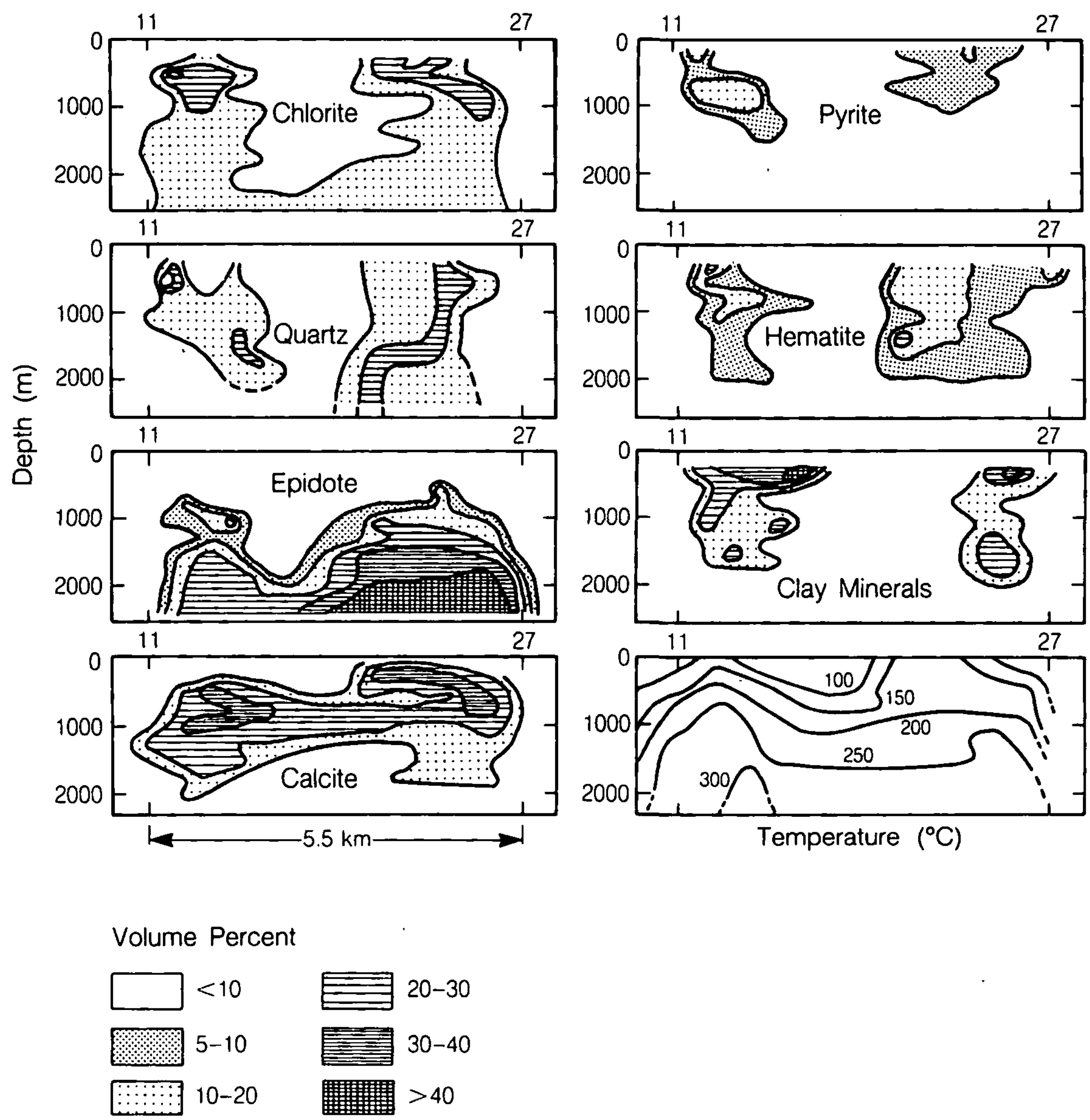


Figure 7. Distribution and concentration of hydrothermal minerals in the Los Azufres geothermal field (after Cathlineau et al., 1985). The cross-section extends from the Tejamaniles zone (well 11) on the south through the Maritaro zone and well 27 on the north. Horizontal and vertical scales are equal. (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 noncondensible 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 (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 Cl^-/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 (H_2O), various noncondensible gases are present (e.g., CO_2, H_2S, CH_4, H_2, N_2) 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 (CO_2, SO_2, and H_2S) from thermal metamorphism of rock containing calcite and sulfides, (d) gas (mainly CH_4) 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 CO_2 and N_2) 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 noncondensible 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 CO_2, H_2S, He, and H_2 (Arnórsson, 1974). Steam arising from active volcanism is distinguishable from steam related to a circulating hot-water system in that the former contains SO_2 and gaseous HCl and HF. SO_2 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 H_2, H_2S, and SO_2, are controlled by mineral buffers. For example, H_2S, SO_2 may be in chemical equilibrium with iron sulfide mineral phases and SO_4^{2-} ions. Armannsson et al. (1982) argued that in Iceland the absence of SO_2 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 noncondensible gases present. N_2-CO_2 waters often represent low-temperature volcanic systems. N_2 is derived from air dissolved in the circulating waters, and CO_2 is derived from a variety of sources, including magmatic and thermal metamorphism of limestone. H_2S-CO_2 waters, on the other hand, are more representative of hydrothermal systems with abyssal heat recharge and are typical of most high-temperature geothermal fields. CO_2 is the principal noncondensible gas present in most thermal areas, and may occur as a result of various reactions. It is generally believed that the major source of CO_2 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

$$CaCO_3 + SiO_2 = CaSiO_3 + CO_2,$$

but there are many others that produce CO_2 plus various metamorphic minerals such as epidote, diopside, and prehnite. Cavaretta et al. (1982) showed that the CO_2 partial pressures at the Larderello and Serrazzano geothermal fields are close to equilibrum for the reaction

$$2 \text{ clinozoisite} + 3 \text{ quartz} + 2 \text{ calcite} = 3 \text{ prehnite} + 2\, CO_2.$$

In such a case it is possible to fit CO_2 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 CO_2 component has been sought because CO_2 is much less soluble in magma than water (steam) and other noncondensible gases. Isotopic analyses of C in CO_2 collected at Casa Diablo Hot Springs, Long Valley caldera, California, show that the $^{13}C/^{12}C$ ratio is consistent with values found in fluid inclusions in igneous rocks. The apparent lack of sedimentary sources of CO_2 beneath the caldera suggests that the CO_2 is exsolving from a contemporary magma (Taylor and Gerlach, 1983).

In addition to CO_2, many geothermal systems contain CH_4 and NH_3. If there is excess CO_2 present, methane may be generated through the Fischer-Tropsch reaction:

$$CO_2 + 4H_2 = CH_4 + 2H_2O,$$

but this reaction does not explain the chemical concentrations and isotopic characteristics of CH_4 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 CO_2, CH_4, and NH_3 (Ellis and Mahon, 1977). The close association of CH_4 with organic-rich sediments and the highly variable concentrations of CH_4 observed in volcanic emissions, hot springs, and fumaroles (Ward, 1978; Graeber et al., 1979; Reitsema, 1979) has detracted from the value of CH_4 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 ^{36}Ar 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 N_2/Ar ratio for the steam plus liquid outflows at Yellowstone was similar to the proportions expected from circulating meteoric water. The very high N_2/Ar ratios and excess H_2 found by Hulston and McCabe (1962) in New Zealand suggest an organic source for some of the N_2 and H_2.

Gaseous emanations in Iceland contain a few to tens of percent H_2. The high concentration of hydrogen is considered typical of active hydrothermal systems associated with volcanic rifts. Most of the geothermal H_2 may result from water disassociation at high temperature and pressure conditions (D'Amore and Nuti, 1977). The actual H_2 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 SiO_2 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 ($^3He/^4He$) 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 $^3He/^4He$ ratio and as the sample ratio normalized to the ratio in air:

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

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 $^3He/^4He$ 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 $^3He/^4He$ 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 $^{18}O/^{16}O$ and D/H in cold- and hot-spring waters. Plotting these ratios in parts-per-thousand ($^{o}/oo$) 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 $\delta^{18}O$ 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.

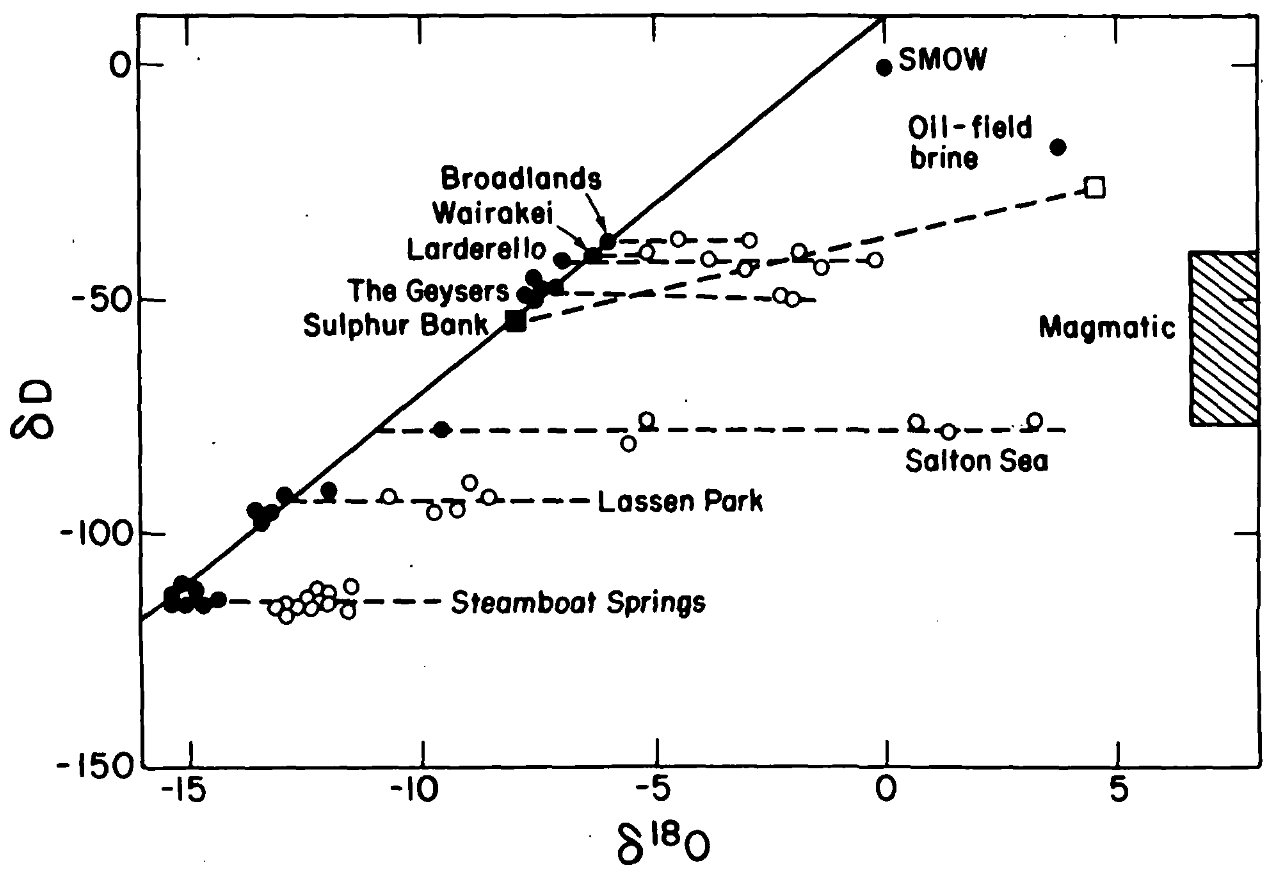


Figure 8. Values of $\delta D/\delta^{18}O$ for various meteoric waters, showing the meteoric water line (solid line) and the "oxygen shift" (broken lines). Filled circles and squares show points for meteoric waters. Open circles and squares show points for thermal waters (from Ellis and Mahon, 1977). (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 $\delta^{18}O$ 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 $^{o}/oo$ and $\delta^{18}O$ values of +3 to +5 $^{o}/oo$ 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, Na^+-K^+-Ca^{2+} relationship, and fractionation of the oxygen isotope $\delta^{18}O$ 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 (SiO_2 concentration in mg/kg)		
Quartz–no steam loss (conductive cooling)	$T(^\circ C) = \frac{1309}{5.19 - \log SiO_2} - 273.15$	0 – 250 °C
Quartz–maximum steam loss (adiabatic cooling)	$T(^\circ C) = \frac{1522}{5.75 - \log SiO_2} - 273.15$	0 – 250 °C
Chalcedony	$T(^\circ C) = \frac{1032}{4.69 - \log SiO_2} - 273.15$	0 – 250 °C
α–Cristobalite	$T(^\circ C) = \frac{1000}{4.78 - \log SiO_2} - 273.15$	0 – 250 °C
β–Cristobalite	$T(^\circ C) = \frac{781}{4.51 - \log SiO_2} - 273.15$	0 – 250 °C
Amorphous silica	$T(^\circ C) = \frac{731}{4.52 - \log SiO_2} - 273.15$	0 – 250 °C
Alkali Geothermometers (Na, K, and Ca concentrations in mg/kg)		
Na/K (Fournier)	$T(^\circ C) = \frac{1217}{\log(Na/K) + 1.483} - 273.15$	>150 °C
Na/K (Truesdell)	$T(^\circ C) = \frac{855.6}{\log(Na/K) + 0.8573} - 273.15$	>150 °C
Na-K-Ca	$T(^\circ C) = \frac{1647}{\log(Na/K) + \beta^* [\log(\sqrt{Ca}/Na) + 2.06] + 2.47} - 273.15$	

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

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

†Throughout this report T is the only symbol used for temperature (°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/cm^3) 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 (^{222}Rn), a gaseous daughter product of ^{238}U, 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 ^{222}Rn (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.

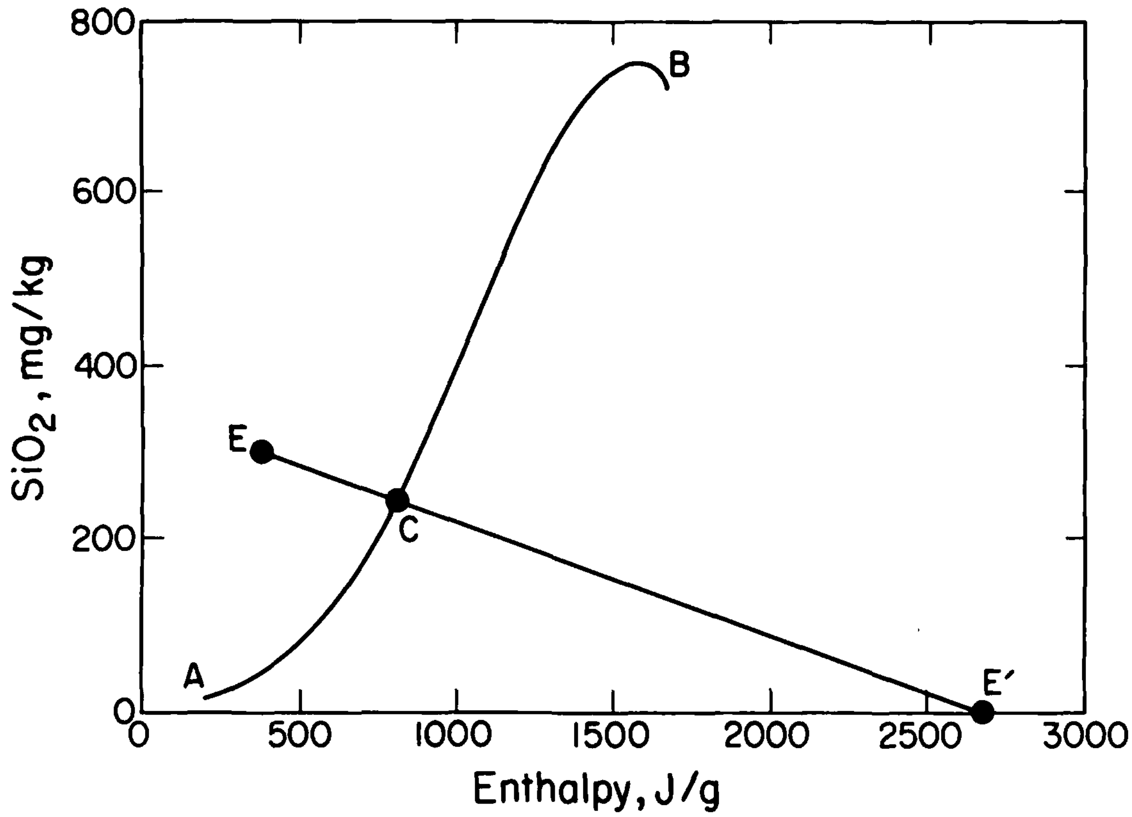


Figure 9. Enthalpy/silica-concentration diagram used for correcting silica concentrations for adiabatic cooling, assuming single-stage steam loss at any temperature. Point *C* represents the silica concentration and enthalpy of the original water. See text for meaning of *AB* and *EE'* lines (from Truesdell 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/mm^2) 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 *A* 1 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).

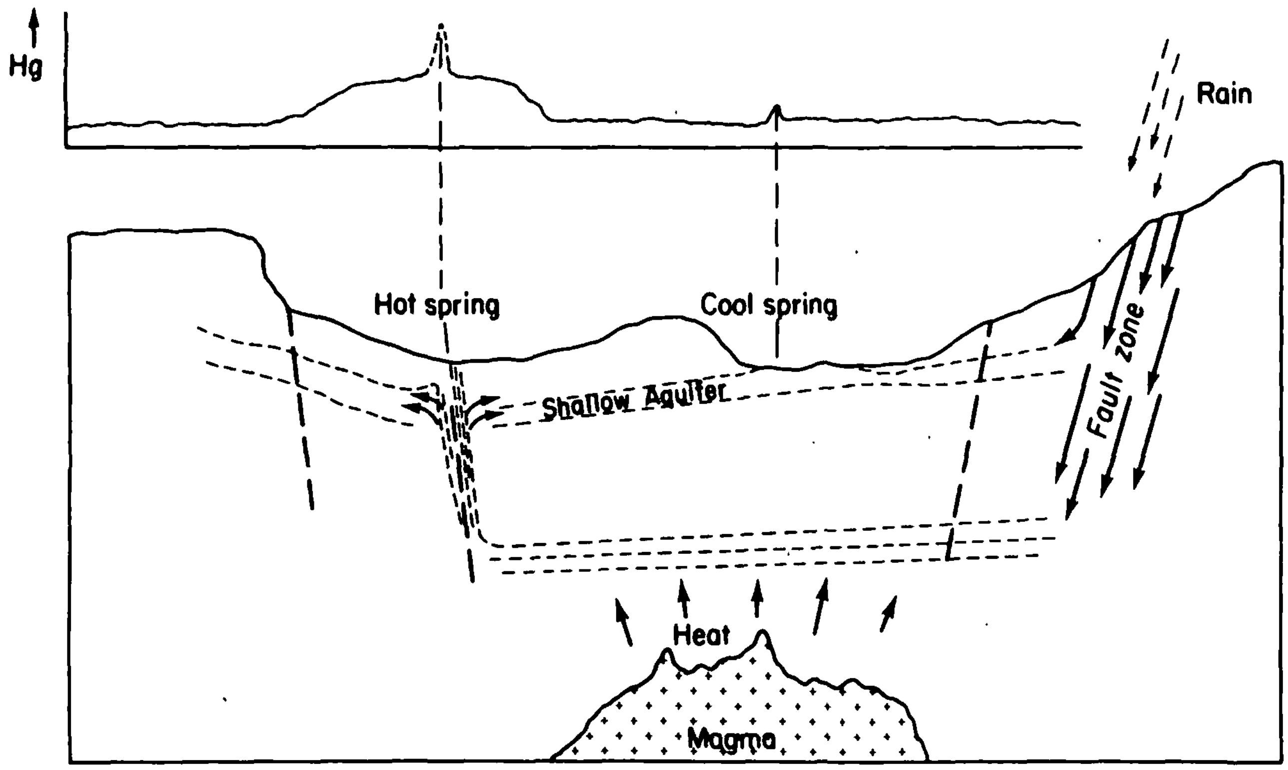


Figure 10. Schematic of a Hg anomaly at a geothermal area, showing peak Hg concentration (top curve) above a zone of ascending hot water (from Varekamp and Buseck, 1983). (XBL 8312-2442)

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-km^2 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.

^{4}Helium

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 C^{3+} 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.

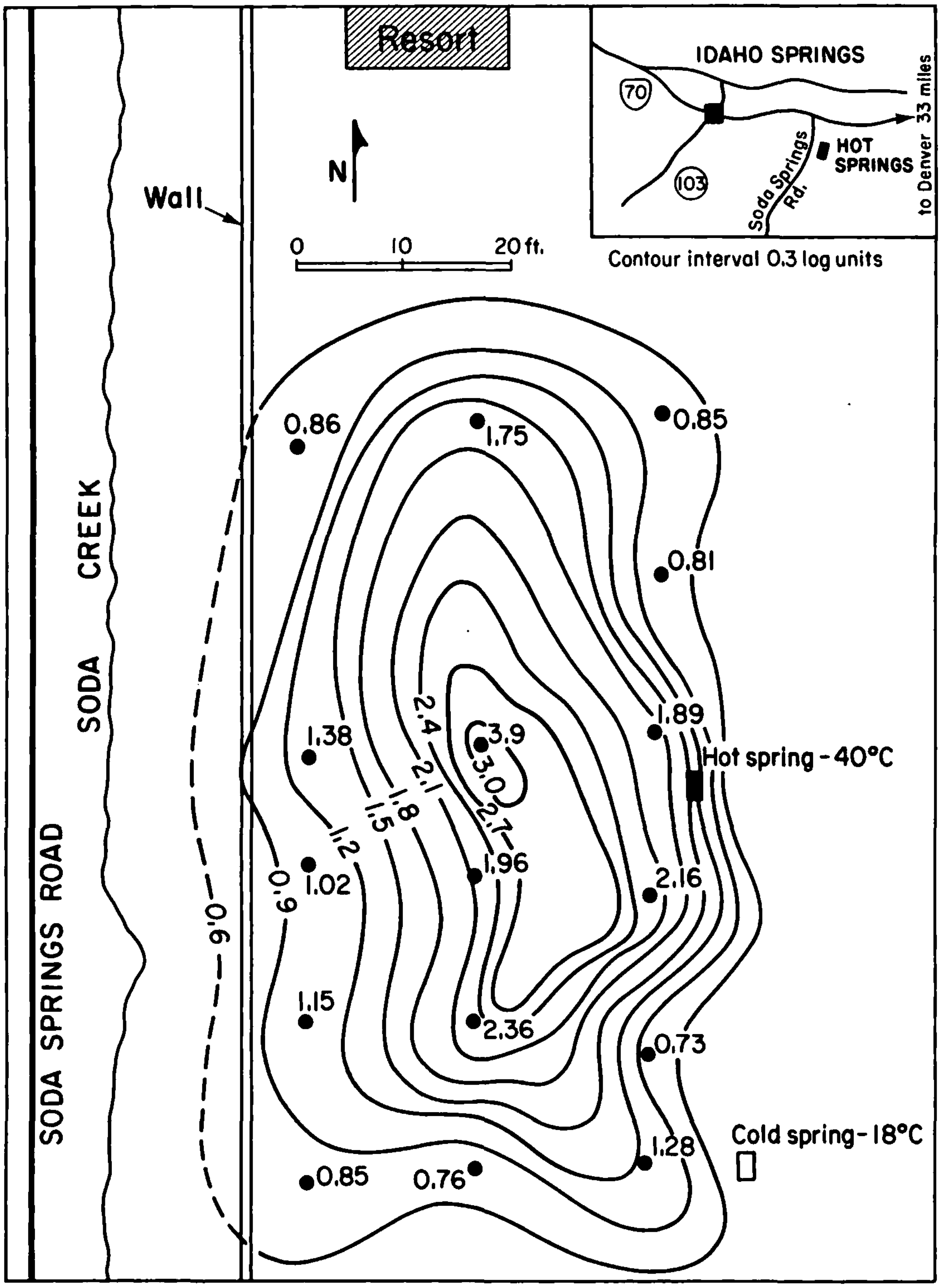


Figure 11. Logarithm of ^{4}He isopleths near a 40 ° C hot spring in Colorado (from Roberts et al., 1975). Notice that the peak of the helium anomaly is displaced from the springs.
(XBL 841-397)

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):

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

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.

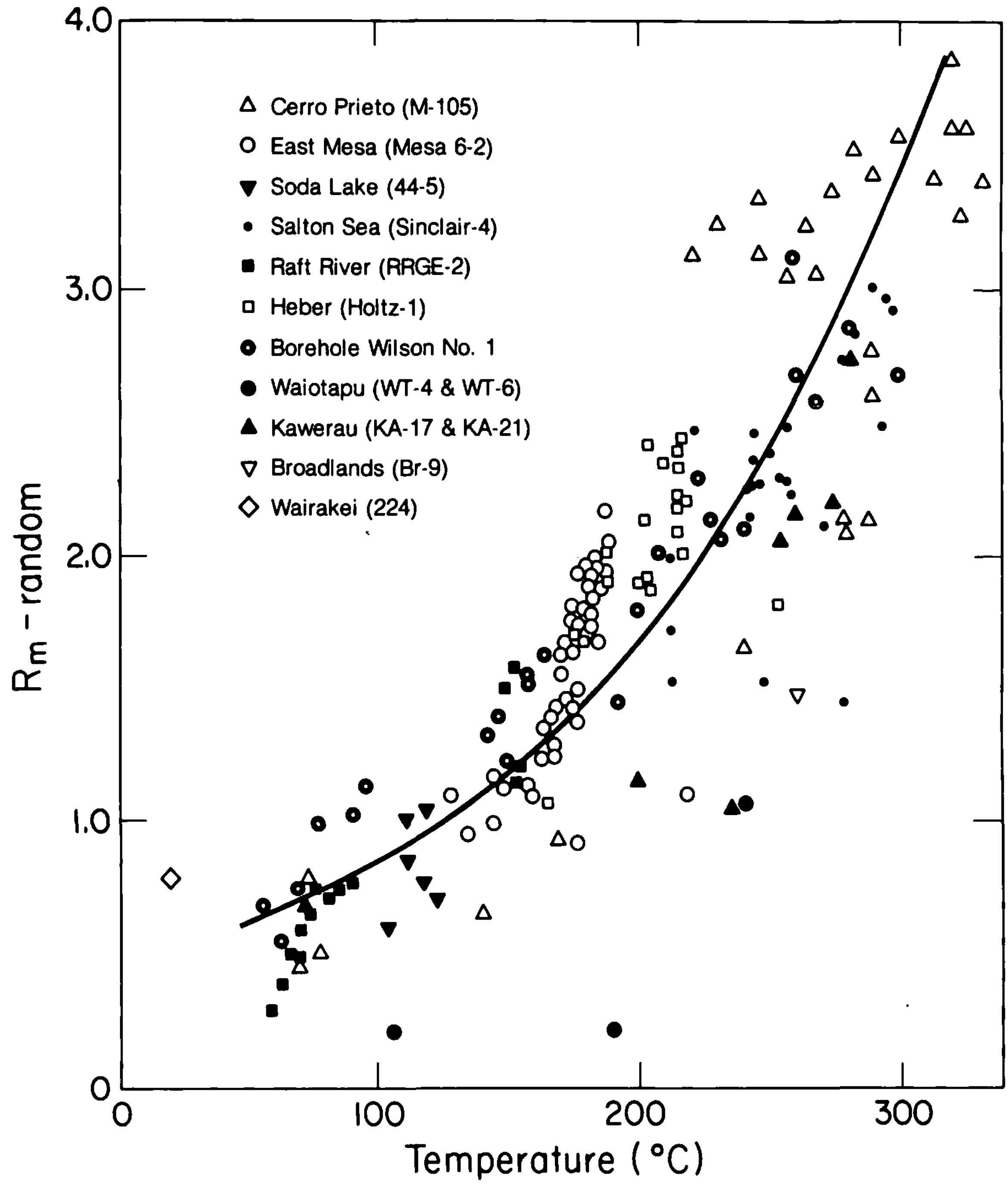


Figure 12. Vitrinite reflectance (R_m) versus temperature for several geothermal areas. The curve describes the regression equation $R_m = 0.435 \exp 0.00683 T$ ($^\circ$C) (from Barker, 1983).
(XBL 8312-2444)

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.,

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

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},

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

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}cal $\cdot$ $cm^{-2} \cdot s^{-1}$ = 41.8 mW $\cdot$ 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

$m \;=\; \sqrt{\pi/\tau\alpha}$,
$\omega \;=\; 2\pi/\tau$,
τ = the period of the wave = 1 year,
Δt = the elapsed time since the summer maximum,
T_a = seasonal temperature amplitude,
$\alpha = \kappa/c\,\delta$ = thermal diffusivity,
κ = thermal conductivity,
c = specific heat of the rock, and
δ = 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 Häenel, 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

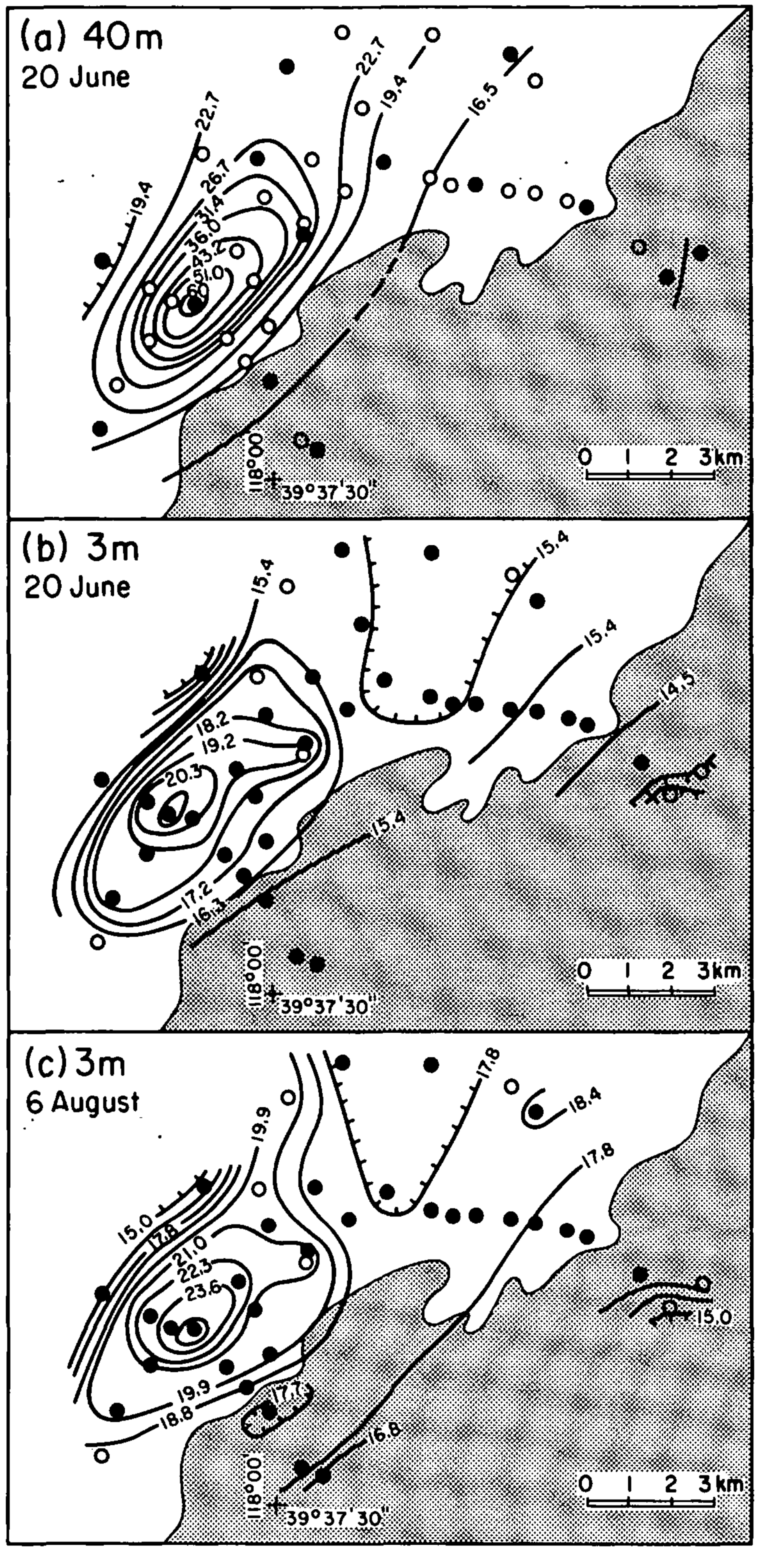


Figure 13. Thermal anomaly mapped at the east side of Dixie Valley, Nevada. (a) Corrected temperatures at the bottom of 40-m-deep holes. (b,c) Corrected temperatures estimated from 3-m-deep holes on two separate days (from Lange et al., 1982).
(XBL 841-9517)

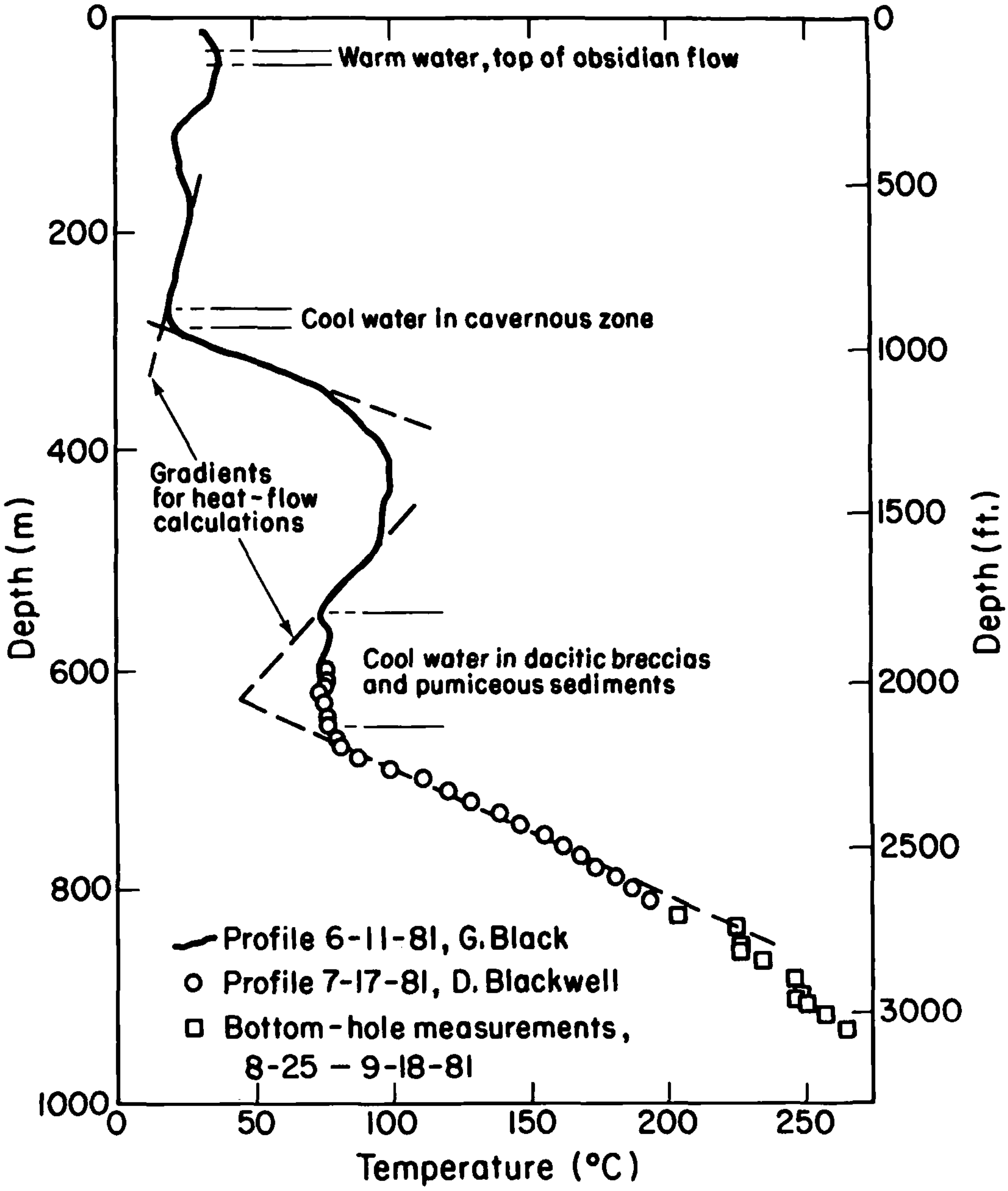


Figure 14. Temperature profile from a USGS test hole at the Newberry Volcano, Oregon. Note the linear temperature gradient below 675 m (from Sammel, 1983). (XBL 841-9518)

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 $\sim$ 2 km, together with low-level surveys flown at constant terrain clearance with a line separation of $\sim$ 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, Souix 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 $x\,Fe_2TiO_4 \cdot (1-x)\,Fe_3O_4$ 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° × 1° 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 ferrimagnetic 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/cm^3 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 volcaniclastic 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 ($> 250^\circ$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 μGal (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 μGal 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 m, but this accuracy level could introduce an uncertainty of $\pm$ 0.2 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 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.

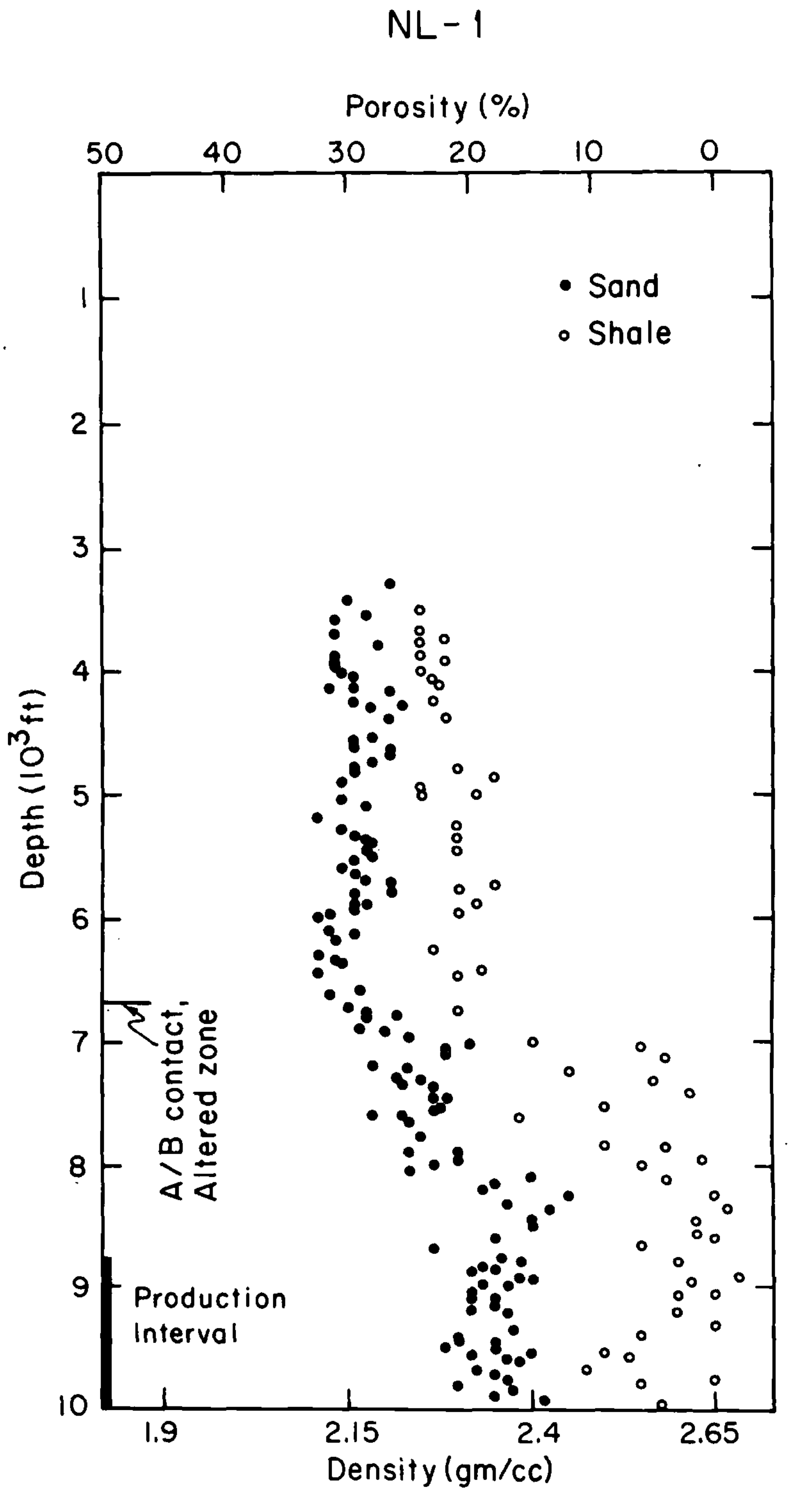


Figure 15. Density versus depth for a geothermal well at the Cerro Prieto geothermal field, Baja California. Note that the density values of both shale and sand are higher in the hydrothermally altered sediments (B) below the A/B contact than they are in the relatively fresh sediments (A) above the A/B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

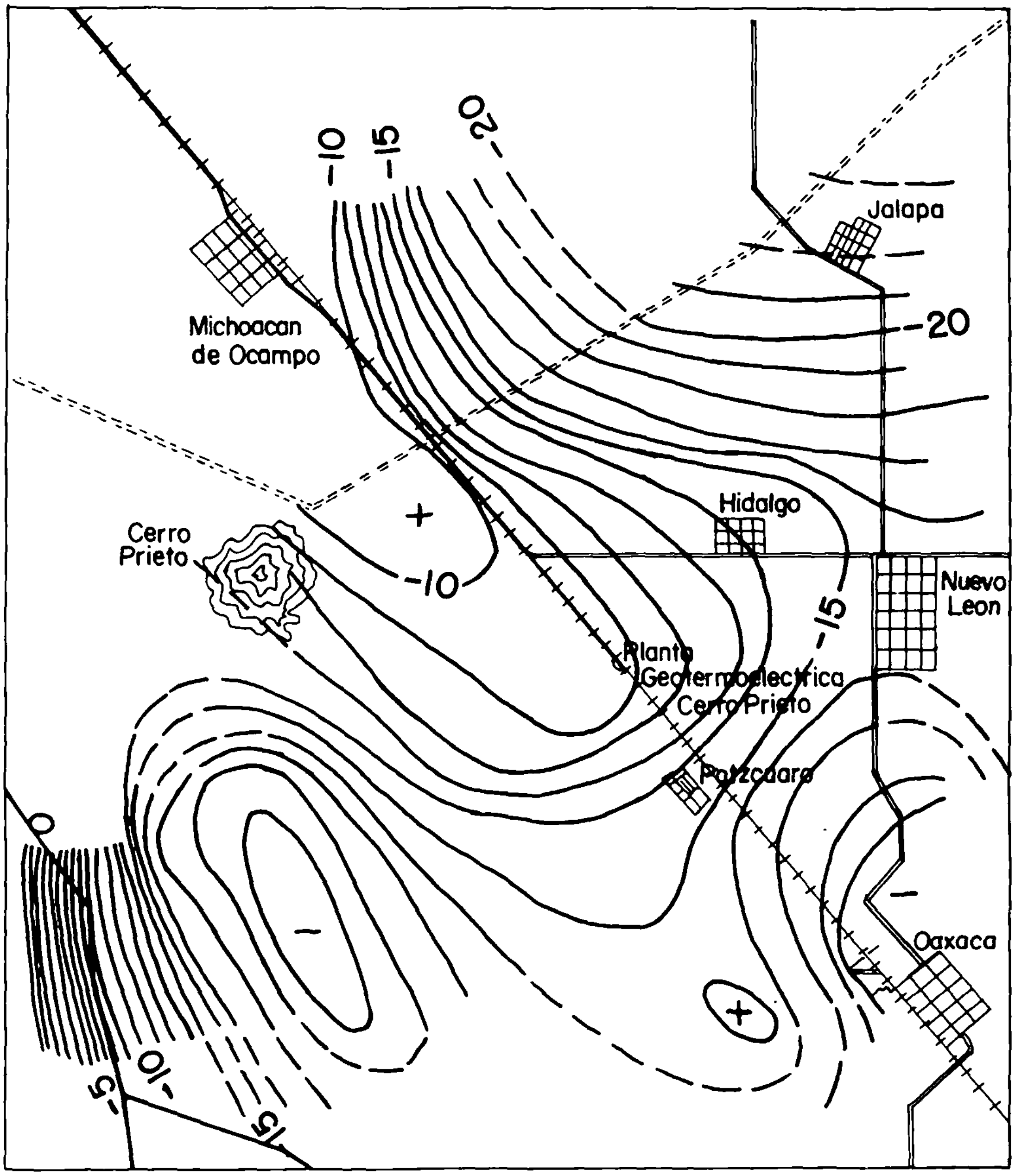


Figure 16. Depth (meters) to the top of the densified zone beneath the East Mesa area, Imperial Valley, California. The mean density contrast of 0.14 g/cm^3 between the anomalous source and the host rocks is attributed to authigenic quartz, sulfides, and calcite deposited as a result of thermal waters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

$$\rho_0 = a \ \rho_w \ \phi^{-m} , \tag{3}$$

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

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

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·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. Ucok 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 $\Lambda = B_0 - B_1 C^{1/2} + B_2 C \ln C$ + 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, deuteric 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).

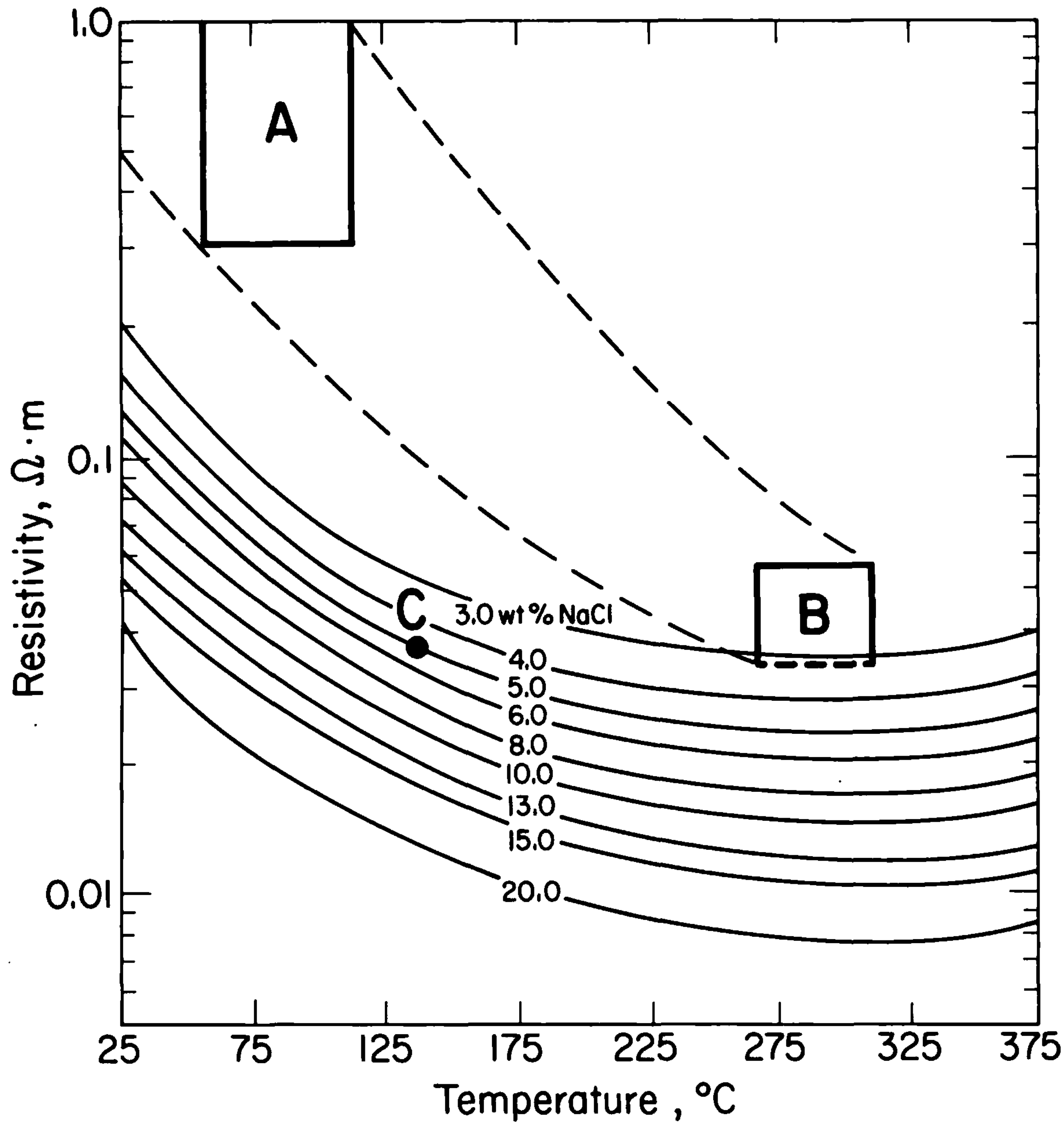


Figure 17. Resistivities of NaCl solutions as a function of temperature. Solution concentrations are typical of geothermal brines. See text for the explanation of A, B, and C (from Ucok, et al., 1980).
(XBL 8312-2445)

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 Krívvik area, Iceland (Arnórsson et al., 1976). In both these areas, the hydrothermally altered and conductive ($\leq$ 10 ohm·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 ohm·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 nonresistive. 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:

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

where ρ, ϵ, and η are electrical resistivity, dielectric constant, and viscosity of the fluid, respectively; ΔP is the pressure drop; and ς 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 ς. They found that ς 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 ς increases with temperature due to H^+ desorption from the solid. They concluded that the streaming potential coefficient $\rho\epsilon\varsigma/\eta$ 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).

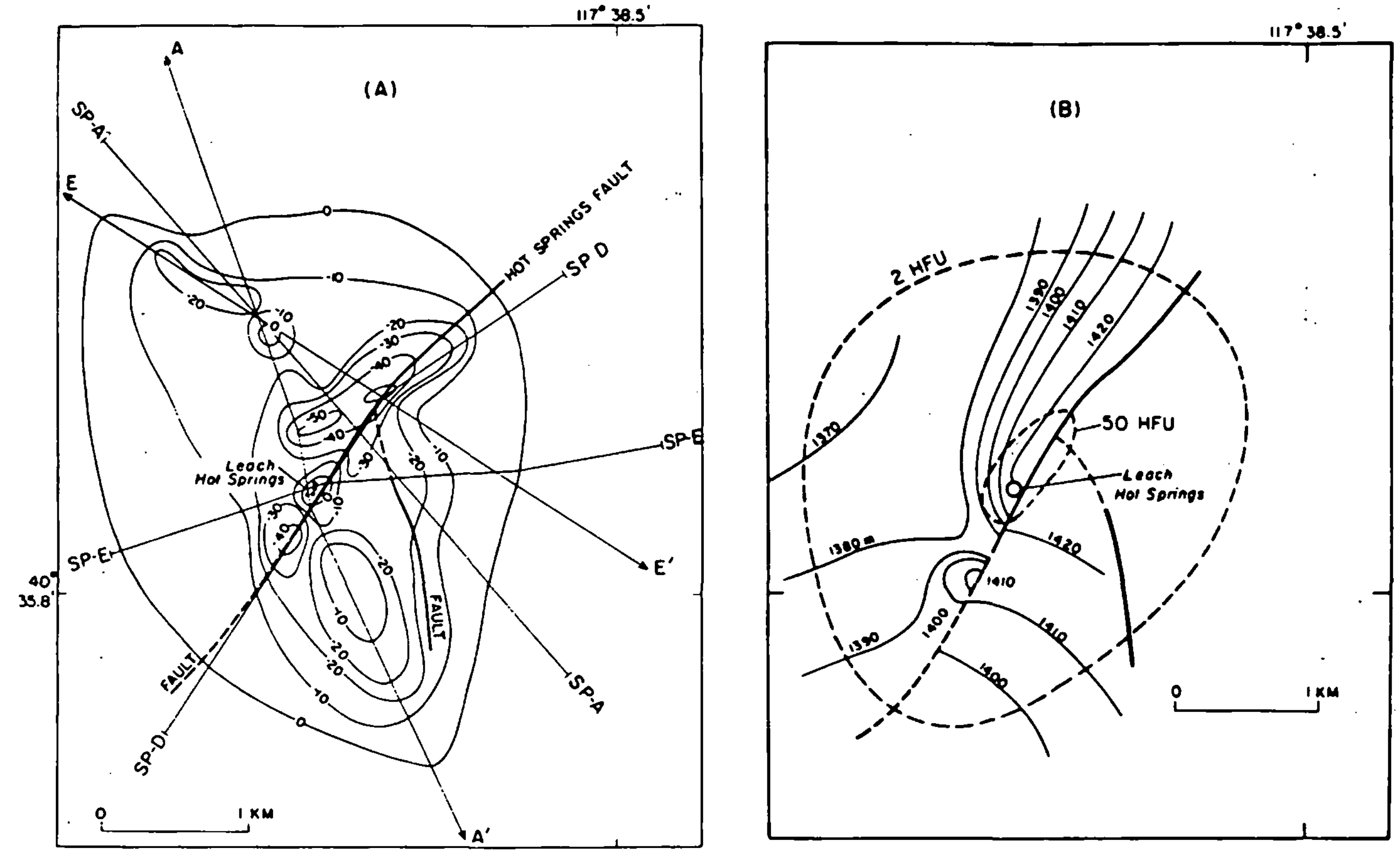


Figure 18. (a) Self-potential distribution in Leach Hot Springs area, Grass Valley, Nevada, based on smoothed data taken in September 1975. SP-A, SP-D, SP-E, A-A′ , and E-E′ are traverse lines along which measurements were made. Electrode spacing was usually 100 m; contour interval is 10 mV. Faults dashed 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)

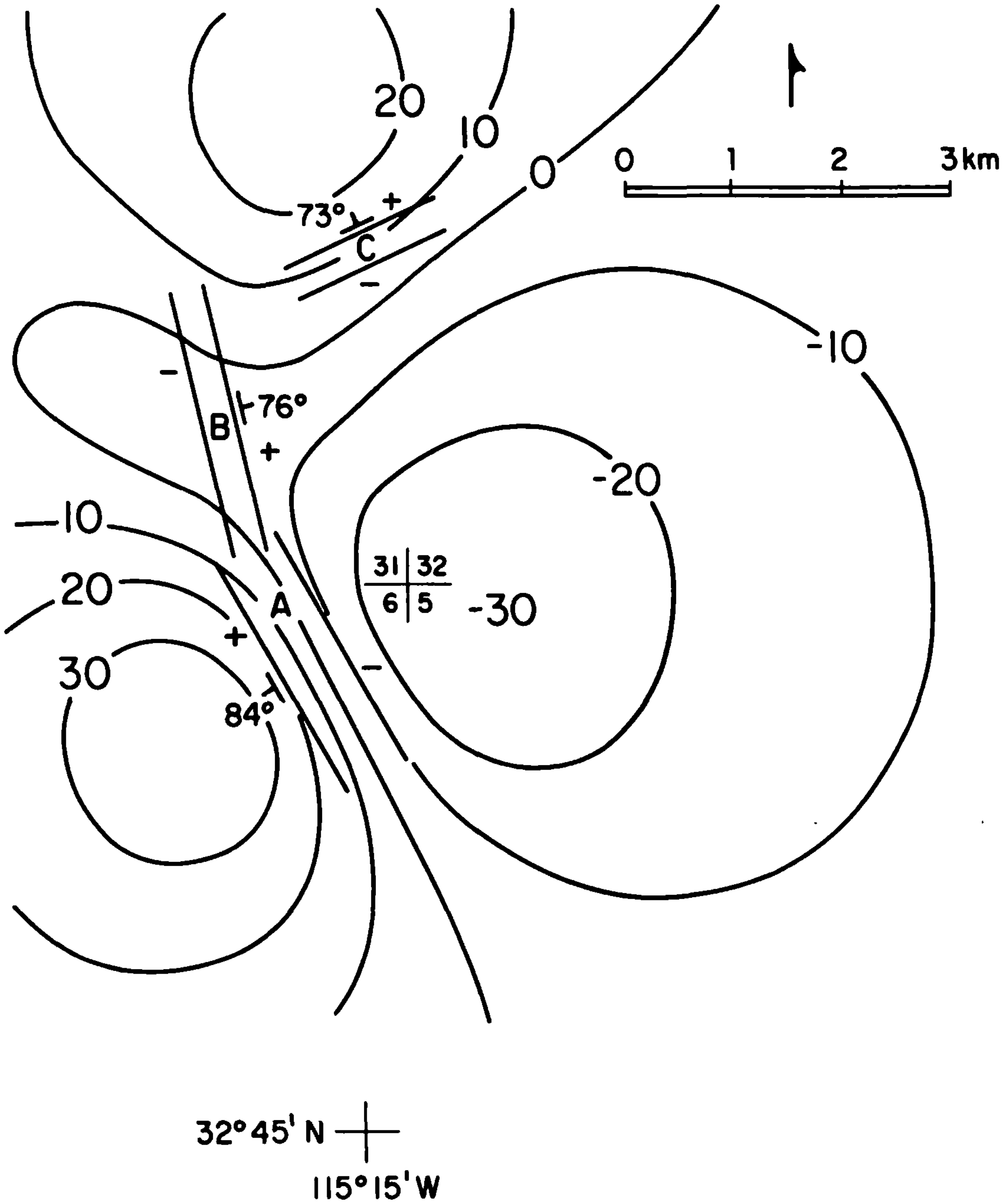


Figure 19. Smoothed self-potential anomaly over the East Mesa geothermal field, California. The axis of the dipole anomaly (A) coincides with the location of the hotter wells and a set of northwest-trending faults (from Corwin et al., 1981). (XBL 8312-2448)

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 Yakeyama 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 discusson to a brief outline and examples of results.

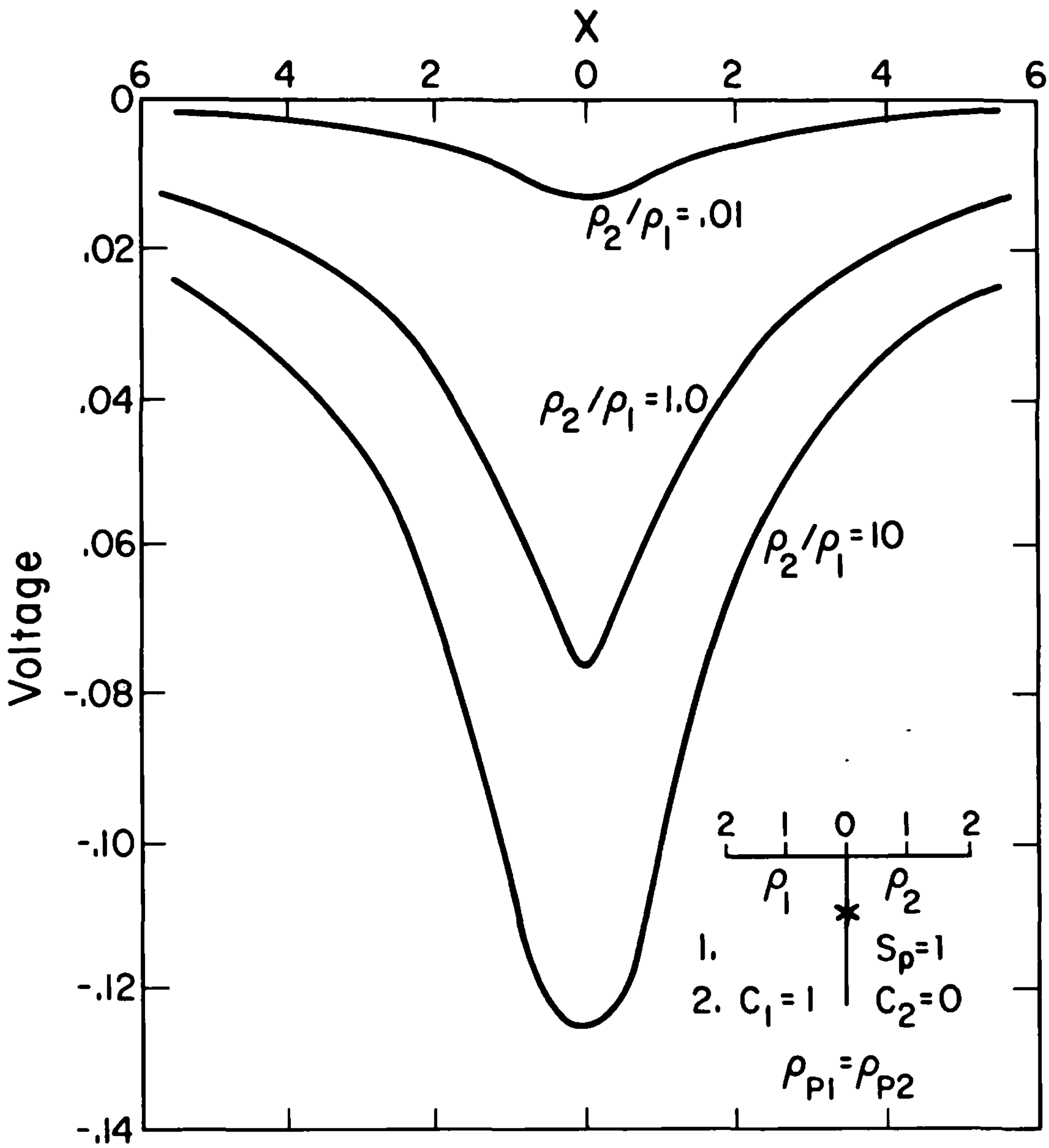


Figure 20. SP surface-voltage curves of a symmetric negative anomaly caused by a point pressure source at a vertical resistivity boundary. ρ = electrical resistivity, ρ_P = hydraulic resistivity or impermeability, and C = the voltage-coupling coefficient, which relates forces to current flow and can be determined experimentally for laboratory models (from Sill, 1983).
(XBL 841-9514)

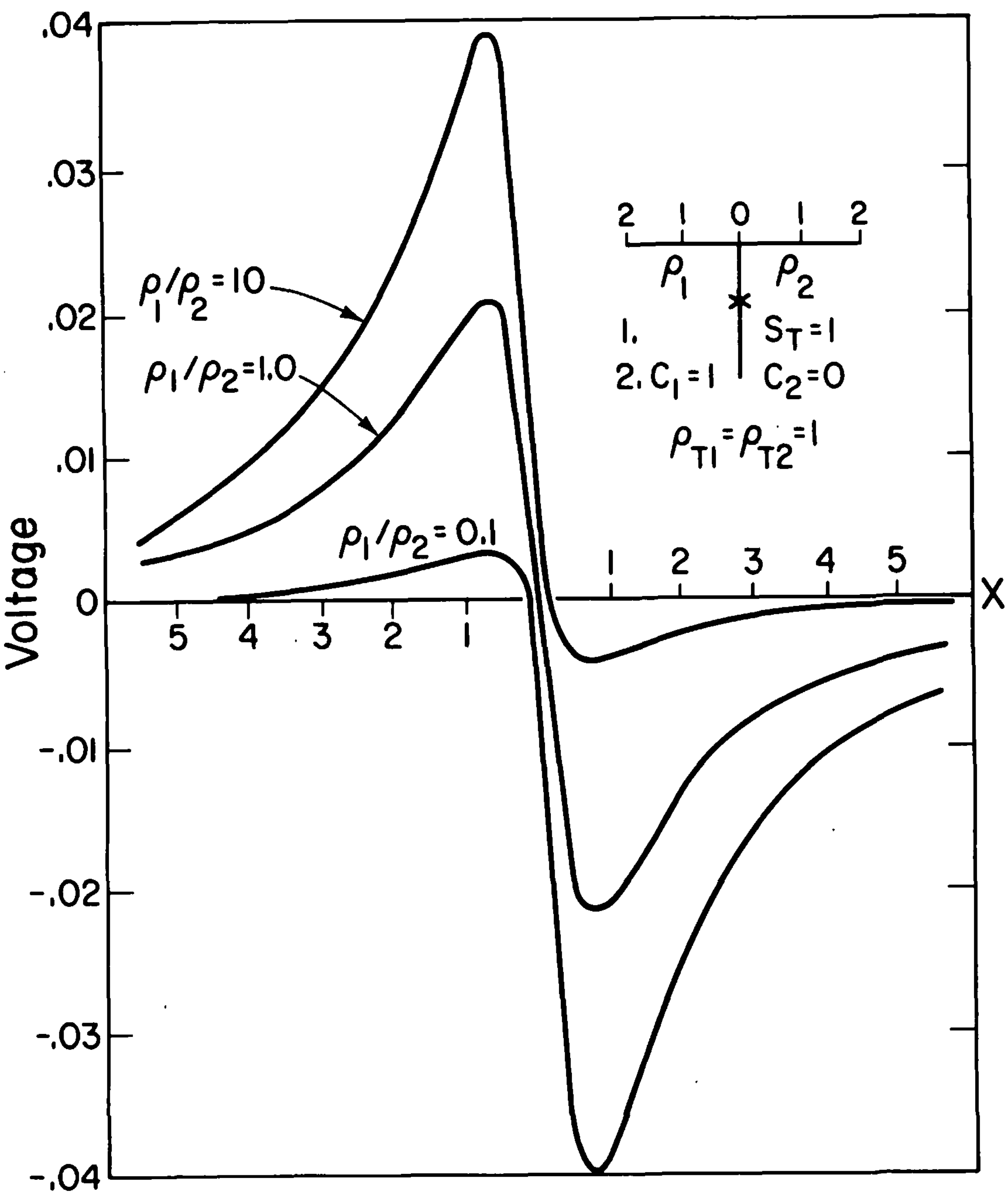


Figure 21. SP surface-voltage curves of a dipolar anomaly caused by a point temperature source at a vertical resistivity boundary. ρ = electrical resistivity, ρ_T = thermal resistivity, and C = the voltage-coupling coefficient, which relates forces to current flow and can be determined experimentally 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

$$S = \sum_{i}^{n} (h_i / \rho_i) \tag{5}$$

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.

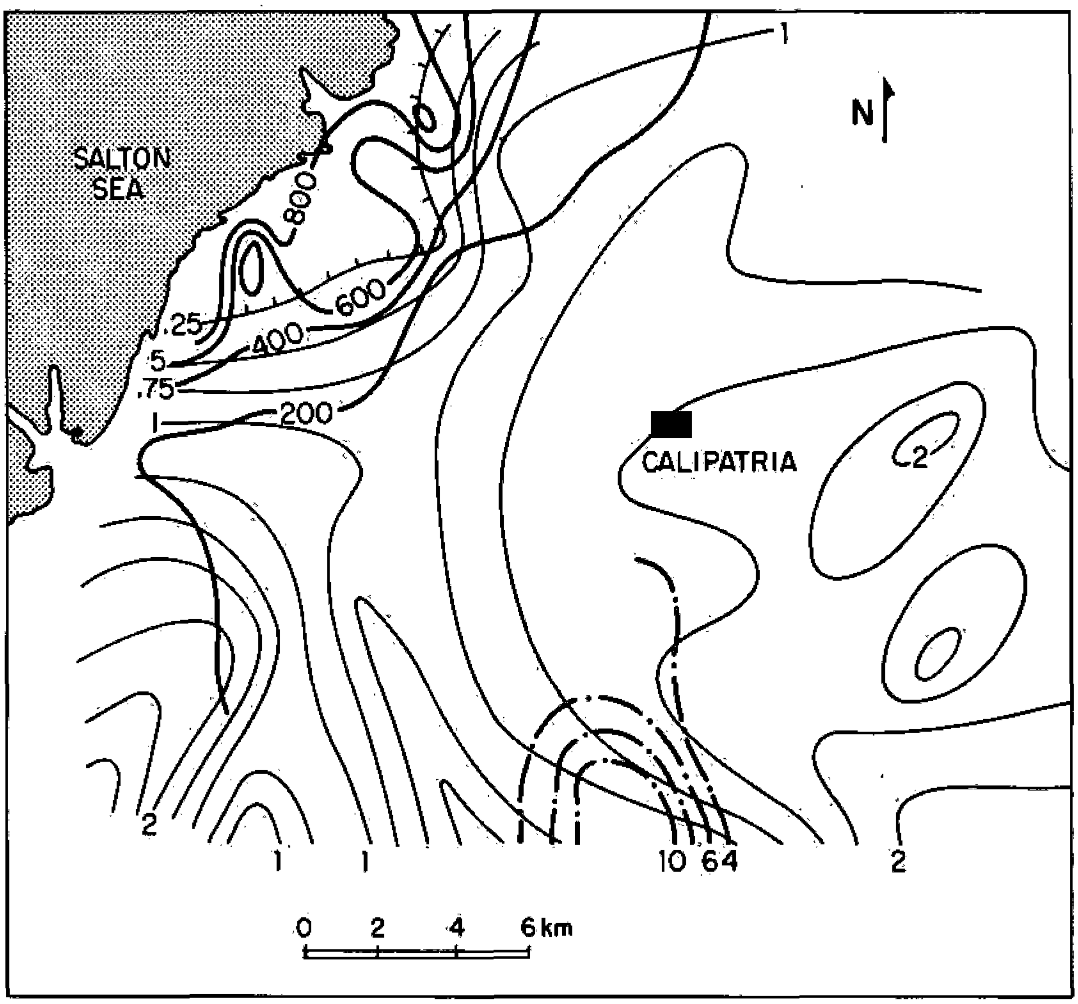


Figure 22. A plot of telluric J values at 67-second period (light solid lines) versus heat flow at the south shore of the Salton Sea geothermal field (after Humphreys, 1978). The heavy solid contours indicate heat flow in mW/m^2 (from Lee and Cohen, 1979). The dash-dot lines are temperature gradients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

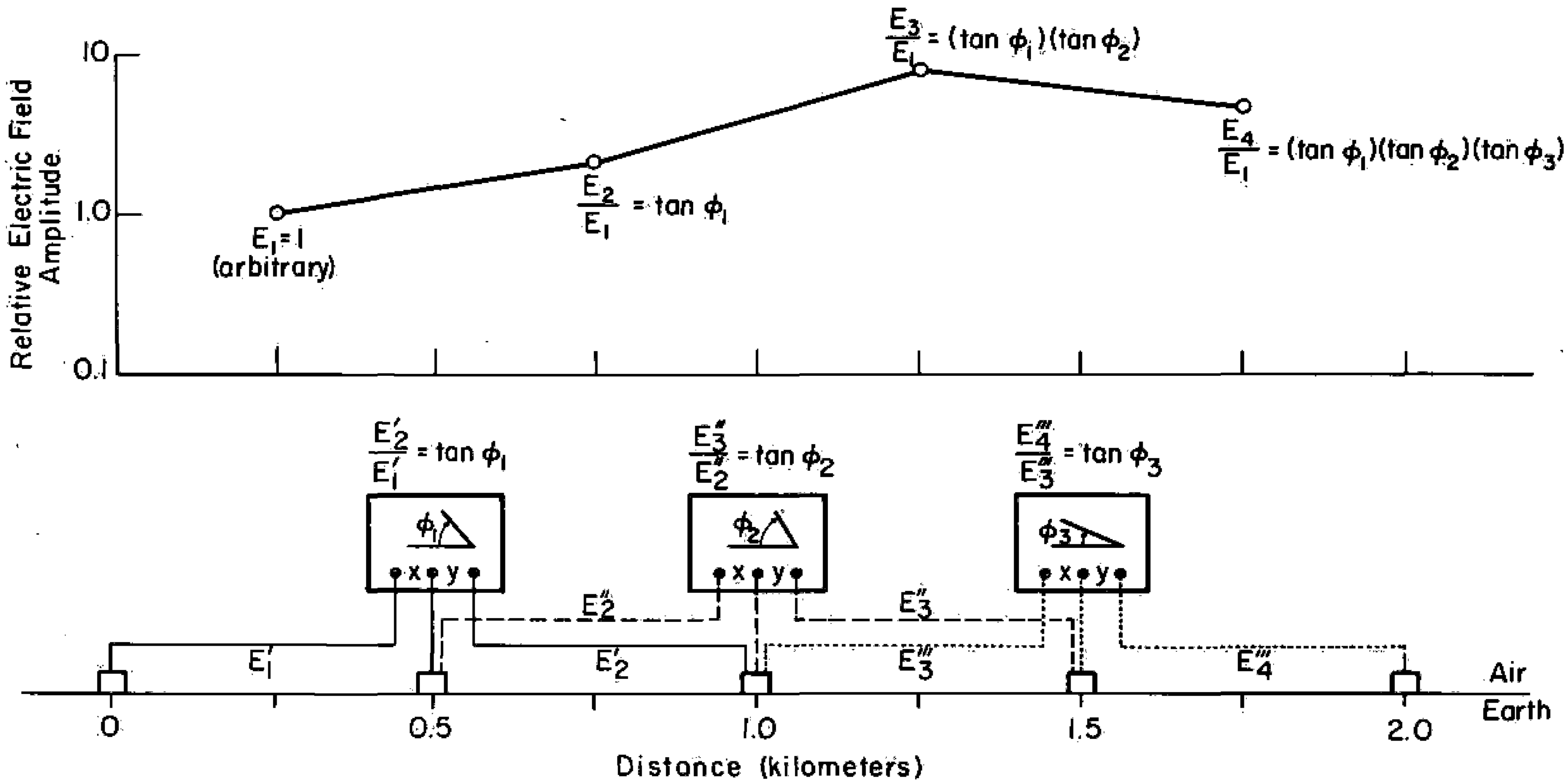


Figure 23. The in-line telluric method. Slopes indicate relative amplitude differences between points on the array. Maxima and minima indicate changing subsurface resistivities within the subsurface region being studied. The depth of the region is an inverse function of the frequency of the wave recorded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

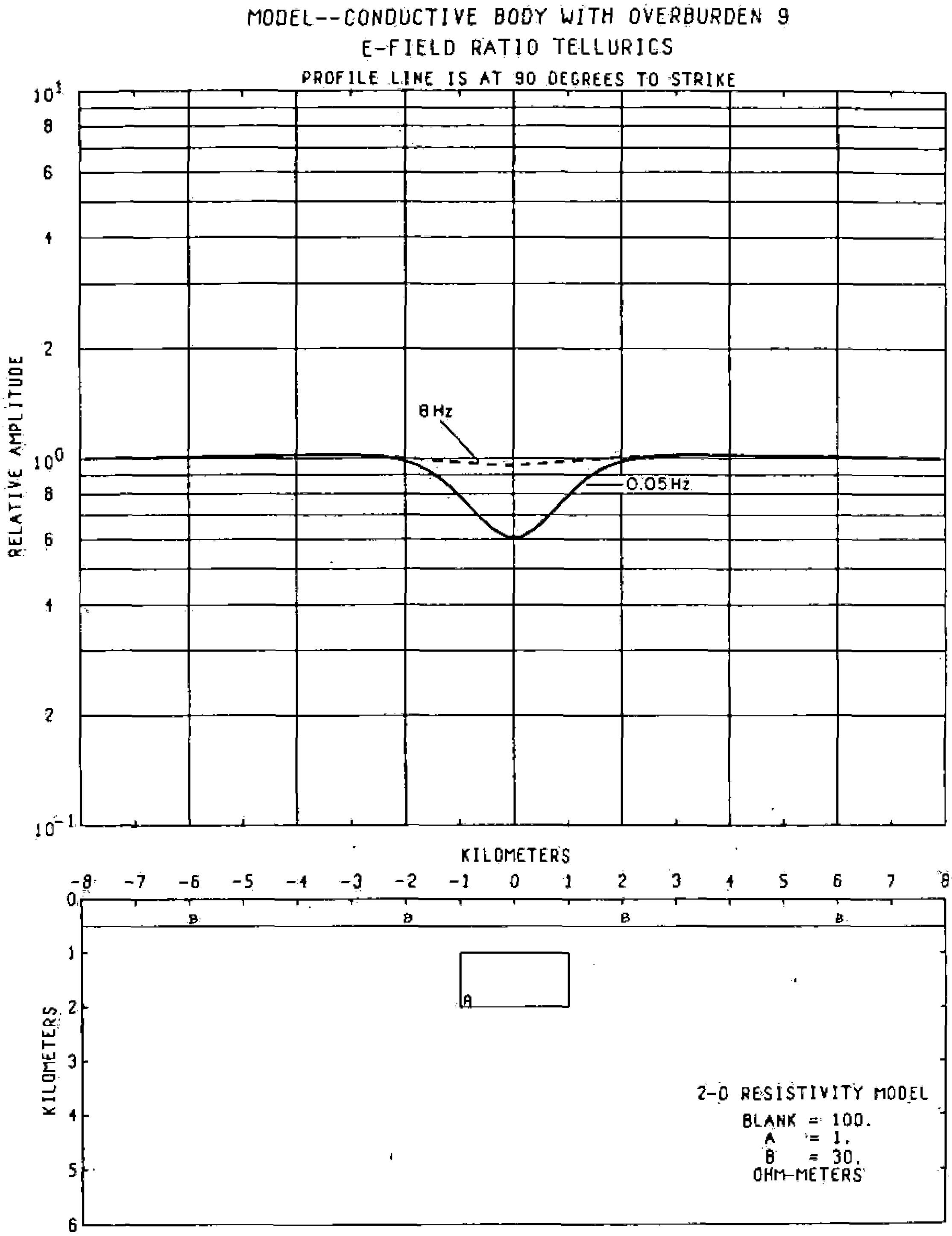


Figure 24. Calculated in-line telluric anomalies at 0.05 and 8 Hz caused by the concealed 2-D conductor beneath an overburden.
(XBL 833-8695)

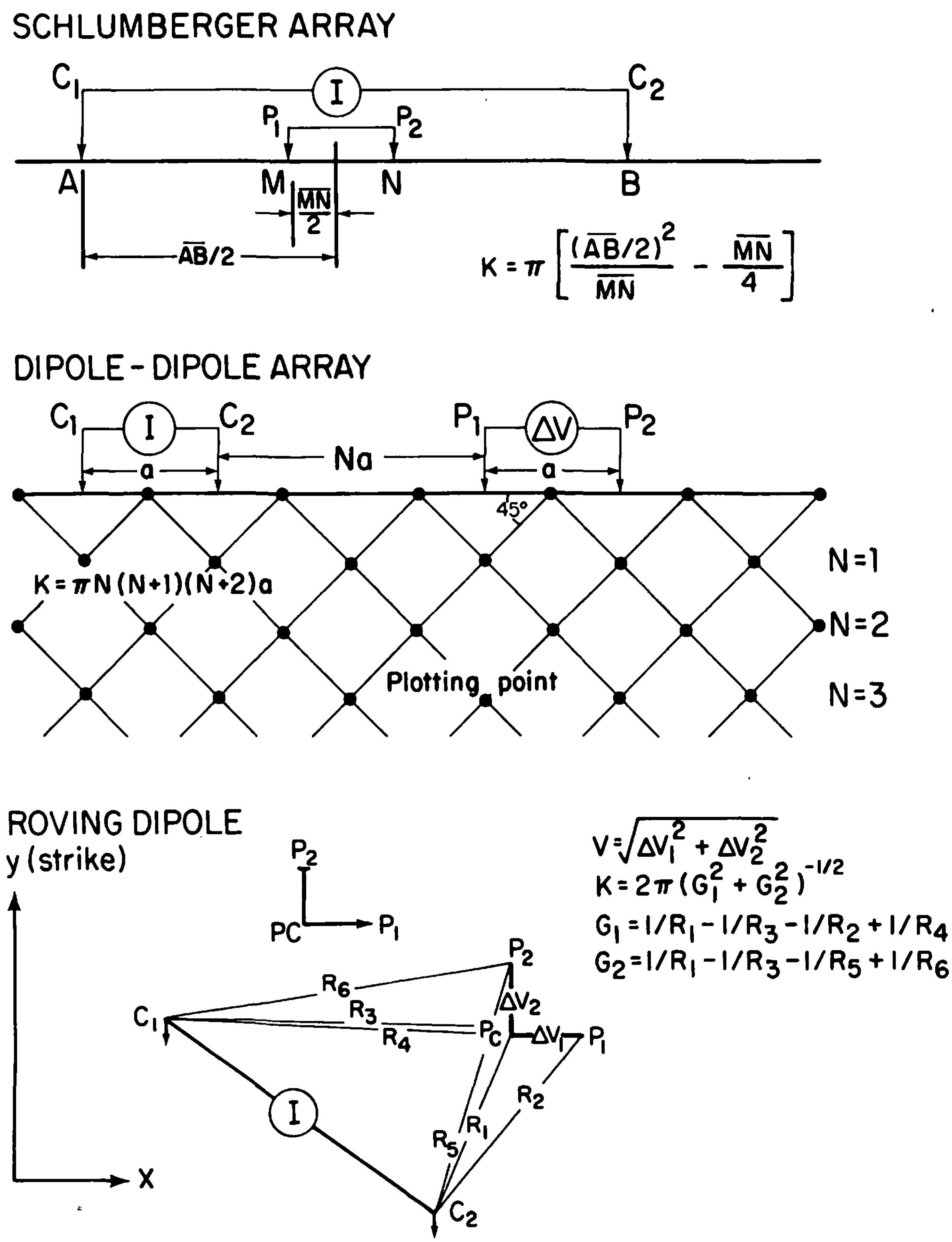


Figure 25. Diagrams of three common methods for conducting dc-resistivity surveys. (XBL 833-1740)

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 (Flóvenz 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 ith layer. Depending on the sounding curve, there can exist a range of values ρ_i and h_i such that $\rho_i h_i$ = constant (T equivalence) or h_i / ρ_i = 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-km^2 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 ET1 and ET2, which were then combined to produce a resultant voltage ET. By changing the ratio of ET1 to ET2, the resultant ET was caused to rotate through 360 degrees, tracing out a field ellipse. The arithmetic

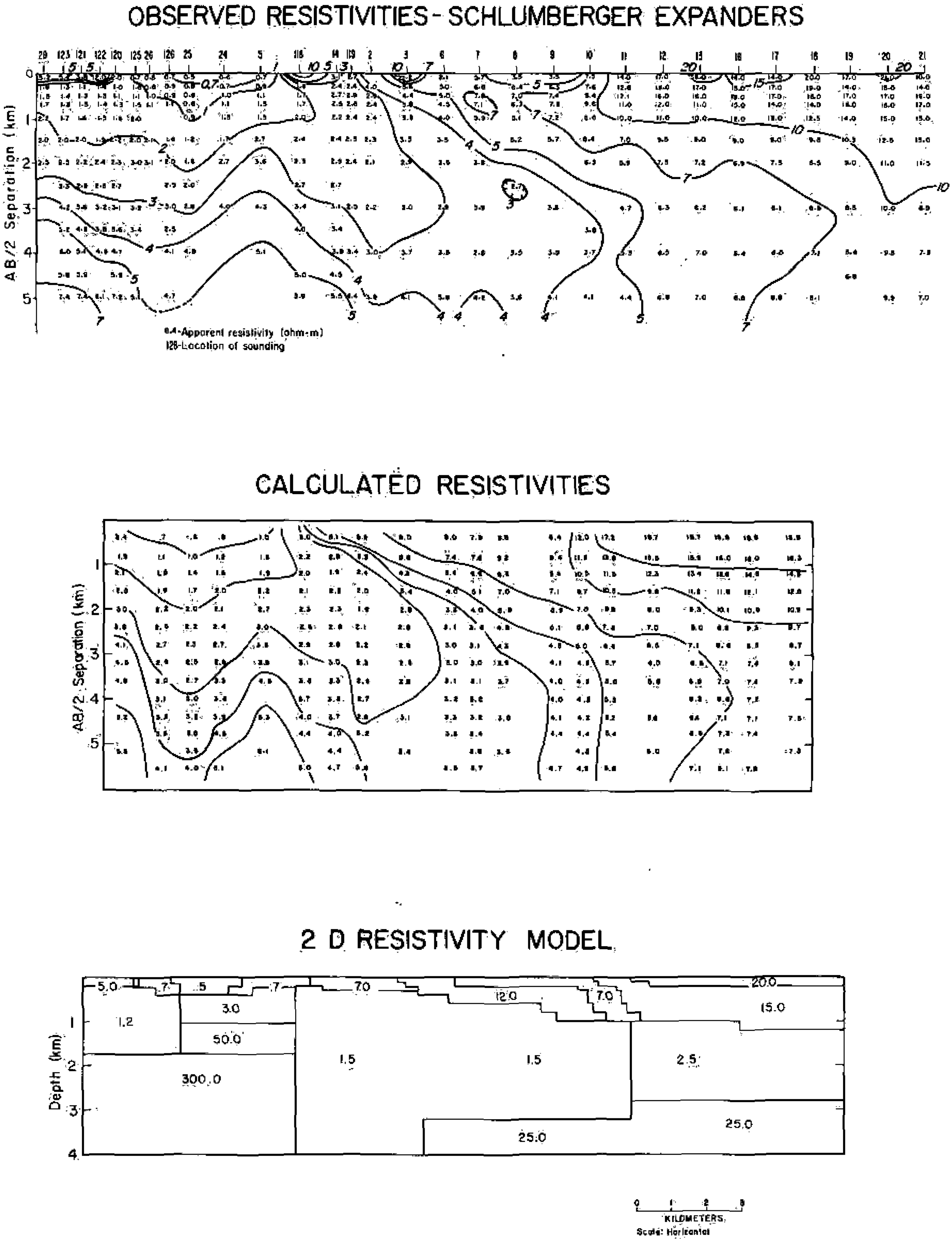


Figure 26. Observed and calculated dc resistivities and 2-D resistivity model of the Cerro Prieto geothermal field, Baja California. The observed resistivities were obtained by Schlumberger expanders. The calculated resistivities were generated by a finite-difference model (from Wilt et al., 1978). (XBL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

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

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

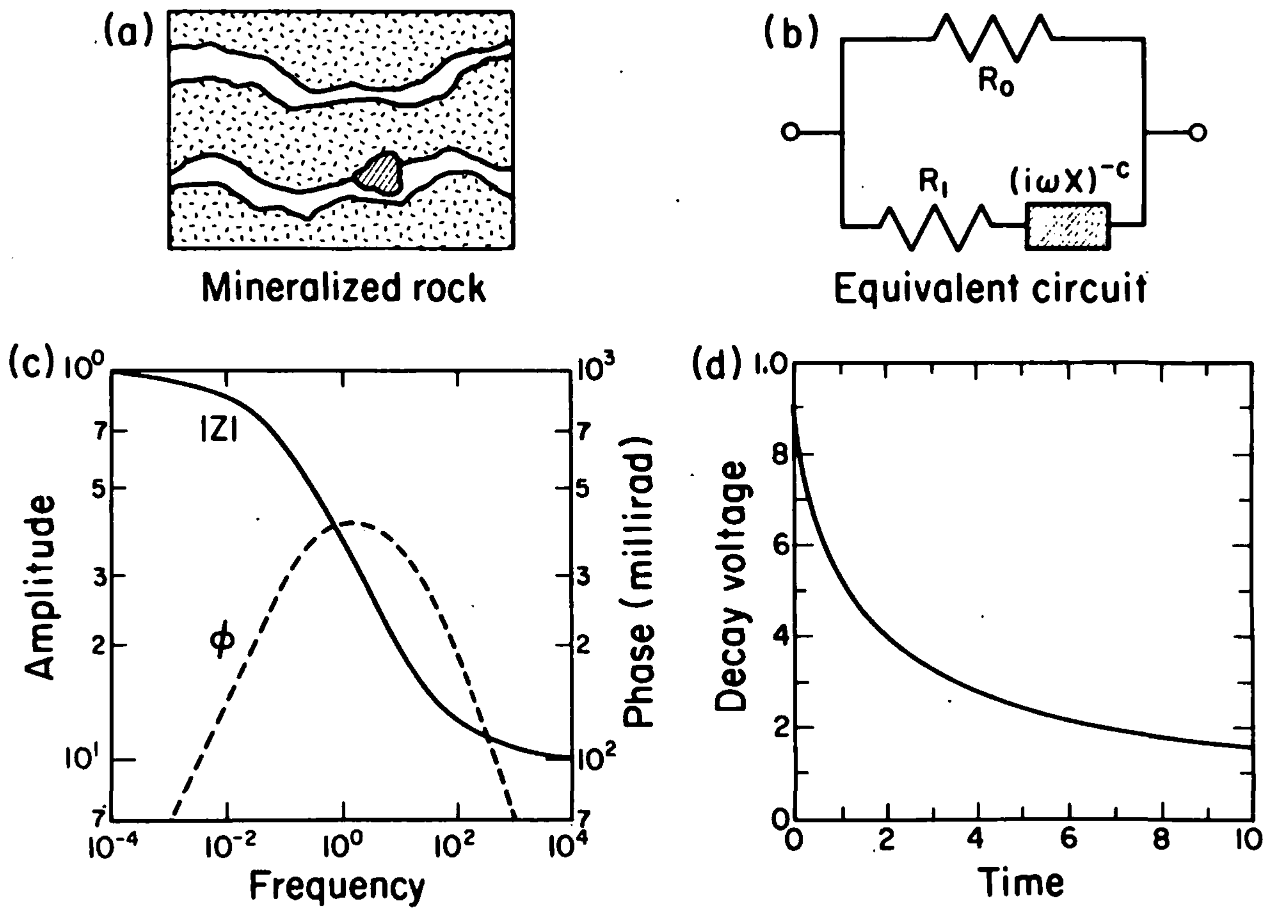


Figure 27. Examples of frequency domain spectra and time domain decay voltages caused by the presence of conductive sulfide grains along ionic flow paths. (XBL 8312-2446)

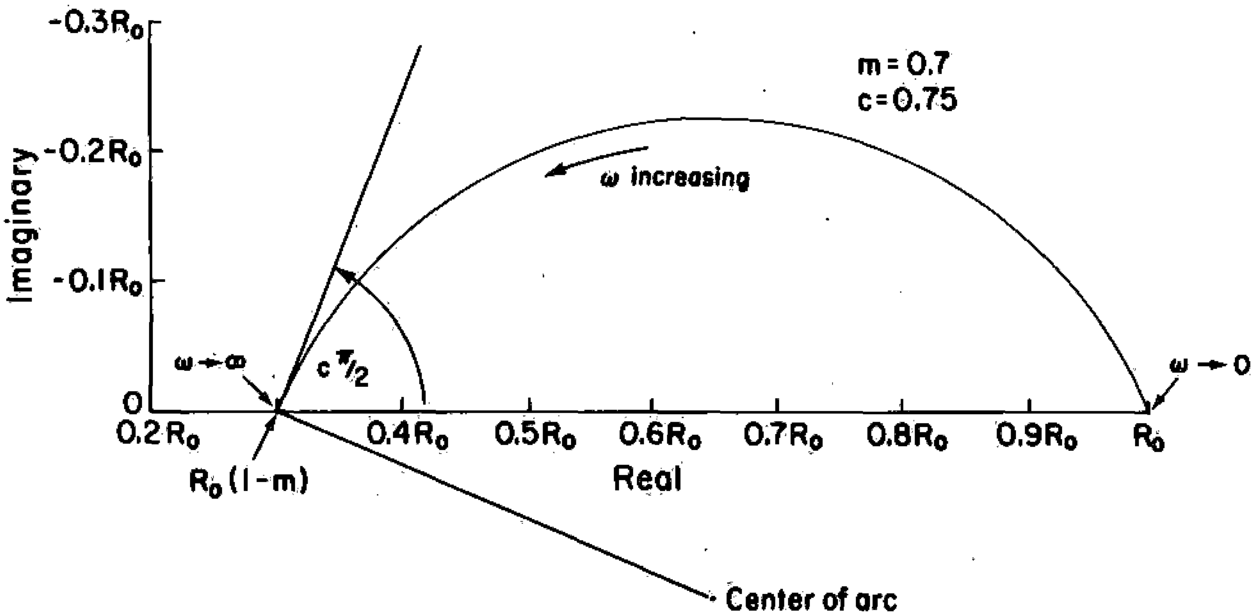


Figure 28. Argand diagram of the voltage measured in Figure 24c. R_0 = dc resistivity value, $Z(0)$; m = chargeability parameter; c = frequency-dependent term; ω = angular frequency.
(XBL 8312-2447)

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 (Keevil and Ward, 1962; Olorunfemi and Griffiths, 1985) that saline pore water (e.g., 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,

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

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):

$$\delta = 0.5 T \rho. \tag{7}$$

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

$$E_x = Z_{xx}\, H_x + Z_{xy}\, H_y, \tag{8}$$

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

Equation (8) is usually shown in tensor notation as

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

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 $\underset{\approx}{Z}$ are minimized (Z_{xx}, Z_{yy} ≈ 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:

$$\rho'_{xy} = \frac{0.2T}{f} \mid Z'_{xy} \mid^2 \tag{9a}$$

and

$$\rho'_{yx} = \frac{0.2T}{f} \mid Z'_{yx} \mid^2, \tag{9b}$$

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

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

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

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).

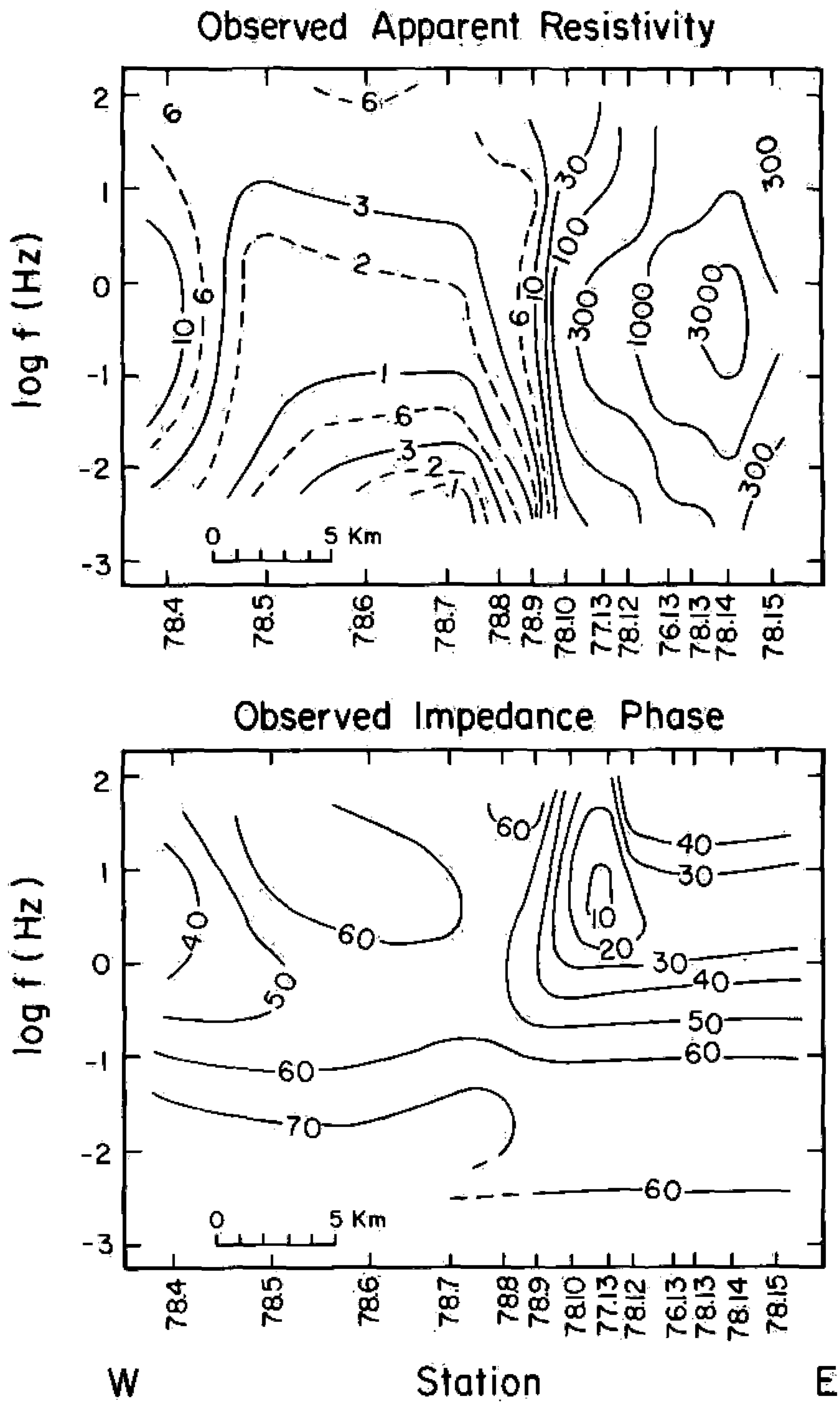


Figure 29. Observed psuedosections of apparent resistivity (ρ_{yx}) and impedance phase (Φ_{yx}), identified as the E-perpendicular-to-strike mode, along a west to east profile crossing the Milford graben and the Mineral Mountains of southwestern Utah (after Wannamaker et al., 1983). Contours of ρ_{yx} are in ohm·m, and those of Φ_{yx} are in degrees.
(XBL 856-10596)

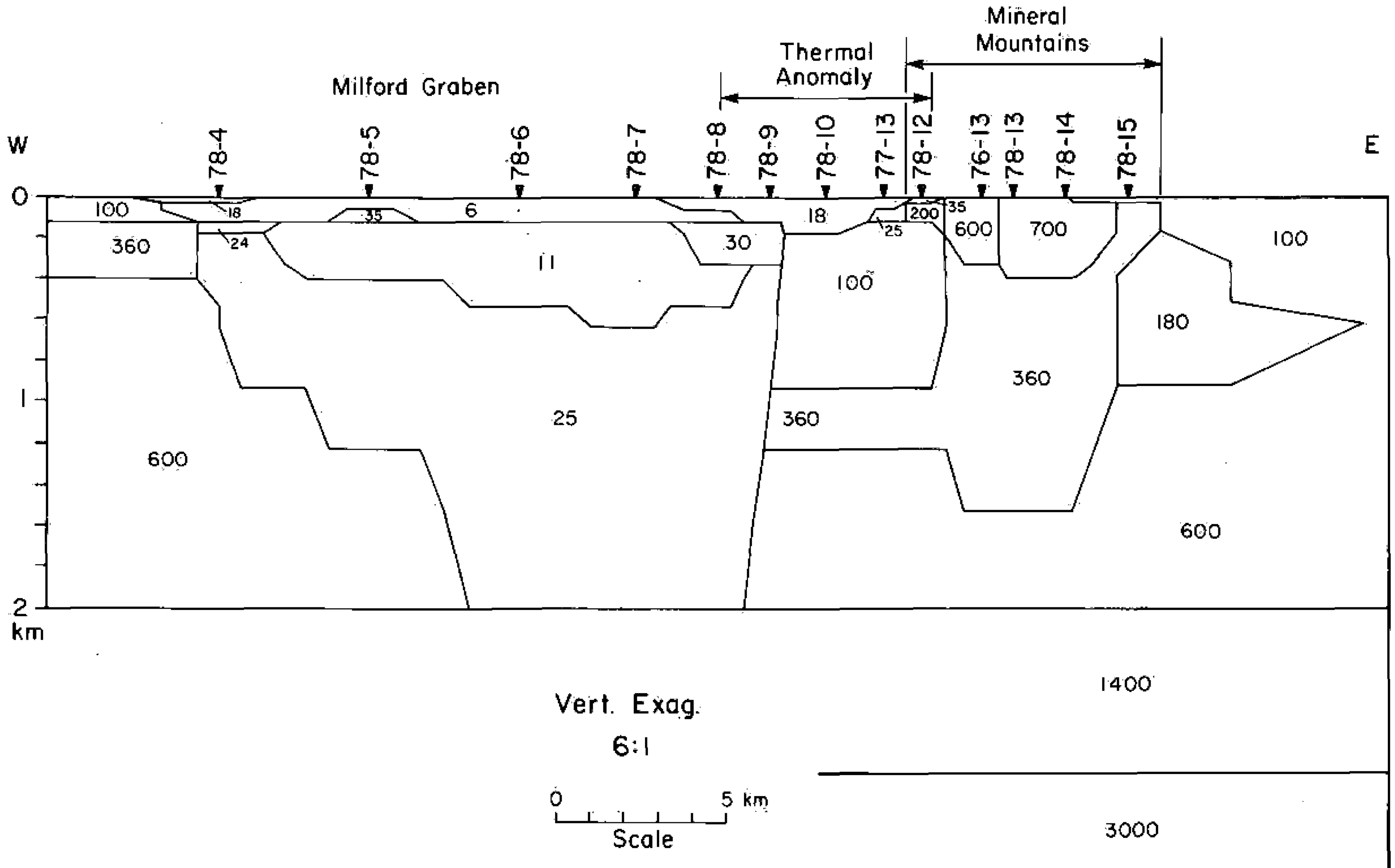


Figure 30. Resistivity cross section for the pseudosections in Figure 26, based on trial-and-error, two-dimensional, finite-element calculations (after Wannamaker et al., 1983). Resistivities are in ohm-meters, and the vertical exaggeration is 6:1. The location of the thermal anomaly associated with 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 bipole) 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$\cdot$ m^2) 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

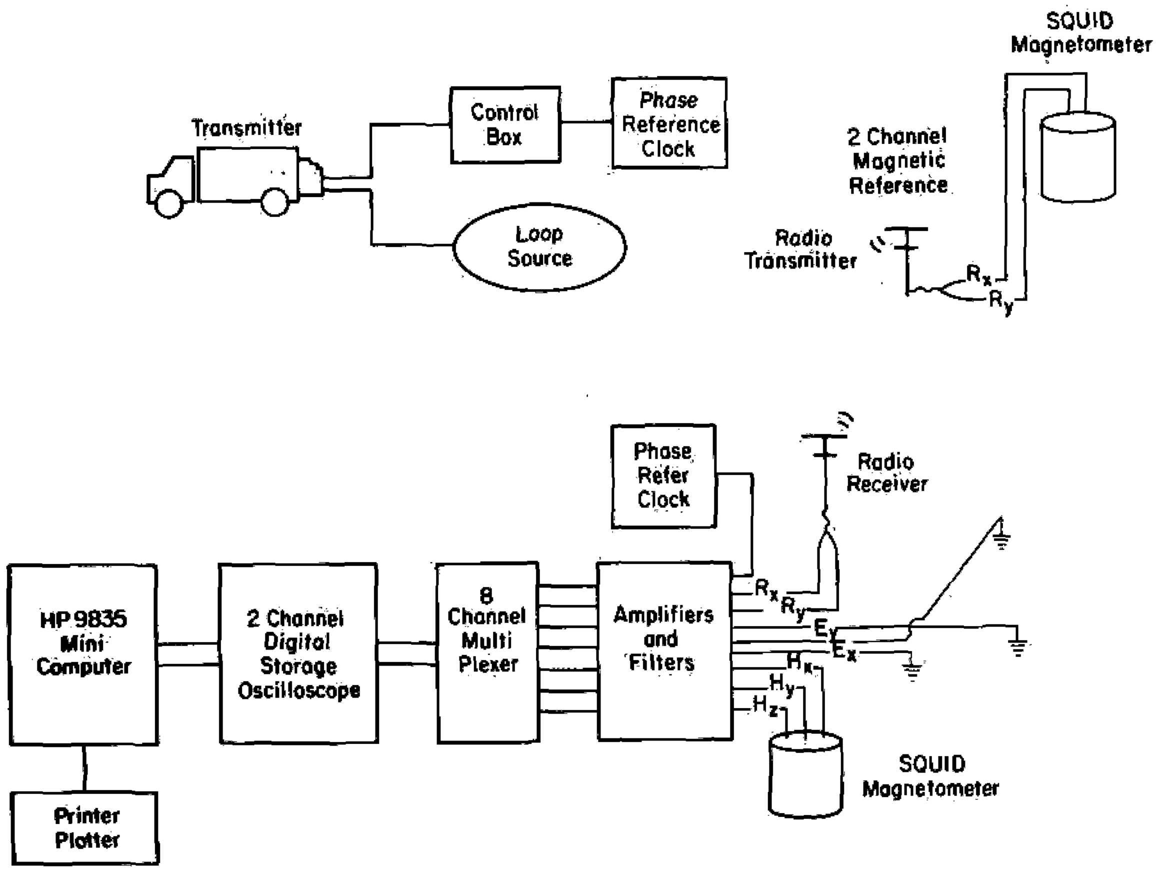


Figure 31. Block diagram of frequency-domain electromagnetic transmitter and in-field processing system developed at Lawrence Berkeley Laboratory and UC Berkeley (from Wilt et al., 1983).
(XBL 818-3383A)

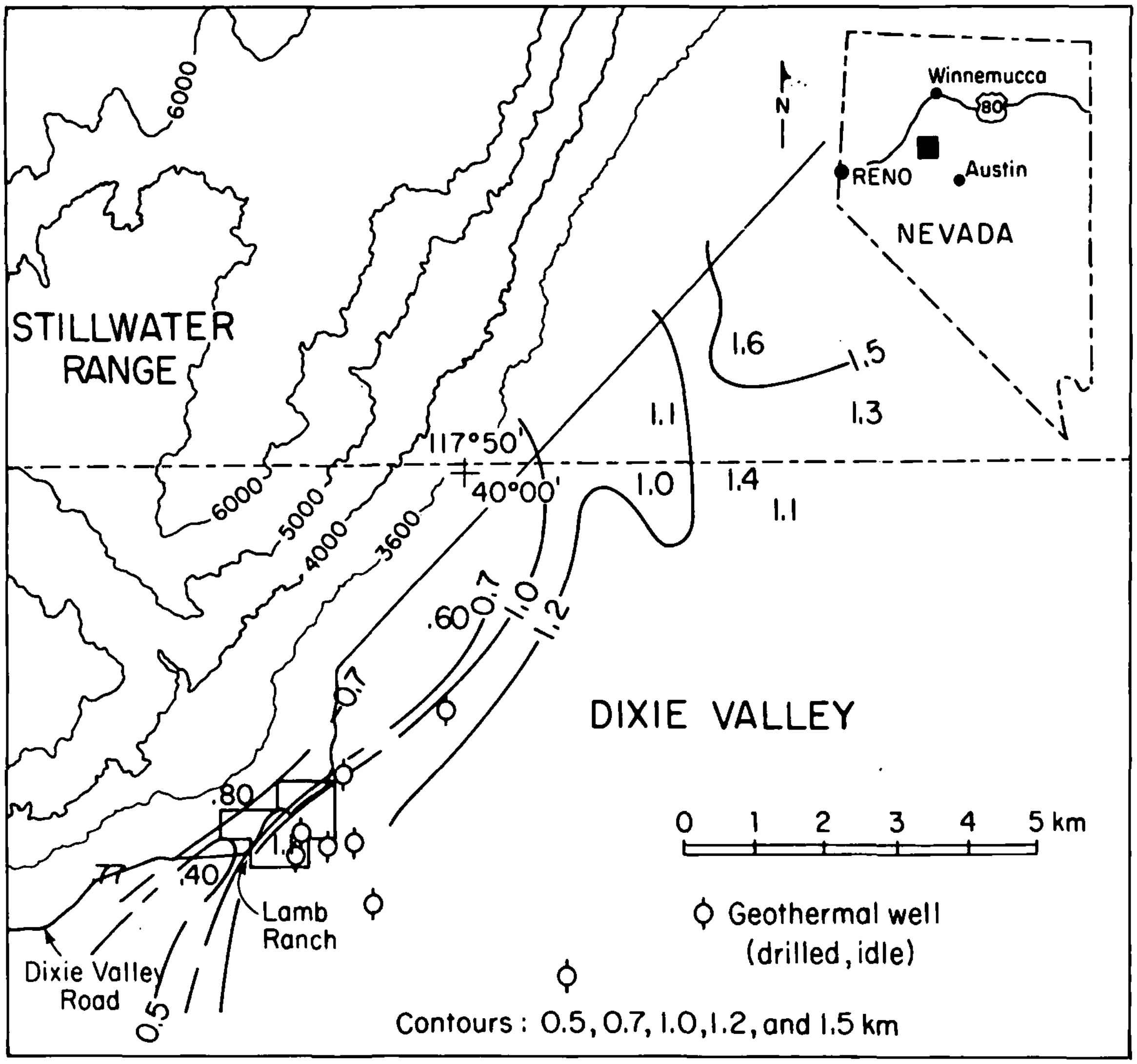


Figure 32. Depth (km) to resistive basement along the east flank of the Stillwater Range, Nevada, as discerned from EM soundings (from Wilt and Goldstein, 1983).
(XBL 8312-2450)

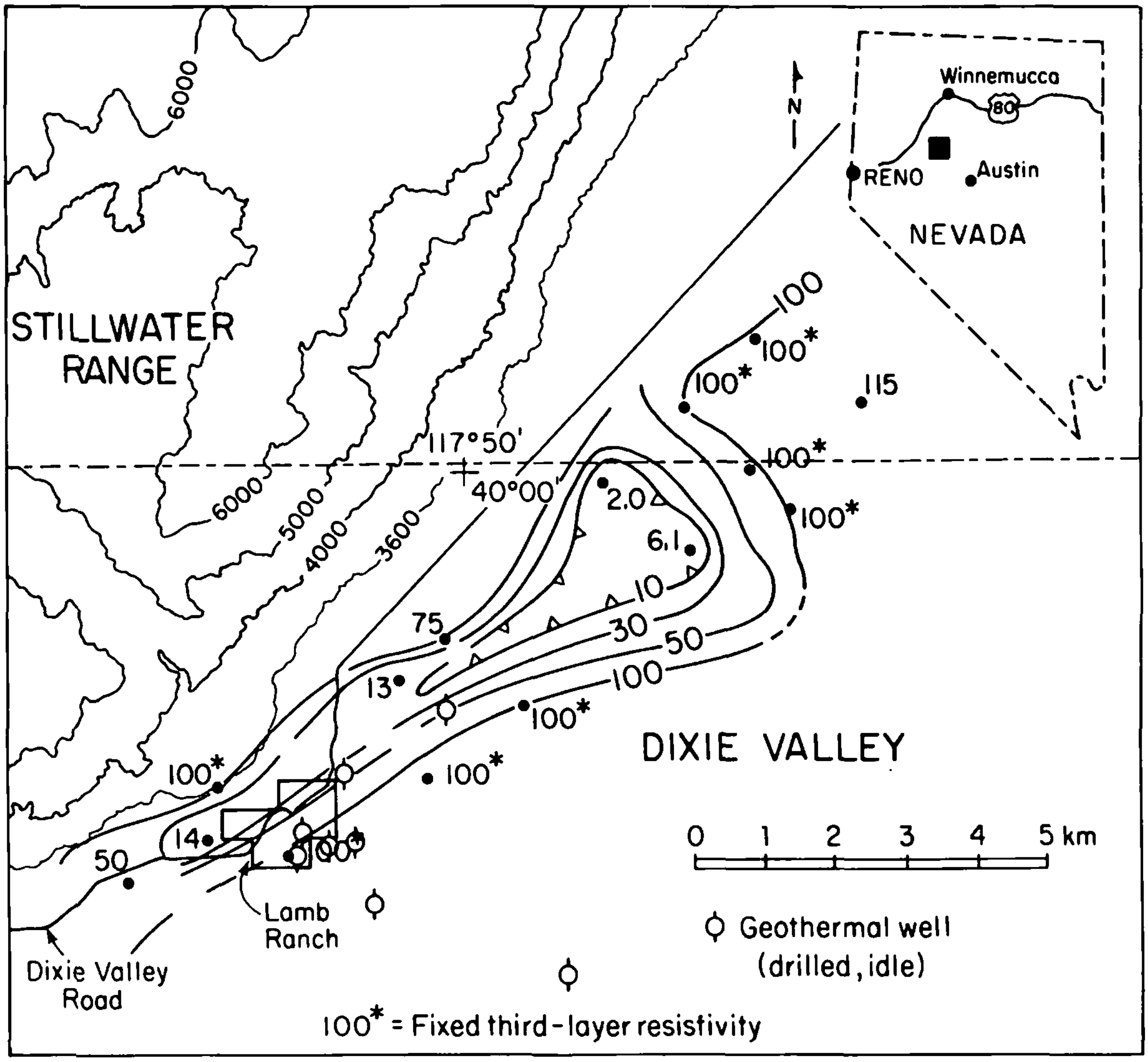


Figure 33. Results of a controlled-source electromagnetic survey in Dixie Valley, Nevada, showing a long northeastward-trending conductive zone found within the basal layer. Contours represent the basal layer resistivity in ohm·meters (from Wilt and Goldstein, 1983).
(XBL 8312-2449)

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 synonomous 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

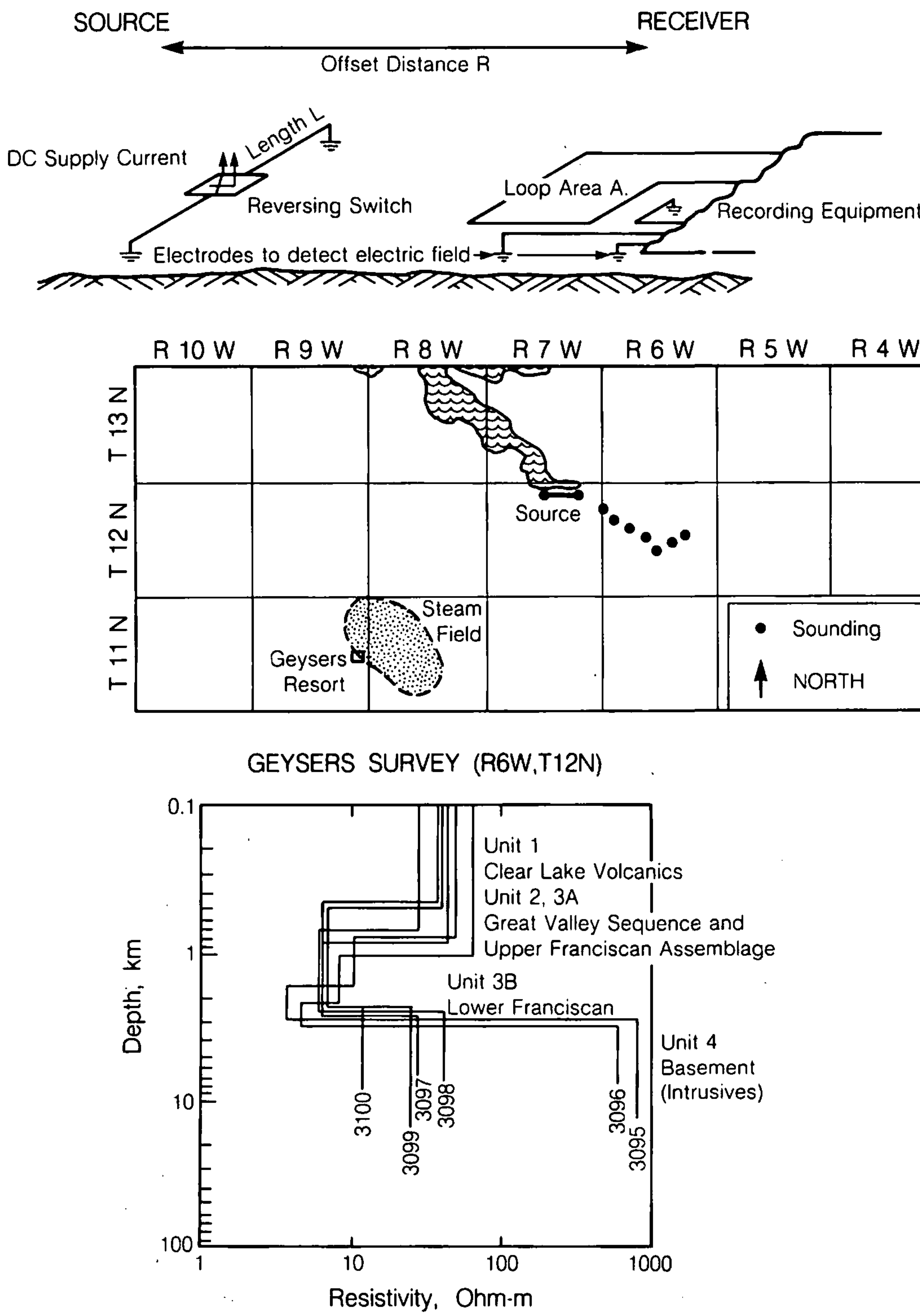


Figure 34. (a) Principal components of the Colorado School of Mines time-domain EM sounding system (from Keller et al., 1984). (b) Composite plot of the resistivity sections obtained by inversion of TDEM soundings in R6W T12N as part of The Geysers survey. The resistive units under the Franciscan assemblage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

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 synonomous 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 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 (Steeples and Iyer, 1976); Coso, California (Reasenberg 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

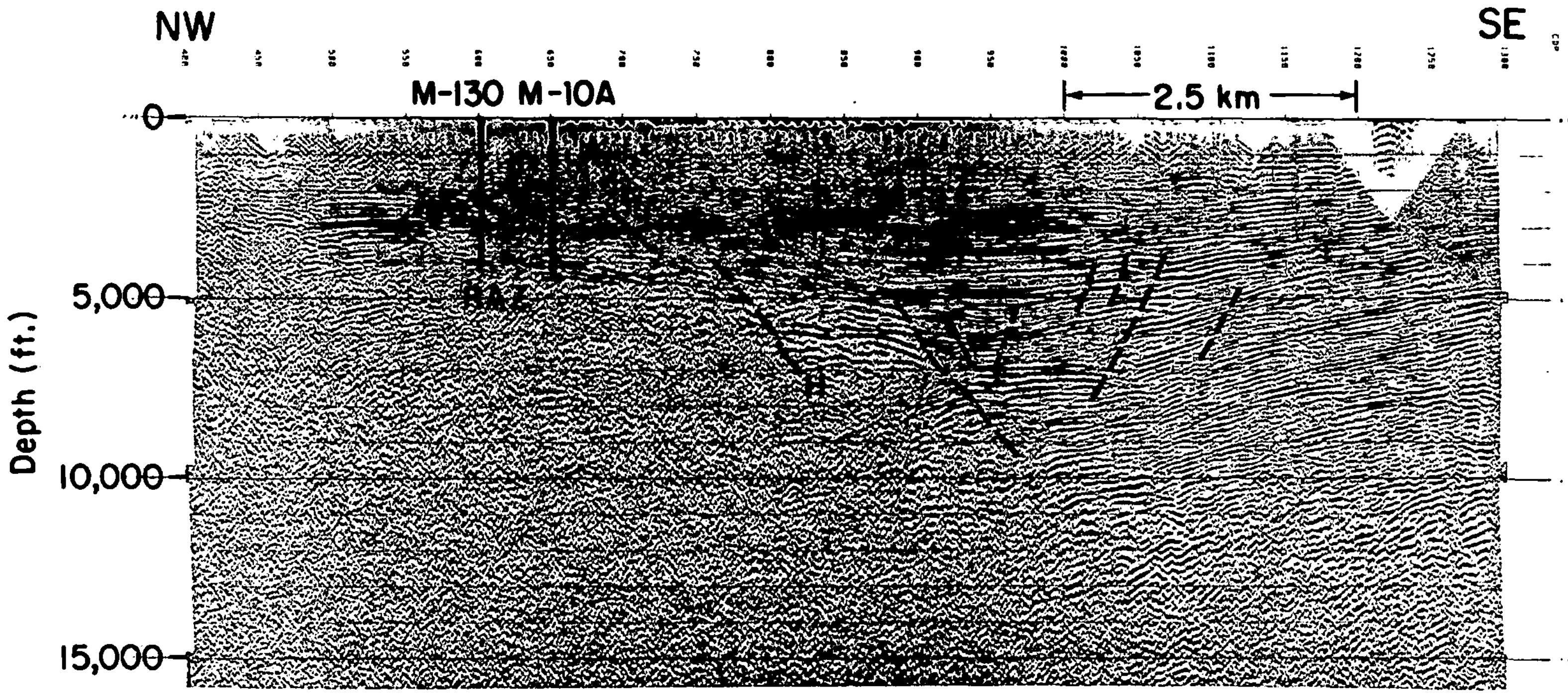


Figure 35. Processed seismic section over the Cerro Prieto field. Over 60,000 traces comprise the data set that was obtained with a Vibroseis® source and linear 14-48 Hz, 16-56 Hz, and 18-60 Hz sweeps. There were 256 channels per shot, an 80-ft receiver spacing, and a 4-ms sampling rate. A number of 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.

ACKNOWLEDGEMENTS

This work was supported by the Assistant Secretary for Conservation and Renewable Energy, Office of Renewable Energy Technologies, Division of Geothermal Technology of the U.S. Department of Energy under Contract No. DE-AC03-76SF00098.

The author would like to thank Alfred Truesdell, Marcelo Lippmann, and Hal Wollenberg, who reviewed parts of the manuscript and who made many helpful suggestions.

REFERENCES

Abrams, M.J., Brown, D., Lepley, L., and Sadowski, R., 1983, Remote sensing for porphyry copper deposits in southern Arizona: *Econ. Geol.*, v. 78, p. 591–604.

Albores, A., Reyes, A., Brune, J.N., Gonzalez, J., Garcilazo, L., and Suaraz, F., 1980, Seismicity studies in the region of the Cerro Prieto geothermal field: *Geothermics,* v. 9, n. 1/2, p. 65–77.

Alter, H.W., and Price, P.B., 1972, Radon detection using track registration material: U.S. Patent 3,665,194.

Anderson, W.L., Frischknecht, F.C., Raab, P.V., Bradley, J.A., Turnross, J., and Buckley, T.W., 1983, Inversion results of time-domain electromagnetic soundings near Medicine Lake, California, geothermal area: U.S. Geol. Survey *Open-File Rept. 82-223.*

Anuta, P.E., 1977, Computer-assisted analysis techniques for remote sensing data interpretations: *Geophysics,* v. 22, p. 468–481.

Archie, G.E., 1942, The electrical resistivity log as an aid in determining reservoir characteristics: *Trans. AIME,* v. 146, p. 54–67.

Armannsson, H., Gilason, G., and Hauksson, T., 1982, Magmatic gases in well fluids aid the mapping of the flow pattern in a geothermal system: *Geochim. et Cosmochim. Acta,* v. 46, p. 167–177.

Arnórsson, S., 1974, The composition of thermal fluids in Iceland and geological features related to the thermal activity, *in* Kristjansson, K., ed., Geodynamics of Iceland and the North Atlantic areas: D. Reidel, Dordrecht, Holland, p. 307–323.

_______, 1975, Application of the silica geothermometer in low temperature hydrothermal areas in Iceland: *Am. J. Sci.,* v. 275, p. 763-784.

Arnórsson, S., Björnsson, A., Gilason, G., and Gudmunsson, G., 1976, Systematic exploration of the Krívvik high-temperature area, Reykjanes Peninsula, Iceland: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 2, p. 853–864.

Bacon, C.R., 1982, Time-predictable bimodal volcanism in the Coso Range, California: *Geology,* v. 10, p. 66–69.

Bailey, R.A., Dalrymple, G.B., and Lanphere, M.A., 1976, Volcanism, structure and geochronology of Long Valley caldera, Mono County, California: *J. Geophys. Res.,* v. 81, p. 725–744.

Balch, A.H., and Lee, M.W., 1984, Vertical seismic profiling: Techniques, applications, and case histories: International Human Resources Dev. Corp., Boston, 488 p.

Bamford, R.W., 1978, Geochemistry of solid materials from two U.S. geothermal systems and its application to exploration: Univ. of Utah Res. Inst., Earth Sciences Laboratory, *Rept. ESL-6,* 196 p.

Bamford, R.W., and Christensen, O.D., 1979, Multielement geochemical exploration data for the Cove Fort-Sulphurdale Known Geothermal Resource Area, Beaver and Millard Counties, Utah: Univ. of Utah Res. Inst., Earth Sciences Laboratory, DOE/ET/28392-28, *Rept. ESL-19.*

Barker, C.E., 1979, Vitrinite reflectance geothermometry in the Cerro Prieto geothermal field, Baja California, Mexico: Inst. of Geophys. and Planetary Phys., Univ. of Calif., Riverside, *Rept. 79/10,* 105 p.

_______, 1983, Influence of time on metamorphism of sedimentary organic matter in liquid-dominated geothermal systems, western North America: *Geology,* v. 11, p. 384–388.

Barker, C.E., and Elders, W.A., 1979, Vitrinite reflectance geothermometry in the Cerro

Prieto geothermal field, Baja California, Mexico: *Geoth. Res. Council, Trans.*, v. 3, p. 27–30.

_______, 1981, Vitrinite reflectance geothermometry and apparent heating duration in the Cerro Prieto geothermal field: *Geothermics*, v. 10, n. 3/4, p. 207–224.

Battacharya, P.K., and Patra, H.P., 1968, Direct Current Geoelectric Sounding, Principles, and Interpretation: Elsevier, New York, 135 p.

Berdichevskiy, M.N., 1960, Electrical surveying by means of telluric currents: Gostoptekhizdat, Moscow. (English translation by Keller, G.V., 1965, Quarterly Colo. School of Mines, v. 60, n. 1, 216 p.).

Berdichevsky, M.N., and Dimitriev, V.I., 1976, Basic principles of interpretation of magnetotelluric sounding curves, *in* Adam, A., ed., Geoelectric and Geothermal Studies: Academy Kiado, Budapest, p. 165–221.

Bergquist, L.E., 1979, Helium, an exploration tool for geothermal sites: *Geoth. Res. Council, Trans.*, v. 3, p. 59–60.

Beyer, H., Dey, A., Liaw, A., Majer, E., McEvilly, T.V., Morrison, H.F., and Wollenberg, H., 1976, Geological and geophysical studies in Grass Valley, Nevada: Preliminary Open-File Report: Lawrence Berkeley Laboratory, *Rept. LBL-5262*, 144 p.

Beyer, J.H., 1977a, Telluric and dc resistivity techniques applied to the geophysical investigation of Basin and Range geothermal systems, Part I: The E-field ratio telluric method: Ph.D. Thesis, Univ. of Calif., Berkeley, and Lawrence Berkeley Laboratory, *Rept. LBL-6325*, 1/3, 135 p.

_______, 1977b, Telluric and dc resistivity techniques applied to the geophysical investigation of Basin and Range geothermal systems, Part II: A numerical model study of the dipole-dipole and Schlumberger resistivity methods: Lawrence Berkeley Laboratory, *Rept. LBL-6325*, 2/3.

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

Bibby, H.M., 1977, The apparent resistivity tensor: *Geophysics*, v. 42, p. 1258–1261.

Biehler, S., 1971, Gravity studies in the Imperial Valley, *in* Rex, R.W., et al., eds., Cooperative Geological-Geophysical-Geochemical Investigations of Geothermal Resources in the Imperial Valley Area of California: Univ. of Calif., Riverside, p. 29–41.

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

Birch, F., Roy, R.F., and Decker, E.R., 1968, Heat flow and thermal history in New England and New York, *in* Zen, E-an, White, W.S., Hadley, J.B., and Thompson, J.B., Jr., eds., Studies of Appalachian Geology: New York, Northern and Maritime: Interscience, p. 437–451.

Björnsson, A., Johnson, G., Sigurdsson, S., Thobergsson, G., and Tryggvason, E., 1979, Rifting of the plate boundary in north Iceland 1975-1978: *J. Geophys. Res.*, v. 84, n. B6, p. 3029–3038.

Blackwell, D.D., and Steele, J.L., 1983, A summary of heat flow studies in the Cascade Range: *Geoth. Res. Council, Trans.*, v. 7, p. 233–236.

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

Blakeslee, S., 1984, Seismic discrimination of a geothermal field: Cerro Prieto: *Geoth. Res. Council, Trans.*, v. 8, p. 183–188.

Boler, F.M., 1978, Aeromagnetic measurements, magnetic source depths, and the Curie point isotherm in the Vale-Owyhee, Oregon: M.S. Thesis, Oregon State Univ., Corvallis.

Bowman, H.R., Hebert, A.J., Wollenberg, H.A., and Asaro, F., 1976, Trace, minor, and major elements in geothermal waters and associated rock formations (North-central Nevada): *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975*, v. 1, p. 699–702.

Brophy, P., 1984, Structural analysis of pre-Cenozoic rocks, Coso geothermal area, California: *Geoth. Res. Council, Trans.*, v. 8, p. 409–415.

Browne, P.R.L., 1978, Hydrothermal alteration in active geothermal fields: *Ann. Rev. Earth Planet. Sci.*, v. 6, p. 229–250.

Browne, P.R.L., and Ellis, A.J., 1970, the Ohaki-Broadlands hydrothermal area, New Zealand: Mineralogy and related geochemistry: *Am. J. Sci.*, v. 269, p. 97–131.

Buddington, A.F., and Lindsley, D.H., 1964, Iron-titanium oxide minerals and synthetic equivalents: *J. Petrology*, v. 6, p. 310–357.

Bussain, A.E., 1983, Electrical conductance in a porous medium: *Geophysics*, v. 48, p. 1258–1268.

Byerly, P.E., and Stolt, R.H., 1977, An attempt to define the Curie Point isotherm in northern and central Arizona: *Geophysics*, v. 42, p. 1394–1400.

Cagniard, L., 1953, Basic theory of the magnetotelluric method of geophysical prospecting: *Geophysics*, v. 18, p. 605–635.

Cantwell, T., 1960, Detection and analysis of low frequency electromagnetic signals: Ph.D. Thesis, Mass. Inst. of Tech.

Capuano, R.M., and Bamford, R.W., 1978, Initial investigation of soil mercury geochemistry as an aid to drill site selection in geothermal systems: Univ. of Utah Res. Inst., Earth Sciences Laboratory, *Rept. ESL-13*, 32 p.

Carlston, K., 1982, Electric field ratio telluric survey of the Roosevelt Hot Springs, Utah: Univ. of Utah, Dept. of Geology and Geophysics, *DOE/ET/28392-26*, 41 p.

Carmichael, I.S.E., Turner, F.J., and Verhoogen, J., 1974, Igneous Petrology: McGraw-Hill Company, New York, 739 p.

Cathlineau, M., Oliver, R., Nieva, D., and Garfias, A., 1985, Mineralogy and distribution of hydrothermal mineral zones in Los Azufres (Mexico) geothermal field: *Geothermics*, v. 14, n. 1, p. 49–57.

Cavarretta, G., Gianelli, G., and Puxeddu, M., 1982, Formation of authigenic minerals and their use as indicators of the physiochemical parameters of the fluid in the Lardarello-Travale geothermal field: *Econ. Geol.*, v. 77, n. 5, p. 1071–1084.

Chapman, R.H., 1975, Geophysical study of the Clear Lake region, California: California Div. Mines and Geol., *Spec. Rept. 116*, 23 p.

Chu, J.J., Sill, W.R., and Ward, S.H., 1979, Induced polarization measurements at Roosevelt Hot Springs thermal area, Utah: Univ. of Utah, Dept. of Geology and Geophysics, *IDO/78-1701.a.2.4.1*, 34 p.

Ciancanelli, E.V., 1983, Geology of the Medicine Lake Volcano, California: *Geoth. Res. Council, Trans.*, v. 7, p. 135–140.

Cifali, G., and Whiteley, J., 1981, Self potential survey of the Woodlawn Orebody, *in* Whiteley, R.J., ed., Geophysical Case Study of the Woodlawn Orebody, New South Wales, Australia: Pergamon Press, Oxford, England, p. 321–332.

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

Cole, D.R., 1983, Time estimates for oxygen exchange during mineral-fluid interaction in hydrothermal systems: *Geoth. Res. Council, Trans.*, v. 7, p. 283–287.

Cole, K.S., and Cole, R.H., 1941, Dispersion and absorption in dielectrics: *J. Chem. Phys.*, v. 9, p. 341.

Combs, J., 1971, Heat flow and geothermal resource estimates for the Imperial Valley, *in* Rex, R.W., et. al., eds., Cooperative Geological-Geophysical-Geochemical Investigations of Geothermal Resources in the Imperial Valley Area of California: Univ. of Calif., Riverside, p. 5–28.

Combs, J., 1980, Heat flow in the Coso geothermal area, Inyo County, California: *J. Geophys. Res.*, v. 85, n. B5, p. 2411–2424.

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

Combs, J., and Rotstein, Y., 1976, Microearthquake studies at the Coso geothermal area, China Lake, California: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975*, v. 2, p. 909–916.

Combs, J., Quiett, S., and Axtell, L.H., 1977, Thermal conductivity measurement techniques for geothermal exploration heat flow surveys: *Geoth. Res. Council, Trans.*, v. 1, p. 49–51.

Compagnie Général de Géophysique, 1963, Master curves for electrical sounding, 2nd revised edition: European Association of Exploration Geophysicists, The Hague, The Netherlands.

Connard, G., Couch, R., and Gemperle, M., 1983, Analysis of aeromagnetic measurements from the Cascade Range in central Oregon: *Geophysics*, v. 48, p. 376–390.

Copp, J.F., 1981, Soil mercury geochemistry in the Coso volcanic field, Inyo County, California: M.S. Thesis, Arizona State Univ.

Cordell, L., and Henderson, R.G., 1968, Iterative three-dimensional solution of gravity data using a digital computer: *Geophysics*, v. 33, p. 596–601.

Corwin, R.F., 1976, Self-potential exploration for geothermal reservoirs: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975*, v. 2, p. 937–945.

Corwin, R.F., and Hoover, D.B., 1979, The self-potential method in geothermal exploration: *Geophysics*, v. 44, p. 226–235.

Corwin, R.F., Morrison, H.F., Diaz C., S., and Rodriguez B., J., 1980, Self potential studies at the Cerro Prieto geothermal field: *Geothermics*, v. 9, n. 1/2, p. 39–48.

Corwin, R.F., DeMoully, G.T., Harding, R.S., Jr., and Morrison, H.F., 1981, Interpretation of self-potential sounding results from the East Mesa geothermal field, California: *J. Geophys. Res.*, v. 86, p. 1841–1848.

Couch, R., and Gemperle, M., 1979, Gravity measurements in the area of Mount Hood, Oregon, *in* Geothermal Resource Assessment of Mount Hood, Final Report: State of Oregon, Dept. of Geology and Mineral Industries, *RLO-1040*, p. 137–189h.

Cox, M.E., 1980, Ground radon survey of a geothermal area in Hawaii: *Geophys. Res. Lett.*, v. 7, p. 283–286.

Cox, M.E., Cuff, K.E., and Thomas, D.M., 1980, Variations of ground radon concentrations with activity at Kilauea Volcano, Hawaii: *Nature*, v. 288, p. 74–76.

Craig, H., 1963, The isotopic geochemistry of water and carbon in geothermal areas, *in* Tongiorgi, E., ed., Nuclear Geology on Geothermal Areas, Spoleto, 1963, Pisa: Consiglio Nazionale delle Ricerche Laboratorio di Geologia Nucleare, p. 17–54.

Craig, H., and Lupton, J.E., 1976, Primordial neon, helium and hydrogen in oceanic basalts: *Earth Planet. Sci. Lett.*, v. 31, p. 369–385.

Craig, H., Lupton, J.E., Welhan, J.A., and Porela, R., 1978, Helium isotope ratios in Yellowstone and Lassen Park volcanic gases: *Geophys. Res. Lett.*, v. 5, p. 897–900.

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

_______, 1984, Effective anisotropic propagation through a cracked solid, *in* Crampin, S., Hipkin, R.G., and Chesnokov, E.M., eds., Proc. of the First Internat. Workshop on Seismic Anisotropy: *Geophys. J. Roy. Astron. Soc.*, v. 76, p. 135–145.

_______, 1985, Evaluation of anisotropy by shear wave splitting: *Geophysics,* v. 50, no. 1, p. 142–152.

Dahlberg, R.S., Jr., 1945, An investigation of natural earth currents: *Geophysics,* v. 10, p. 494–506.

D'Amore, F., and Nuti, S., 1977, Notes on the chemistry of geothermal gases: *Geothermics,* v. 6, n. 1/2, p. 39–46.

D'Amore, F., and Panichi, C., 1985, Geochemistry in geothermal exploration: *Int. J. Energy Res., Spec. Issue on Geothermal Energy,* v. 9, n. 3, p. 277–298.

de Boer, J., 1979, Paleomagnetism of the Quaternary Cerro Prieto, Crater Elegante, and Salton Buttes volcanic domes in the northern part of the Gulf of California rhombochasm: *Proc., 2nd Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, Mexicali, October 1979,* Comisión Federal de Electricidad, p. 91–98.

de la Peña L., A., Puente C., J., and Díaz C., E., 1979, Geologic model of the Cerro Prieto geothermal field: *Proc., 2nd Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, Mexicali, October 1979,* Comisión Federal de Electricidad, p. 29.

del Grande, N.K., 1982, Airborne temperature survey maps of heat flow anomalies for exploration geology: *Geoth. Res. Council Bull.,* v. 14, March 1985, p. 3–13.

Denlinger, R.P., and Kovach, R.L., 1977, Seismic-reflection investigations in a geothermal area: *Geoth. Res. Council, Trans.,* v. 1, p. 77–78.

_______, 1981, Seismic-reflection investigations at Castle Rock Springs in The Geysers geothermal area, *in* Research in the Geysers-Clear Lake geothermal area: U.S. Geol. Survey *Prof. Paper 1141,* p. 117–128.

Dey, A., and Morrison, H.F., 1977, An analysis of the bipole-dipole method of resistivity surveying: *Geothermics,* v. 6, n. 1/2, p. 47–82.

Doicin, D., 1976, Quadripole-quadripole arrays for direct resistivity measurements–Model studies: *Geophysics,* v. 41, p. 79–95.

Douze, E.J., and Laster, S.J., 1979, Seismic noise studies at the Roosevelt Hot Springs, Utah, geothermal area: *Geophysics,* v. 44, p. 1570–1583.

Douze, E.J., and Sorrells, G.G., 1972, Geothermal groundnoise surveys: *Geophysics,* v. 37, p. 813–824.

Duffield, W.A., and Bacon, C.R., 1980, Geologic map of the Coso volcanic field and adjacent areas, Inyo County, California: Misc. Geol. Invest. *Map I-2000,* U.S. Geol. Survey.

Duffield, W.A., Bacon, C.A., and Dalrymple, G.B., 1980, Late Cenozoic volcanism, geochronology, and structure of the Coso Range, Inyo County, California: *J. Geophys. Res.,* v. 85, no. B5, p. 2381–2404.

Eichelberger, J.C., and Gooley, R., 1977, Evolution of silicic magma chambers and their relationship to basaltic volcanism: Am. Geophys. Union *Monograph 20,* p. 57–77.

Elders, W.A., Rex, R.W., Meidav, T., Robinson, P.T., and Biehler, S., 1972, Crustal spreading in southern California: *Science,* v. 178, p. 15–24.

Elders, W.A., Hoagland, J.R., and Williams, A.E., 1979, Distribution of hydrothermal mineral zones in the Cerro Prieto geothermal field of Baja California, Mexico: *Proc., 2nd Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, Mexicali, October 1979,* Comisión Federal de Electricidad, p.57–69.

Ellis, A.J., and Mahon, W.A.J., 1964, Natural hydrothermal systems and experimental hot-water/rock interactions: *Geochim. et Cosmochim. Acta,* v. 28, p. 1323–1357.

_______, 1977, Chemistry and geothermal systems: Academic Press, New York, 392 p.

Evison, F.F., Robinson, R., and Arabasz, W.J., 1976, Microearthquakes, geothermal activity, and structure, Central North Island, New Zealand: *N.Z. J. Geol. and Geophys.*, v. 19, p. 625–637.

Fairbank, B.D., Openshaw, R.E., Souther, J.G., and Stauder, J.J., 1981, Meager Creek geothermal Project–An exploration case history: *Geoth. Res. Council Bull.*, v. 10, n. 6, p. 3–7.

Fink, J.H., and Pollard, D.D., 1983, Ground cracks as indicators of geothermal potential (abstr.): *Eos, Trans. Am. Geophys. Union,* v. 64, n. 45, p. 898.

Finn, C., and Williams, D.L., 1982, Gravity evidence for a shallow intrusion under Medicine Lake Volcano, California: *Geology,* v. 10, p. 503–507.

Finn, C., and Williams, D., 1983, Gravity studies in the Cascade Range: *Geoth. Res. Council Trans.*, v. 7, p. 247–251.

Fitterman, D.V., 1979, Calculation of self-potential anomalies near vertical contacts: *Geophysics,* v. 44, p. 195–205.

Fitterman, D.V., and Corwin, R.F., 1982, Inversion of self-potential data from the Cerro Prieto geothermal field, Mexico: *Geophysics,* v. 47, p. 938–945.

Fleischer, R.L., Price, P.B., and Walker, R.M., 1975, Nuclear Tracks in Solids: Berkeley, Univ. of Calif. Press.

Flövenz, O.G., and Georgsson, L.S., 1982, Prospecting for near vertical aquifers in low temperature geothermal areas in Iceland: *Geoth. Res. Council, Trans.*, v. 6, p. 19–22.

Fournier, R.O., 1973, Silica in geothermal waters: Laboratory and field investigations: *Proc., International Symposium on Hydrologeochemistry and Biogeochemistry, Japan, 1970,* Vol. 1, Hydrogeochemistry, Washington, D.C., p. 122–139.

_______, 1977, Chemical geothermometers and mixing models for geothermal systems: *Geothermics,* v. 5, p. 41–50.

_______, 1981, Application of water geochemistry to geothermal exploration and reservoir engineering, *in* Rybach, L., and Muffler, L.J.P., eds., Principles and Case Histories: John Wiley & Sons, Ltd, p. 109–143.

_______, 1982, Geochemistry in geothermal exploration and production: *Proc., Pacific Geothermal Conference, 1982, Univ. of Auckland Geothermal Institute, Auckland, New Zealand,* Part 1, p. 209–215.

_______, 1983, Active hydrothermal systems as analogues of fossil systems: Geoth. Res. Council, *Spec. Rept. 13,* p. 263–284.

Fournier, R.O., and Potter, R.W., 1978, A magnesium correction for the Na-K-Ca chemical geothermometer: U.S. Geol. Survey *Open-File Rept. 78-486,* 24 p.

Fournier, R.O., and Potter, R.W., II, 1982, A revised and expanded silica (quartz) geothermometer: *Geoth. Res. Council Bull.*, v. 11, n. 10, p. 3–9.

Fournier, R.O., and Truesdell, A.H., 1973, An empirical Na-K-Ca geothermometer for natural waters: *Geochim. et Cosmochim. Acta,* v. 37, p. 1255–1275.

_______, 1974, Geochemical indicators of subsurface temperature–Part 2, Estimation of temperature and fraction of hot water mixed with cold water: *U.S. Geol. Survey J. Res.*, v. 2, n. 3, p. 263–270.

Fox, R.C., 1978, Low-altitude aeromagnetic survey of a portion of the Coso Hot Springs KGRA, Inyo County, California: Univ. of Utah Res. Inst., Earth Sciences Laboratory, *Rept. IDO/77.S.7,* 19 p.

Fox, R.C., Hohmann, G.W., Killpack, T.J., and Rijo, L., 1980, Topographic effects in resistivity and induced polarization surveys: *Geophysics,* v. 45, p. 75–93.

Fridleifsson, I.B., 1979, Applied volcanology in geothermal exploration in Iceland, *in* Rybach, L., and Stegena, L., eds., Geothermics and Geothermal Energy: Birkhäuser Verlag, Basel, p. 242–252.

Gal'perin, E.I., 1974, Vertical seismic profiling: Soc. Expl. Geophys., Tulsa, *Spec. Publ. No. 12,* 270 p.

Gamble, T.D., Goubau, W.M., and Clarke, J., 1979, Magnetotellurics with a remote reference: *Geophysics,* v. 44, n. 1, p. 53–68.

Gamble, T.D., Goubau, W.M., Goldstein, N.E., Miracky, R., Starke, M., and Clarke, J., 1981, Magnetotelluric studies at Cerro Prieto: *Geothermics,* v. 10, n. 3/4, p. 169–182.

Gianelli, G., Puxeddo, M., and Squarci, P., 1978, Structural setting of the Larderello-Travale geothermal region: Societá Geologica Italiana, 69th Congress, Perugia.

Goetz, A.F.H., Rock, B.N., and Rowan, L.C., 1983, Remote sensing for exploration: An overview: *Econ. Geol.,* v. 78, p. 573–590.

Gold, T., and Soter, S., 1980, The deep-earth-gas hypothesis: *Sci. Am.,* v. 242, p. 154–161.

_______, 1982, Abiogenic methane and the origin of petroleum: *Energy Explor. Exploit.,* v. 1, p. 89–104.

Goldstein, N.E., and Carle, S., 1986, Faults and gravity anomalies over the East Mesa hydrothermal-geothermal system: Lawrence Berkeley Laboratory, *Rept. LBL-21459,* 6 p. (also accepted for publication in *Geoth. Res. Council. Trans.,* v. 10).

Goldstein, N.E., and Paulsson, B., 1978, Interpretation of gravity surveys in Grass and Buena Vista Valleys, Nevada: *Geothermics,* v. 7, p. 29–50.

Goldstein, N.E., Beyer, H., Corwin, R., di Somma, D.E., Majer, E., McEvilly, T.V., Morrison, H.F., Wollenberg, H.A., and Grannell, R., 1976, Geosciences Studies in Buena Vista Valley, Nevada: Lawrence Berkeley Laboratory, *Rept. LBL-5913,* 41 p.

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

Goldstein, N.E., Wilt, M.J., and Corrigan, D.J., 1984, Analysis of the Nuevo León magnetic anomaly and its possible relation to the Cerro Prieto magmatic-hydrothermal system: *Geothermics,* v. 13, n. 1/2, p. 3–11.

Goss, R., and Combs, J., 1975, Thermal conductivity measurement and prediction from geophysical well log parameters with borehole application: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 3, p. 1014–1027.

Goupillaud, P.L., and Cherry, J.T., 1977, Utilization of seismic exploration technology for high resolution mapping of a geothermal reservoir: S-Cubed, *Final Rept. SSR-R-77-3176,* 55 p.

Goyal, K.P., and Kassoy, D.R., 1981, A plausible two-dimensional vertical model of the East Mesa geothermal field, California: *J. Geophys. Res.,* v. 86, p. 10719–10733.

Graeber, E.J., Modeski, P.J., and Gerlach, T.M., 1979, Composition of gases collected during the 1977 east rift eruption, Kilauea, Hawaii: *J. Volcanol. Geoth. Res.,* v. 5, p. 344–377.

Grannell, R.B., Whitcomb, J.H., Aronstam, P.S., and Clover, R.C., 1981, Assessment of precise gravity measurements for monitoring the response of a geothermal reservoir to exploitation: Lawrence Berkeley Laboratory, *Rept. LBL-12910,* 105 p.

Griscom, A., and Muffler, L.J.P., 1971, Salton Sea aeromagnetic map: USGS Geophysical Map, *Map GP-754,* Washington, D.C.

Gunter, B.D., 1973, Aqueous phase-gaseous phase material balance studies of argon and nitrogen in hydrothermal features at Yellowstone National Park: *Geochim. et Cosmochim. Acta,* v. 37, p. 495–513.

Gupta, H.K., Ward, R.W., and Lin, T.L., 1982, Seismic wave velocity investigation at the

Geysers-Clear Lake geothermal field, California: *Geophysics,* v. 47, p. 819–824.
Halfman, S.E., Lippmann, M.J., Zelwer, R., and Howard, J.H., 1984, A geological interpretation of the geothermal fluid movement in the Cerro Prieto field, Baja California, Mexico: *Am. Assoc. Petrol. Geol. Bull.,* v. 68, n. 1, p. 18–30.
Halliday, M.E., and Cook, K.L., 1978, Gravity and ground magnetic surveys in the Monroe and Joseph KGRA's and surrounding region, South Central Utah: Univ. of Utah, Dept. of Geology and Geophysics, *DOE/EY/76/S/07/1601,* 164 p.
Hamilton, R.M., and Muffler, L.J.P., 1972, Microearthquakes at The Geysers geothermal area, California: *J. Geophys. Res.,* v. 77, p. 2081–2086.
Harthill, N., 1978, A quadripole resistivity survey of the Imperial Valley, California: *Geophysics,* v. 43, p. 1485–1500.
Haught, J.R., Wilt, M.J., and Goldstein, N.E., 1981, Deep induction sounding for geothermal exploration from an arbitrarily oriented magnetic dipole (abstr.): *Geophysics,* v. 46, p. 458.
Hayakawa, M., 1970, The study of underground structures and geophysical state in geothermal areas by seismic exploration: *Geothermics,* v. 2, pt. 1, p. 347–357.
Hearn, B.C., Jr., Donnelly-Nolan, J.M., and Goff, F.E., 1981, The Clear Lake volcanics: Tectonic setting and magma sources, *in* McLaughlin, R.J., and Donnelly-Nolan, J.M., eds., Research in the Geysers-Clear Lake geothermal area, Northern California: U.S. Geol. Survey *Prof. Paper 1141,* p. 25–45.
Hedenquist, J.W., and Henley, R.W., 1985, Hydrothermal eruptions in the Waiatapu geothermal system, New Zealand: Their origin, associated breccias, and relation to precious metal mineralization: *Econ. Geol.,* v. 80, p. 1640–1668.
Heiken, G., 1978, Plinian-type eruptions in the Medicine Lake Highland, California, and the nature of the underlying magma: *J. Volc. Geotherm. Res.,* v. 4, p. 375–402.
Henley, R.W., and Ellis, A.J., 1983, geothermal systems ancient and modern, a geological review: *Earth Sci. Rev.,* v. 19, p. 1–50.
Henley, R.W., Truesdell, A.H., and Barton, P.B., Jr., 1984, Fluid-mineral equilibria in geothermal systems: *Rev. Econ. Geol.*, v. 1, 267 p.
Henry, S.G., and Pollock, H.N., 1982, Terrane correction for terrestrial heat flow: *Technical Program Abstracts, Soc. Expl. Geophys., 52nd Annual International Meeting and Exposition, Dallas,* p. 493–494.
Hermance, J.F., 1982, The asymptotic response of three-dimensional basin off-sets to magnetotelluric fields at long periods: The effects of current channeling: *Geophysics,* v. 47, p. 1562–1573.
Hewitt, E.J., 1963, The essential nutrient elements: Requirements and interactions in plants, *in* Stewart, F.C., ed., Plant Physiology–A Treatise: New York, Academic Press, v. 3, p. 137–360.
Higgins, M.W., 1973, Petrology of Newberry Volcano, central Oregon: *Geol. Soc. Am. Bull.,* v. 84, p. 455–488.
Hildreth, W., 1981, Gradients in silicic magma chambers: Implications for lithospheric magmatism: *J. Geophys. Res.,* v. 86, n. B11, p. 10153–10192.
Hinkle, M.E., and Kilburn, J.E., 1980, Survey of helium in soil gases of Long Valley, California: U.S. Geol. Survey *Open-File Rept. 80-612.*
Hochstein, M.P., 1976, Geophysical exploration of the Kawah Kamojang geothermal field, West Java: *Proc., 2nd Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 2, p. 1049–1058.
Hochstein, M.P., and Hunt, T.M., 1970, Seismic gravity and magnetic studies, Broadlands geothermal field, New Zealand: *Proc., U.N. Symposium on the Development and*

Utilization of Geothermal Resources, Pisa, (Geothermics, Spec. Iss. 2), v. 2, pt. 1, p. 333.

Holland, H.D., 1967, Gangue minerals in hydrothermal deposits, *in* Barnes, H.L., ed., Geochemistry of Hydrothermal Ore Deposits: Holt, Rinehart and Winston, New York, p. 382–436.

Hoover, D.B., Frischknecht, F.C., and Tippens, C.L., 1976, Audiomagnetotelluric sounding as a reconnaissance exploration technique in Long Valley, California: *J. Geophys. Res.*, v. 81, p. 801–809.

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

Hose, R.K., and Taylor, B.E., 1974, Geothermal systems of northern Nevada: U.S. Geol. Survey *Open-File Rept. 74-271.*

Hulen, J.B., 1983, Structural control of the Baltazor Hot Springs geothermal system, Humboldt County, Nevada: *Geoth. Res. Council, Trans.*, v. 7, p. 157–162.

Hulen, J.B., and Nielson, D.L., 1982, Stratigraphic permeability in the Baca geothermal system, Redondo Creek area, Valles Caldera, New Mexico: *Geoth. Res. Council, Trans.*, v. 6, p. 27–30.

Hulen, J.B., and Nielson, D.L., 1986, Hydrothermal alteration in the Baca system, Redondo dome, Valles Caldera, New Mexico: *J. Geophys. Res.*, v. 91, p. 1867–1886.

Hulston, J.R., and McCabe, W.J., 1962, Mass spectrometric measurements in the thermal areas of New Zealand: 1. Carbon dioxide and residual gas analysis: *Geochim. et Cosmochim. Acta*, v. 26, p. 383–397.

Humphreys, E.D., 1978, Telluric sounding and mapping in the vicinity of the Salton Sea geothermal area, Imperial Valley, California: M.S. Thesis, Univ. of Calif., Riverside, Dept. of Earth Sciences and Inst. of Geophys. and Planetary Phys. *Rept. UCR/IGPP-78/17*, 116 p.

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

Inman, J.R., 1975, Resistivity inversion with ridge regression: *Geophysics*, v. 40, p. 798–817.

Irving, E., 1964, Paleomagnetism and Its Application to Geological and Geophysical Problems: John Wiley & Sons, Inc., New York, 399 p.

Isherwood, W.F., 1975, Gravity and magnetic studies of the Geysers-Clear Lake geothermal region, California: U.S. Geol. Survey *Open-File Rept. 75-368*, 37 p.

Ishido, T., and Mizutani, H., 1981, Experimental and theoretical basis of electrokinetic phenomena in rock-water systems and its application to geophysics: *J. Geophys. Res.*, v. 86, p. 1763–1775.

Ivanov, V.V., 1967, Principal geochemical environments and processes of the formation of hydrothermal waters in regions of recent volcanic activity, *in* Vinogradov, A.P., ed., Chemistry of the Earth's Crust, v. 2: Jerusalem, Israel Prog. Scientific Translation Ltd., p. 260–281.

Iyer, H.M., 1979, Deep structure under Yellowstone National Park, U.S.A.: A continental hot-spot: *Tectonophysics*, v. 56, p. 165–197.

Iyer, H.M., and Hitchcock, T., 1976, Seismic noise as a geothermal exploration tool: Techniques and results: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975*, v. 2, p. 1075–1083.

Iyer, H.M., Oppenheimer, D.H., and Hitchcock, T., 1979, Abnormal *P*-wave delays in the Geysers-Clear Lake geothermal area, California: *Science*, v. 204, p. 495–497.

Jachens, R.C., and Roberts, C.W., 1985, Temporal and areal gravity investigations at Long Valley caldera, California: *J. Geophys. Res.*, v. 90, n. B13, p. 11210–11218.

Jackson, D.B., and O'Donnell, J.E., 1980, Reconnaissance electrical surveys in the Coso Range, California: *J. Geophys. Res.*, v. 85, n. B5, p. 2502–2516.

Japan's Sunshine Project, 1984, Summary of geothermal Energy R&D: Sunshine Project Promotion Headquarters, Agency of Industrial Science and Technology, Ministry of International Trade and Industry, Tokyo, 186 p.

Jupp, D.L.B., and Vozoff, K., 1977, Two-dimensional magnetotelluric inversion: *Geophys. J. Royal Astr. Soc.*, v. 50, p. 333–352.

Kam, M.N.S., 1980, Determination of Curie isotherm from aeromagnetic data in the Imperial Valley, California: M.S. Thesis, Univ. of Calif., Riverside, *UCR/IGPP-80/4/*.

Kappelmeyer, O., and Haenel, R., 1974, Geothermics with special reference to application: Geoexploration Monographs, series 1, n. 4, Gebrüder Borntraeger, Berlin.

Kaufman, A.A., and Keller, G.V., 1981, The Magnetotelluric Sounding Method: Elsevier, New York, 595 p.

_______, 1983, Frequency and Transient Soundings: Elsevier, New York, 685 p.

Keevil, N.B.J., and Ward, S.H., 1962, Electrolyte activity: Its effect on induced polarization: *Geophysics*, v. 27, p. 677–690.

Keller, G.V., and Frischknecht, F.C., 1966, Electrical Methods in Geophysical Prospecting: Pergamon Press, New York, p. 1–33.

Keller, G.V., and Furgerson, R.B., 1977, Determining the resistivity of a resistive layer in the crust: Am. Geophys. Union *Monograph 20*, p. 440–469.

Keller, G.V., and Jacobson, J.J., 1983a, Deep electromagnetic soundings northeast of The Geysers steam field: *Geoth. Res. Council, Trans.*, v. 7, p. 497–503.

_______, 1983b, Megasource electromagnetic survey in the Bruneau-Grandview area, Idaho: *Geoth. Res. Council, Trans.*, v. 7, p. 505–511.

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

Keller, G.V., Taylor, K., and Souto, J.M., 1982, Megasource EM method for detecting deeply buried conductive zones in geothermal exploration (abstr.): *Geophysics*, v. 47, n. 2, p. 420.

Keller, G.V., Pritchard, J.I., Jacobson, J.J., and Harthill, N., 1984, Megasource time-domain electromagnetic sounding methods, *in* Nabighian, M.N., ed., Time Domain Electromagnetic Methods of Exploration: *Geophysics, Spec. Iss.*, v. 49, p. 993–1009.

Klusman, R.W., and Landress, R.A., 1978, Secondary controls on mercury in soils of geothermal areas: *J. Geochem. Explor.*, v. 9, p. 75–91.

Klusman, R.W., Cowling, S.S., Culvey, B., Roberts, C., and Schwab, A.P., 1977, Preliminary evaluation of secondary controls on Hg in soils of geothermal districts: *Geothermics*, v. 6, p. 1–8.

Koga, A., and Noda, T., 1976, Geochemical prospecting in vapor-dominated fields for geothermal exploration: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975*, v. 1, p. 761–766.

Kristiansson, K., and Malmquist, L., 1982, Evidence for nondiffussive transport of $_{86}{}^{222}$Rn in the ground and a new physical model for the transport: *Geophysics*, v. 47, p. 1444–1452.

Kunetz, G., 1958, Principles and applications of telluric prospecting: Erdöl-Z., Heft 9, Urban-Verlag, Hamburg, p. 3–15 (in German).

Labovitz, M.L., Masuoka, E.J., Bell, R., Siegrist, A.W., and Nelson, R.F., 1983, The application of remote sensing to geobotanical exploration for metal sulfides–Results from the 1980 field season at Mineral, Virginia: *Econ. Geol.*, v. 78, p. 750–760.

Lachenbruch, A.H., 1968, Preliminary geothermal model of the Sierra Nevada: *J. Geophys. Res.*, v. 73, p. 6977–6989.

Lachenbruch, A.H., and Sass, J.H., 1977, Heat flow in the United States and the thermal regime of the crust, *in* Hencock, J.G., ed., The Nature and Physical Properties of the Earth's Crust: Am. Geophys. Union, *Monograph 20*.

_______, 1978, Models of an extending lithosphere and heat flow in the Basin and Range Province: *Geol. Soc. Am., Memoir 152*, p. 209-250.

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

Lange, A.L., and Westphal, W.H., 1969, Microearthquakes near The Geysers, Sonoma County, California: *J. Geophys. Res.*, v. 74, p. 4377–4378.

Lange, A.L., Pilkington, H.D., and Deymonay, J., 1982, Comparative studies of geothermal anomalies in 3-meter and temperature gradient holes: *Geoth. Res. Council, Trans.*, v. 6, p. 133–136.

Leary, P.C., and Henyey, T.L., 1985, Anisotropy and fracture zones about a geothermal well from P-wave velocity profiles: *Geophysics*, v. 50, no. 1, pp. 25–36.

Lee, K.H., 1978, Electromagnetic scattering by a two-dimensional inhomogeneity due to an oscillating magnetic dipole: Ph.D. Dissertation, Univ. of Calif., Berkeley, Dept. of Engineering Geosciences, and Lawrence Berkeley Laboratory, *Rept. LBL-8275*.

Lee, K.H., Pridmore, D.F., and Morrison, H.F., 1981, A hybrid three-dimensional electromagnetic modeling scheme: *Geophysics*, v. 46, p. 796–805.

Lee, T-C., 1977, On shallow-hole temperature measurements–A test study in the Salton Sea geothermal field: *Geophysics*, v. 42, p. 572–583.

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

LeShack, L.A., and Lewis, J.E., 1983, Geothermal prospecting with Shallow-Temp surveys: *Geophysics*, v. 48, p. 975–996.

Levinson, A.A., 1974, Introduction to Exploration Geochemistry: 2nd ed., Applied Publishing Co., Wilmette, Illinois, 924 p.

Liaw, A.L., and McEvilly, T.V., 1977, The role of ambient microseisms in geothermal exploration: *Geoth. Res. Council, Trans.*, v. 1, p. 187–188.

_______, 1979, Microseisms in geothermal exploration–Studies in Grass Valley, Nevada: *Geophysics*, v. 44, p. 1097–1115.

Liaw, A., and Suyenaga, W., 1982, Detection of geothermal microtremors using seismic arrays (abstr.): *52nd Annual International Meeting, Society of Exploration Geophysicists, Dallas*, p. 486–489.

Lippmann, M.J., and Mañón M., A., 1986, The Cerro Prieto geothermal field: Lawrence Berkeley Laboratory, *Rept. LBL-21796*.

Lyons, D.J., and van de Kamp, P.C., 1979, Subsurface geological study and geophysical study of the Cerro Prieto geothermal field: *Proc., 2nd Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, Mexicali, October 1979*, Comisión Federal de Electricidad, p. 173–186.

Madden, T., and Nelson, P., 1964, A defense of Cagnaird's magnetotelluric method: Tech. Report, Mass. Inst. of Tech., *ONR Project NR-371-401*.

Madden, T.R., and Park, S., 1982, MT modeling for a crustal environment: Mass. Inst. of Tech., Final Report, U.S. Geol. Survey *Grant 14-08-0001-G-643*, 74 p.

Mahon, W.A.J., 1976, Review of hydrogeochemistry of geothermal systems-prospecting, development and use: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975*, v. 1, p. 775–784.

Majer, E.L., 1978, Seismological investigations in geothermal regions: Ph.D. Thesis, Univ. of Calif., Berkeley.

Majer, E.L., and McEvilly, T.V., 1979, Seismological investigations at The Geysers geothermal field: *Geophysics,* v. 44, p. 246–269.

_______, 1981, Detailed microearthquake studies at the Cerro Prieto geothermal field: *Proc., 3rd Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, San Francisco, March 1981,* Lawrence Berkeley Laboratory, *Rept. LBL-11967,* p. 347–352.

Majer, E.L., McEvilly, T.V., Albores, A., and Díaz C., S., 1980, Seismological studies at Cerro Prieto: *Geothermics,* v. 9, n. 1/2, p. 79–88.

Majer, E.L., McEvilly, T.V., Eastwood, F., and Myer, L., 1986, Fracture detection using P- and S-wave VSP's at The Geysers geothermal field: Lawrence Berkeley Laboratory, *Rept. LBL-20100,* (submitted to *Geophysics*).

Major, J., and Silic, J., 1981, Restrictions on the use of Cole-Cole dispersion models in complex resistivity interpretation: *Geophysics,* v. 46, p. 916–931.

Mann, P., Hempton, M.R., Bradley, D.C., and Burke, K., 1983, Development of pull-apart basins: *J. Geol.,* v. 91, p. 529–554.

Mariner, R.H., Presser, T.S., Rapp, J.B., and Willey, L.M., 1975, The minor and trace elements, gas, and isotope compositions of the prinicipal hot springs of Nevada and Oregon: U.S. Geol. Survey *Open-File Report,* 27 p.

Marshall, D.J., and Madden, T.R., 1959, Induced polarization: A study of its causes: *Geophysics,* v. 24, p.790.

Matlick, J.S., and Buseck, P.R., 1976, Exploration for geothermal areas using mercury: A new geochemical technique: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 1, p. 785–792.

Mazor, E., 1976, Atmospheric and radiogenic noble gases in thermal waters: Their potential application to prospecting and steam production studies: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 1, p. 793–802.

McBirney, A.R., 1978, Volcanic evolution of the Cascade Range: *Am. Rev. Earth Planet. Sci.,* v. 6, p. 437–456.

McCarthy, K.P., 1983, Helium exploration survey for the Animas Valley, Colorado: *Geoth. Res. Council Bull.,* v. 12, n. 7, p. 9–14.

McEvilly, T.V., and Majer, E.L., 1982, ASP: An automated seismic processor for microearthquake networks: *Bull. Seis. Soc. Am.,* v. 72, p. 303–325.

McEvilly, T.V., and Schechter, B., 1978, East Mesa seismic study: Lawrence Berkeley Laboratory, *Rept. UCID 4030.*

McLaughlin, R.J., 1981, Tectonic setting of pre-Tertiary rocks and its relation to geothermal resources in the Geysers-Clear Lake area: U.S. Geol. Survey *Prof. Paper 1141,* p. 3–24.

McNerney, J.J., Buseck, P.R., and Hanson, R.C., 1972, Mercury detection by means of thin gold films: *Science,* v. 178, p. 611–612.

McNitt, J.R., 1973, The role of geology and hydrology in geothermal exploration, *in* Geothermal Energy, A Review of Research and Development: UNESCO, Paris, France, p.33–40.

McNitt, J.R., 1976, Summary of United Nations geothermal exploration experience, 1965 to 1975: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 2, p. 1127–1134.

Meidav, T., 1970, Application of electrical resistivity and gravimetry in deep geothermal exploration: *Proc., U.N. Symposium on the Development and Utilization of*

Geothermal Resources, Pisa, (Geothermics, Spec. Iss. 2), v. 2, pt. 1, p. 333.

Meyer, C., and Hemley, J.J., 1967, Wall rock alteration, *in* Barnes, H.L., ed., Geochemistry of Hydrothermal Ore Deposits: Holt, Rinehart and Winston, Inc., New York, p. 166–235.

Miller, C.D., 1985, Holocene eruptions at the Inyo volcanic chain, California: Implications for possible eruptions in Long Valley caldera: *Geology,* v. 13, p. 14–17.

Milton, N.M., 1983, Use of reflectance spectra of native plant species for interpreting airborne multispectral scanner data in the East Tintic Mountains, Utah: *Econ. Geol.,* v. 78, p. 761–769.

Mogro-Campero, A., and Fleischer, R.L., 1977, Subterrestrial fluid convection: A hypothesis for long-distance migration of radon within the earth: *Earth Plant. Sci. Lett.,* v. 34, p. 321–325.

Moore, J.N., Christensen, O.D., and Bamford, R.W., 1982, Mercury–Its role in the exploration of vapor-dominated systems (abstr.), *in* Stringfellow, J., ed., The Industry Coupled Case Study Program: Univ. of Utah Res. Inst., Earth Sciences Laboratory, *Rept. ESL-13,* 32 p.

Moos, D., 1983, Velocity, attenuation, and natural fractures in shallow boreholes: Ph.D. Thesis, Stanford University, California.

Morrison, H.F., Goldstein, N.E., Wilt, M.J., Lee, K.H., and Turnross, J.L., 1983, Controlled-source electromagnetic soundings at the Long Valley caldera, California (abstr.): *Eos, Trans., Am. Geophys. Union,* v. 64, n. 45, p. 890.

Morrison, H.F., Conti, U., Labson, V.F., Nichols, E., and Goldstein, N.E., 1984, Field tests of noise in SQUID and induction coil magnetometers: Lawrence Berkeley Laboratory, *Rept. LBID-901,* 10 p.

Moskowitz, B., and Norton, D., 1977, A preliminary analysis of intrinsic fluid and rock resistivity in active hydrothermal systems: *J. Geophys. Res.,* v. 82, n. 36, p. 5787–5795.

Mozley, E.C., 1982, An investigation of the conductivity distribution in the vicinity of a Cascade volcano: Ph.D. Thesis, Univ. of Calif., Berkeley, Lawrence Berkeley Laboratory, *Rept. LBL-15671,* 385 p.

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

Myer, L.R., Hopkins, D., and Cook, N.G.W., 1985, Effects of contact area of an interface on acoustic wave transmission characteristics, *in* Balkenia, A.A., ed., Research and Engineering Applications in Rock Masses: *Proc., 26th U.S. Symp. on Rock Mechanics,* South Dakota School of Mines, v. 1, p. 565–572.

Nabighian, M.N., ed., 1984, Time domain electromagnetic methods of exploration: *Geophysics, Spec. Iss.,* v. 49, n. 7.

Nagata, T., 1961, Rock Magnetism: Maruzen Company Ltd., Tokyo, 350 p.

Nelson, P.H., 1977, Induced polarization effects from grounded structures: *Geophysics,* v. 42, p. 1241–1253.

Nettleton, L.L., 1976, Gravity and Magnetics in Oil Prospecting: McGraw Hill, New York, 463 p.

Neuenschwander, E.F., and Metcalf, D.F., 1942, A study of electrical earth noise: *Geophysics,* v. 7, n. 1, p. 69–77.

Nielson, D.L., 1978, Radon emanometry as a geothermal exploration technique: Theory and an example from Roosevelt Hot Springs KGRA, Utah: Univ. of Utah Res. Inst., *Rept. ICO/78.*

Nielson, D.L., Sibbett, B.S., and McKinney, D.B., 1978, Geology of Roosevelt Hot Springs

KGRA, Beaver County, Utah: Univ. of Utah Res. Inst., Earth Sciences Laboratory, *Rept. 12,* 121 p.

_______, 1979, Geology and structural control of the geothermal system at Roosevelt Hot Springs, Beaver County, Utah (abstr.): *Am. Assoc. Petrol. Geol. Bull.,* v. 63, p. 836.

Nourbehecht, B., 1963, Irreversible thermodynamic effects in inhomogeneous media and their application in certain geoelectric problems: Ph.D. Thesis, Mass. Inst. of Tech.

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

Oldenburg, D.W., 1979, One-dimensional inversion of natural source magnetotelluric observations: *Geophysics,* v. 44, p. 1218–1244.

Olhoeft, G.R., 1981, Electrical properties of granite with implications for the lower crust: *J. Geophys. Res.,* v. 86, n. B2, p. 931–936.

Olhoeft, G.R., 1983, Electrical properties of water- and steam-saturated zeolite (abstr.): *Eos, Trans. Am. Geophys. Union,* v. 64, n. 45, p. 692.

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

Olorunfemi, M.O., and Griffiths, D.H., 1985, A laboratory investigation of the induced polarization of the Triassic Sherwood sandstone of Lancashire and its hydrogeological applications: *Geophys. Pros.,* v. 33, p. 110–217.

Onodera, S., 1982, Determination of drilling target in Onikobe geothermal area: *U.S. Geoth. Res. Council, Trans.,* v. 6, p. 149–152.

Orellana, E., and Mooney, H.M., 1966, Master tables and curves for vertical electric sounding over layered structures: *Madrid Interciscia,* 150 p., 66 tables.

Pálmasson, G., 1976, Geophysical methods in geothermal exploration: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 2, p. 1175–1184.

Panichi, C., Ferrara, G.C., and Gonfiantini, R., 1976, Isotope geothermometry in the Larderello geothermal field: *Geothermics,* v. 5, p. 81–88.

Park, S.K., Orange, A.S., and Madden, T.R., 1983, Effects of three-dimensional structure on magnetotelluric sounding curves: *Geophysics,* v. 48, p. 1402–1405.

Pelton, W.H., Ward, S.H., Hallof, P.G., Sill, W.R., and Nelson, P.H., 1978, Mineral discrimination and removal of inductive coupling with multifrequency IP: *Geophysics,* v. 43, p. 588–609.

Phelps, D., and Buseck, P.R., 1980, Distribution of soil mercury and the development of soil mercury anomalies in the Yellowstone geothermal area, Wyoming: *Econ. Geol.,* v. 75, p. 730–740.

Piller, H., 1977, Microscope Photometry: Springer-Verlag, New York, 253 p.

Podwysocki, M.H., Segal, D.B., and Abrams, M.J., 1983, Use of multispectral scanner images for assessment of hydrothermal alteration in the Marysvale, Utah, mining area: *Econ. Geol.,* v. 78, p. 675–687.

Poppendiek, H.F., Connelly, D.J., and Sellers, A.J., 1982, Development of downhole heat flux and thermal conductivity transducers: *Geoth. Res. Council, Trans.,* v. 6, p. 161–164.

Quist, A.S., and Marshall, W.L., 1968, Electrical conductances of aqueous sodium chloride solutions from 0 to 800° and pressures to 4000 bars: *J. Phys. Chem.,* v. 72, p. 684–702.

Reader, J.F., and Fairbank, B.D., 1983, Heat flow in the vicinity of the Meager Volcanic Complex, southwestern British Columbia: *Geoth. Res. Council, Trans.,* v. 7, p. 535–539.

Reasenberg, P., Ellsworth, W., and Walter, A., 1980, Teleseismic evidence for a low-velocity body under the Coso geothermal area: *J. Geophys. Res.,* v. 85, p. 2471–2487.

Reitsema, R.H., 1979, Gases of mud volcanoes in the Copper River Basin, Alaska: *Geochim. et Cosmochim. Acta,* v. 43, p. 183–187.

Rigby, F., and McEuen, R.B., 1982, An example of a resistivity log extrapolation with magnetotelluric data for a geothermal site: *J. Pet. Tech.,* November, p. 2683–2688.

Risk, G.F., 1981, Induced polarization (IP) anomaly in the Broadlands geothermal field: *Proc., New Zealand Geothermal Workshop, 1981, Auckland, N.Z.,* Univ. of Auckland, p. 43–48.

Risk, G.F., Macdonald, W.J.P., and Dawson, C.B., 1970, DC resistivity surveys at the Broadlands geothermal region, New Zealand: *Geothermics, Spec. Iss. 2,* p. 287–294.

Roberts, A.A., Friedman, I., Donovan, T.J., and Denton, E.H., 1975, Helium survey, a possible technique for locating geothermal reservoirs: *Geophys. Res. Lett.,* v. 2, p. 209–210.

Robbins, J.C., 1973, Zeeman spectrometer for measurement of atmospheric mercury vapor: *Geochem. Expl. 1972,* IMM, London, p. 315–323.

Robinson, P.T., and Elders, W.A., 1976, Quaternary volcanism in the Salton Sea geothermal field, Imperial Valley, California: *Geol. Soc. Am. Bull.,* v. 87, p. 347–360.

Robinson, R., 1981, A microearthquake survey at the Ngawha geothermal field, New Zealand: *Geophysics,* v. 46, p. 1467–1468.

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

Roquemore, G.R., 1984, Ground magnetic survey in the Coso Range, California: *J. Geophys. Res.,* v. 89, n. B5, p. 3309–3314.

Ross, H.P., Nielson, D.L., and Moore, J.N., 1982, Roosevelt Hot Springs geothermal system–Case study: *Am. Assoc. Petrol. Geol. Bull.,* v. 66, n. 7, p. 879–902.

Ross, L.W., Brown, L.L., and Williams, R.E., 1977, Pressure and temperature measurements and sampling in geothermal wells: *Geoth. Res. Council, Trans.,* v. 1, p. 265–267.

Roy, A., and Apparao, A., 1971, Depth of investigation in direct current methods: *Geophysics,* v. 36, p. 943–959.

Roy, R.F., Decker, E.R., Blackwell, D.D., and Birch, F., 1968, Heat flow in the United States: *J. Geophys. Res.,* v. 73, p. 5207–5221.

Ryall, A.S., and Ryall, F.D., 1982, Propagation effects and seismicity associated with magma in Long Valley caldera, eastern California (abstr.): Seis. Soc. Am., *Earthquake Notes,* v. 53, p. 46–47.

Rybach, L., 1981, Geothermal systems, conductive heat flow, *in* Rybach, L., and Muffler, L.J.P., eds., Geothermal systems, Principles and Case Histories: John Wiley & Sons, New York, p. 3–36.

Rybach, L., and Stegena, L., eds., 1979, Geothermics and Geothermal Energy: Birkhäuser Verlag, Basel, 341 p.

Rymer, H., and Brown, G.C., 1986, Gravity fields and the interpretation of volcanic structures: Geological discrimination and temporal evolution: *J. Volcan. Geotherm. Res.,* v. 27, n. 3/4, p. 229–254.

Sammel, E.A., 1983, The shallow hydrothermal system at Newberry Volcano, Oregon: A conceptual model: *Geoth. Res. Council, Trans.,* v. 7, p. 325–330.

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

Sanford, S.J., 1981, Dating thermal events at Cerro Prieto using fission track annealing: M.S. Thesis, Univ. of Calif., Riverside, Inst. of Geophys. and Planetary Phys., *Rept. 81/5,* 167 p.

Sanford, S.J., and Elders, W.A., 1981, Dating thermal events at Cerro Prieto using fission track annealing: *Proc., 3rd Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, San Francisco, March 1981,* Lawrence Berkeley Laboratory, *Rept. LBL-11967,* p. 114–119.

Sasaki, Y., 1982, Automatic interpretation of induced polarization data over two-dimensional structures: Mem. Faculty of Engineering, Kyushu Univ., v. 42, n. 1, p. 59–74.

Sass, J.H., Kennelly, J.P., Wendt, W.E., Moses, T.H., Jr., and Ziagos, J.P., 1981, In-situ determination of heat flow in unconsolidated sediments: *Geophysics,* v. 46, p. 76–83.

Sawyer, R.F., and Cook, K.L., 1977, Gravity and ground magnetic surveys of the Thermo Hot Springs KGRA region, Beaver County, Utah: Univ. of Utah, Dept. of Geology and Geophysics, 142 p.

Schoenberg, M., 1980, Elastic wave behavior across linear slip interfaces: *J. Acoust. Soc. Am.,* v. 68, no. 5, pp. 1516–1521.

Schoenberg, M., 1983, Reflection of elastic waves from periodically stratified media with interfacial slip: *Geophys. Prosp.,* v. 31, pp. 265–292.

Seigel, H.O., 1959, Mathematical formulation and type curves for induced polarization: *Geophysics,* v. 24, p. 547–565.

Sharman, G.F., Reicale, M.S., and Brune, J.N., 1976, Detailed study of relative plate motion in the Gulf of California: *Geology,* v. 4, p. 206–210.

Shuey, R.T., Schellinger, D.K., Tripp, A.C., and Alley, L.B., 1977, Curie depth determination from aeromagnetic data: *Geophys. J. Roy. Astr. Soc.,* v. 50, p. 75–102.

Sibbett, B.S., 1983, Structural control and alteration at Beowawe KGRA, Nevada: *Geoth. Res. Council, Trans.,* v. 7, p. 187–191.

Siegal, B.S., and Gillespie, A.R., 1980, Remote sensing in geology: John Wiley & Sons, Inc., New York, 702 p.

Silberman, M.L., and White, D.E., 1975, Limits on the duration of hydrothermal activity at Steamboat Springs, Nevada, by K-Ar ages of spatially associated altered and unaltered volcanic rocks (abstr.): *Geol. Soc. Am. Abstr. with Prog.,* v. 7, n. 7, p. 1272–1273.

Sill, W.R., 1982, Self-potential effects due to hydrothermal convection-velocity cross coupling: Univ. of Utah, Dept. of Geology and Geophysics, *Rept. DOE/ID/12079-68.*

_______, 1983, Self-potential modeling from primary flows: *Geophysics,* v. 48, p. 76–86.

Sill, W.R., and Johng, D.S., 1979, Self-potential survey, Roosevelt Hot Springs, Utah: Univ. of Utah Dept. of Geol. and Geophysics, *Topical Rept. IDO/78-1701.a.2.3.*

Simmons, G., 1967, Interpretation of heat flow anomalies, 1, Contrast in heat production: *Rev. of Geophys.,* v. 5, n. 1, p. 42–52.

Smith, R.L., and Bailey, R.A., 1968, Resurgent cauldrons, *in* Coats, R.R., Hay, R.L., and Anderson, C.A., eds., Studies in Volcanology: *Geol. Soc. Am., Memoir 116,* p. 613–662.

Smith, R.L., and Shaw, H.R., 1973, Volcanic rocks as geologic guides to geothermal exploration and evaluation: *Eos, Am. Geophys. Union Trans.,* v. 54, p. 1213.

_______, 1975, Igneous-related geothermal systems, *in* White, D.E., and Williams, D.L., eds., Assessment of Geothermal Resources of the United States–1975: U.S. Geol. Survey *Circular 726,* p. 58–83.

Sorey, M.L., Lewis, R.E., and Olmstead, F.H., 1978, The hydrothermal system of Long Valley caldera, California: U.S. Geol. Survey *Prof. Paper 1044A.*

Spector, A., and Grant, F.S., 1970, Statistical models for interpreting aeromagnetic data: *Geophysics,* v. 35, p. 293–302.

Stanley, W.D., 1982, A regional magnetotelluric survey of the Cascade Mountains region: U.S. Geol. Survey *Open-File Rept. 82-126.*

Stanley, W.D., and Tinkler, R.D., 1982, A practical low-noise coil system for magnetotellurics: U.S. Geol. Survey *Open-File Rept. 83-85.*

Stearns, D.W., and Friedman, M., 1972, Reservoirs in fractured rock, *in* Stratigraphic Oil and Gas Fields: *Am. Assoc. Petrol. Geol., Memoir 16,* p. 82–106.

Steele, J.L., and Blackwell, D.D., 1982, Heat flow in the vicinity of the Mount Hood volcano, Oregon, *in* Priest, G.R., and Vogt, B.F., eds., Geology and Geothermal Resources of the Mount Hood Area, Oregon: Department of Geology and Mineral Industries, Oregon, *Spec. Paper 14,* p. 31–42.

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

Stodt, J.A., 1979, Documentation of a finite element program for solution of geophysical problems governed by the inhomogeneous 2-D scalar Helmholtz equation: Univ. of Utah, Technical Report, NSF Grant AER 76-11155.

Stoker, A., and Kruger, P., 1976, Radon in geothermal reservoirs: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 2, p. 1797–1804.

Stoyer, C.H., 1982, Time-domain electromagnetic method: *Mines Magazine,* Colorado School of Mines, Golden, v. 74, n. 8, p. 9.

Sumner, J.S., 1976, Principles of Induced Polarization for Geophysical Exploration: Elsevier, New York, 277 p.

Swift, C.M., Jr., 1967, A magnetotelluric investigation of electrical conductivity anomaly in the southwestern United States: Ph.D. Thesis, Mass. Inst. of Tech.

Tanner, A.B., 1964, Radon migration in the ground: A review, *in* Adams, J.A.S., and Lowder, W.M., eds., The Natural Radiation Environment: Univ. of Chicago Press, p. 161–190.

_______, 1980, Radon migration in the ground: A supplementary review, *in* Gesell, T.F., and Lowder, W.M., eds., Natural Radiation Environment III: Symp. Proc., Houston, April 1978, v. 1, p. 5–56.

Tasci, M.T., 1975, Exploration for a geothermal system in the Lualualei Valley, Oahu, Hawaii: Appendix A, *in* Grose, L.T., and Keller, G.V., Geothermal Energy in the Pacific region: Colo. School of Mines.

Taylor, B.E., and Gerlach, T.M., 1983, Chemical and isotopic composition of Casa Diablo Hot Spring: Magmatic CO_2 near Mammoth Lakes, Calif. (abstr.): Am. Geophys. Union Poster Paper presented at AGU Fall Meeting, December 3–7, 1983, San Francisco.

Telford, W.M., Geldart, L.P., Sheriff, R.E., and Keys, D.A., 1976, Applied Geophysics: Cambridge University Press, Cambridge, 860 p.

Ting, S.C., and Hohmann, G.W., 1981, Integral equation modeling of three-dimensional magnetotelluric response: *Geophysics,* v. 46, p. 182–197.

Toksöz, M.N., Johnston, D.H., and Timur, A., 1979, Attenuation of seismic waves in dry and saturated rocks: 1. Laboratory measurements: *Geophysics,* v. 44, p. 681–690.

Torgersen, T., and Jenkins, W.J., 1982, Helium isotopes in geothermal systems: Iceland, The Geysers, Raft River and Steamboat Springs: *Geochim. et Cosmochim. Acta,* v. 46, p. 739–748.

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

Tripp, A.C., Hohmann, G.W., and Swift, C.M., Jr., 1984, Two-dimensional resistivity inversion: *Geophysics,* v. 49, p. 1708–1717.

Truesdell, A.H., 1984, Chemical geothermometers for geothermal exploration, *in* Henley, R.W., Truesdell, A.H., and Barton Jr., P.B., eds., Fluid-Mineral Equilibria in Hyrothermal systems: *Rev. Econ. Geol.,* v. 1, p. 31–43.

Truesdell, A.H., 1976, Summary of Section III–Geochemical techniques in exploration: *Proc., 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco, May 1975,* v. 1, p. *liii-lxiii.*

Truesdell, A.H., and Fournier, R.O., 1977, Procedure of estimating the temperature of a hot water component in a mixed water using a plot of dissolved silica vs. enthalpy: U.S. Geol. Survey, *J. Res.,* v. 5, n. 1, p. 49–52.

Truesdell, A.H., Thompson, J.M., Coplen, T.B., Nehring, N.L., and Janik, C.J., 1981, The origin of the Cerro Prieto geothermal brine: *Proc., 2nd Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, Mexicali, October 1979,* Comisión Federal de Electricidad, p. 224–240.

Ucok, H., Ershaghi, I., and Olhoeft, G.R., 1980, Electrical resistivity of geothermal brines: *J. Pet. Tech.,* April, p. 717–727.

Varekamp, J.C., and Buseck, P.R., 1983, Hg anomalies in soils: A geochemical exploration method for geothermal areas: *Geothermics,* v. 12, n. 1, p. 29–47.

Vonder Haar, S., and Howard, J.H., 1981, Intersecting faults and sandstone stratigraphy at the Cerro Prieto geothermal field: *Geothermics,* v. 10, n. 3/4, p. 145–168.

Von Herzen, R., and Maxwell, A.E., 1959, The measurement of thermal conductivity of deep-sea sediments: *J. Geophys. Res.,* v. 64, p. 1557–1563.

Vozoff, K., 1972, The magnetotelluric method in the exploration of sedimentary basins: *Geophysics,* v. 37, p. 98–141.

Wannamaker, P.E., and Hohmann, G.W., 1982, Electromagnetic modeling of three-dimensional bodies in layered earths using integral equations: Univ. of Utah Res. Inst., Earth Sciences Laboratory, *Rept. DOE/ID/12079-50,* 51 p.

Wannamaker, P.E., Ward, S.H., Hohmann, G.W., and Sill, W.R., 1980, Magnetotelluric models of the Roosevelt Hot Springs Thermal Area, Utah: Univ. of Utah, *Topical Rept. DOE/ET/27002-8,* 213 p.

Wannamaker, P.E., Ward, S.H., Hohmann, G.W., and Sill, W.R., 1983, Deep resistivity structure in southwestern Utah and its geothermal significance: Univ. of Utah, Earth Science Laboratory, *Rept. DOE/ID/12079-89,* 96 p.

Ward, D.M., 1978, Thermophilic methanogenesis in a hot-spring algal bacterial mat: *Appl. Environ. Microbiol.,* v. 35, p. 1019–1026.

Ward, S.H., 1977, Report on workshop on electrical methods in the geothermal environment: *Geophysics,* v. 42, p. 664–666.

Ward, S.H., and Sill, W.R., 1982, Resistivity, induced polarization and self-potential methods in geothermal exploration: Univ. of Utah Res. Inst., Earth Sciences Laboratory, DOE/ID/12079-90, *Rept. ESL-108,* 100 p.

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

Waxman, M.H., and Smits, L.J.M., 1968, Electrical conductivities in oil-bearing shaly sand: *Soc. Petrol. Eng. J.,* v. 8, p. 107–122.

Weaver, C.S., Green, S.M., and Iyer, H.M., 1982, Seismicity of Mount Hood and structure as determined from teleseismic P-wave delay studies: *J. Geophys. Res.,* v. 87(B4), p. 2782–2792.

Webring, M., 1985, SAKI: A FORTRAN program for generalized linear inversion of gravity and magnetic profiles: U.S. Geol. Survey *Open-File Rept. 85-122,* 108 p.

White, D.E., 1957a, Thermal waters of magmatic origin: *Geol. Soc. Am. Bull.,* v. 68, p. 1637–1658.

_______, 1957b, Magmatic, connate and metamorphic waters: *Geol. Soc. Am. Bull.,* v. 68, p. 1659–1682.

_______, 1965, Saline waters of sedimentary rocks, *in* Fluids in Subsurface Environment–A Symposium: *Am. Assoc. Petrol. Geol., Memoir,* v. 4, p. 342.

_______, 1974, Diverse origins of hydrothermal fluids: *Econ. Geol.,* v. 69, p. 954–973.

White, D.E., Hem, J.D., and Waring, G.A., 1963, Chemical composition of subsurface waters, *in* Fleischer, M., ed., Data of Geochemistry: U.S. Geol. Survey *Prof. Paper 440-F,* p. F1–F67.

Whitehead, N.E., Gingrich, J.E., and Fisher, J.C., 1983, A New Zealand test of the Track-Etch® method of prosecting for geothermal steam: *J. Volcan. and Geoth. Res.,* v. 15, p. 339–354.

Williams, D.L., Berkman, F., and Mankinen, E.A., 1976, Implications of a magnetic model of the Long Valley caldera, California: *J. Geophys. Res.,* v. 82, n. 20, p. 3030–3038.

Williams, D.L., Hull, D.A., Ackermann, H.D., and Beeson, M.H., 1982, The Mt. Hood region: Volcanic history, structure and geothermal energy potential: *J. Geophys. Res.,* v. 87, n. B4, p. 2767–2781.

Wilt, M.J., and Goldstein, N.E., 1983, Electromagnetic soundings over a geothermal reservoir in Dixie Valley, Nevada: Lawrence Berkeley Laboratory, *Rept. LBL-15526.*

Wilt, M., and Vonder Haar, S., 1986, A geological and geophysical appraisal of the Baca geothermal field: *J. Volcan. Geotherm. Res.,* v. 27, n. 3/4, p. 349–370.

Wilt, M.J., Goldstein, N.E., and Razo M., A., 1978, LBL resistivity studies at Cerro Prieto: *Geothermics, Spec. Iss., Proc., 1st Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, Mexicali, 1979,* Comisión Federal de Electricidad, Part 2, p. 15–26.

Wilt, M., Goldstein, N.E., Stark, M., Haught, J.R., and Morrison, H.F., 1983, Experience with the EM-60 electromagnetic system for geothermal exploration in Nevada: *Geophysics,* v. 48, p. 1090–1101.

Winkler, K.W., and Nur, A., 1982, Seismic attenuation: Effects of pore fluids and frictional sliding: *Geophysics,* v. 47, p. 1–15.

Wollenberg, H.A., 1974, Radon alpha-track survey of potential geothermal resource area: Lawrence Berkeley Laboratory, *Rept. LBL-3225.*

Won, I.J., and Son, K.H., 1983, Determination of depth to the Curie isotherm from aeromagnetic data, *in* Workshop on Exploration for Hot Dry Rock Geothermal Systems: Los Alamos National Laboratory, *Rept. LA-9697-C,* p. 51–57.

Word, D.R., Smith, H.W., and Bostic, F.X., Jr., 1970, An investigation of the magnetotelluric tensor impedance method: Univ. of Texas, Electrical Geophys. Res. Laboratory, *Tech. Rept. 82.*

Worthington, P.F., 1982, The influence of shale effects on the electrical resistivity of reservoir rocks: *Geophys. Pros.,* v. 30, p. 673–687.

Wynn, J.C., 1974, Electromagnetic coupling in induced polarization: Ph.D. Dissertation, Univ. of Arizona, University Microfilms, Ann Arbor, Michigan.

Wynn, J.C., and Zonge, K.L., 1975, EM coupling, its intrinsic value, its removal and the cultural coupling problem: *Geophysics,* v. 40, p. 831–851.

_______, 1977, Electromagnetic coupling: *Geophys. Pros.,* v. 25, p. 29–51.

Yungul, S.H., 1965, Telluric current method of determining ellipse area by simultaneously measuring two voltages with a collinear three electrode array: U.S. Patent Office, *No. 3,188,559.*

Yungul, S.H., 1977, The telluric methods in the study of sedimentary structure–A survey: *Geoexploration,* v. 15, p. 207–238.

Yungul, S.H., Hembree, M.R., and Greenhouse, J.P., 1973, Telluric anomalies associated with isolated reefs in the Midland Basin, Texas: *Geophysics,* v. 38, n. 3, p. 545–556.

Zhody, A.A.R., 1973, A computer program for the automatic interpretation of Schlumberger soundings over horizontally layered media: Nat. Tech. Information Service, Springfield, Va., *PB232 702/A5,* 25 p.

Zhody, A.A.R., and Bisdorf, R.J., 1982, Schlumberger Soundings in the Medicine Lake area, California: U.S. Geol. Survey *Open-File Rept. 82-887,* 162 p.

Zhody, A.A., Anderson, L.A., and Muffler, L.J.P., 1973, Resistivity, self-potential and induced polarization surveys over a vapor-dominated geothermal system: *Geophysics,* v. 38, p. 1130–1144.

LAWRENCE BERKELEY LABORATORY
TECHNICAL INFORMATION DEPARTMENT
UNIVERSITY OF CALIFORNIA
BERKELEY, CALIFORNIA 94720

EXPLORATION STRATEGIES FOR REGIONAL ASSESSMENT OF HYDROTHERMAL RESOURCES

by

Stanley H. Ward
Duncan Foley
Joseph N. Moore
Dennis L. Nielson
Howard P. Ross
Phillip M. Wright

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

LIST OF FIGURES

Figure 1 Heat Flow in the U. S.

Figure 2 Hydrothermal Provinces in the U. S.

Figure 3 Exploration Strategy

Figure 4 Geologic Map of the Roosevelt Hot Springs, Utah, Geothermal Area and the Adjacent Mineral Mountains

Figure 5 Boron and Chloride in Wells and Springs in the Milford, Utah, Area

Figure 6 Portion of the Aeromagnetic Map of Utah

Figure 7 Portion of the Simple Bouguer Gravity Anomaly Map of Utah

ABSTRACT

This chapter briefly reviews the exploration strategies and techniques in general use for regional assessment of hydrothermal resources. The objective of regional assessment is to select from a large region those several areas that have highest potential for occurrence of a resource. Detailed exploration can then be carried on in these smaller areas. Most of the known hydrothermal resources and all those of high-temperature in the U. S. occur in the west including Hawaii and Alaska. There is diversity in the nature of and controls on hydrothermal resources that must be understood for the particular region under consideration before a specific assessment program, designed to optimize the ever-present trade-off between certainty of resource detection and cost, can be applied. Techniques from geology, geochemistry, geophysics and hydrology can then be applied in the most cost-effective combination during the assessment.

OBJECTIVE

It is ~~our~~ the objective of this ~~chapter~~ article to describe exploration methodologies and their integration as applied to regional assessment of hydrothermal resources in the United States. Achieving this objective necessitates describing, briefly, ~~our~~ the state of knowledge of regional geothermal resources in the U.S., outlining a basic exploration strategy suited to defining these resources in a regional context and, finally, discussing the geological, geochemical, and geophysical ingredients of the basic strategy. The basic strategy is expected to have variants suited to each region or even subsets of these regions. ~~We must also clearly state that~~ Technology appropriate to the task of reliable regional assessment is in an early stage of development as evidenced by poor success ratios for new wildcat geothermal wells (Ehni, 1981). Development of better technology is needed but ~~is probably not forthcoming from the private sector~~ will be delayed because of budget constraints brought about largely by the fact that geothermal development is ~~only~~ most economic today at a few of the highest grade resources.

The exploration methodologies and philosophies ~~which we~~ outlined in this ~~paper~~ ~~chapter~~ article have been formulated through ~~our~~ work with industry in ~~The~~ the Department of Energy (DOE) Industry Coupled Program. In addition ~~our work in~~ exploration technique development has been supported under DOE's Exploration and Assessment Technology Program. This Federal support has ~~greatly~~ benefited geothermal exploration in recent years by placing exploration data in the public domain and by sponsoring research which is specifically directed toward improving the state-of-the-art of geothermal exploration.

BACKGROUND

Geothermal energy is derived from the heat of the earth.

> *"The Earth's interior is a gigantic but delicately balanced heat engine fueled by radioactivity, which has much to do with how the surface evolved. Were it running more slowly, geological activity would have proceeded at a slower pace.*
> *...if there had been more radioactive fuel and a faster running engine, volcanic gas and dust would have blotted out the Sun, the atmosphere would have been oppressively dense, and the surface would have been racked by daily earthquakes and volcanic explosions."* *(Press and Siever, 1974).*

The earth, fortunately, is a heat engine running at ~~exactly~~ the right speed for ~~our~~ man's survival. The average heat flowing out through the earth's surface is 0.08 watts per m^2. If ~~we multiply~~ this value is multiplied by the total surface area of the earth (5.1×10^{14} m^2) ~~we~~ one obtains the total heat flowing from the earth as 41,000,000 megawatts. Only a fraction of this energy can be extracted economically under current market conditions. However, the crust of the earth contains local hot spots from which extraction of energy, either for direct heat applications or for conversion to electricity, is economical at present. Geothermal hot spots are manifested as a continuum of seven accepted resource types: convective hydrothermal, normal geothermal gradient, deep sedimentary basin, geopressured, radiogenic, hot dry rock, and magma. We shall be concerned in this article only with the convective hydrothermal, normal geothermal gradient, and deep sedimentary basin types of resources; collectively they are usually referred to as *hydrothermal resources* and all involve movement of thermal water.

Hot-water-dominated convective hydrothermal systems are generally classified as high temperature (>150°C), intermediate temperature (90 to 150°C), and low temperature (<90°C) (White and Williams, 1975; Muffler, 1979). Although some of these systems may derive their heat from still molten or hot, crystallized plutonic masses (Smith and Shaw, 1975), others show no association with recent plutonic or volcanic activity but derive their heat from deep circulation along fault zones in areas of high thermal gradients. Any regional exploration program for hydrothermal resources should be based on the search for the components of hydrothermal systems, heat, water, and permeability, rather than the unlikely and obvious occurrence of these components together. unclear

(4)

HYDROTHERMAL SYSTEMS IN THE U.S.

Introduction

The U.S. may be divided into two broad heat flow provinces which have much local and regional variation (Figure 1). The eastern U.S. is generally lower in heat flow with localized regions of higher heat flow occurring in areas of radiogenic plutons. The broader zones of higher heat flow in the West are due to higher thermal flux from the mantle (Simmons and Roy, 1969). This higher flux implies that the West, as a region, presents better opportunities to find hydrothermal systems. Indeed, the preponderance of known hydrothermal resources and all of the known high-temperature resources occur in the West, including Hawaii and Alaska. Young volcanic systems in the West form the thermal targets of potentially highest grade.

The prospects for alternate systems is discussed in chapter

At ~~the time of this writing~~ present, economically viable development of hydrothermal systems requires that adequate water be naturally present. ~~This~~ Today, ~~means~~ a system must have both sufficient quantities of water available for recharge and adequate permeability, most commonly through fractures. Regional identification of fracture systems can help guide the explorationist to favorable terrains of fracture permeability.

The U.S. may be divided into hydrothermal provinces (Figure 2) based on regional heat-flow patterns, similarity of geologic environments and similarity of resource target models. The discussion below is based on these provinces, the margins of which, particularly west of the Great Plains, form attractive regional exploration targets in many places.

Alaska and Hawaii

Areas of young volcanic activity form the prime thermal targets in Alaska and Hawaii. In Alaska, most of the volcanos are found along the Alaska Peninsula and in the Aleutian Island Chain. Secondary Alaskan targets for hydrothermal resources are zones of high heat flow along regional strike-slip faults, areas of deep circulation, and deep sedimentary basins (Turner et al., 1980; Motyka et al., 1980; Motyka and Moorman, 1981). In Hawaii, volcanic rift zones around the margins of active volcanos form the best targets found to date (Thomas et al., 1980; Furumoto, 1978). Electric power development from the Puna rift resource is continuing and shows substantial promise (Chen et al., 1980).

Cascade Mountains

The young volcanos of the Cascade range of Washington, Oregon, and California have long been an attractive, but enigmatic, hydrothermal terrain. Although the resource potential is high (Brook et al., 1979; Youngquist, 1981), institutional barriers have slowed development (Bloomquist, 1981). Extensive resource investigations at Mt. Hood have failed to demonstrate an electric quality resource, although direct heat applications are planned (Riccio, 1979; Bowen, 1981). Recent U.S. Geological Survey drilling at Newberry caldera in central Oregon has demonstrated, however, that 265°C resources may be hidden beneath cold near-surface groundwater regimes (Sammel, 1981).

(site of a national park)

The recent eruption of Mt. St. Helens, the high-temperature fumeroles on Mt. Baker and Mt. Hood, the vapor-dominated hydrothermal system at Mt. Lassen, and the ongoing geothermal development at Meager Creek in British Columbia (Fairbanks et al., 1981) all suggest that there are extensive resources yet to be discovered and developed in the Cascades.

systems, such as Thermopolis, Wyoming, apparently have water that circulates to depth in stratigraphic horizons and rises to the surface along the limbs of anticlines (Heasler, 1981). Exploration programs in the Rocky Mountains should be targeted to identify favorable deep circulation paths for water because local heat sources, such as very young volcanos, are rare.

Great Plains

Hydrothermal resources in the Great Plains are dominated by the Madison Group of Paleozoic carbonate rocks. These rocks are found predominantly in Montana, Wyoming, and the Dakotas. Temperatures are suitable for direct applications. In North Dakota, much of the water is of poor quality (Harris et al., 1980). In Nebraska, waters are found in the Cretaceous Dakota Formation which, although cooler than the deep waters in the Madison (Gosnold and Eversoll, 1981), typically have much better water quality.

Regional exploration in the Great Plains is based on a deep sedimentary basin model, where exploration is concerned with identification of appropriate stratigraphic horizons at suitable depths for the thermal gradient to provide sufficient temperatures. Areas of anomalously high heat flow within these basins can produce water several tens of degrees hotter than water elsewhere at the same depth.

Ouachita Belt

Hydrothermal resources in central Texas are principally found in Cretaceous sandstone and limestone aquifers along the buried Ouachita fold belt (Woodruff and McBride, 1979). Water heated in the thermal gradient of the earth circulates either downward from the west or upward from the deeper parts of the gulf basin to the east. Downward-circulating waters are typically fresh while upward-circulating waters are more saline. Woodruff and

McBride (1979) have identified a broad region favorable for geothermal resources. Exploration within this region is directed toward identifying suitable stratigraphic horizons at sufficient depth.

Rio Grande Rift

Non-volcanic related-hydrothermal systems in the Rio Grande rift are apparently the result of upward flow of waters from deep rift basins at groundwater constrictions (Harder et al., 1980; Morgan et al., 1981). These geothermal resources have potential for direct applications but one related to volcanic rocks is being explored for generation of electricity in the Valles Caldera in northern New Mexico (Dondanville, 1978). Also, a deep system suitable for electric generation may exist in the vicinity of Soccorro, New Mexico (Chapin et al., 1978).

Exploration for resources suitable for direct heat application should focus on identification of groundwater flow patterns in basins, particularly in areas where constrictions may cause water to flow to the surface, while exploration for resources suited to electric power generation should focus on local regions of young volcanic activity.

Colorado Plateau

Isolated, generally cool, and low-flow rate hydrothermal systems exist in the Colorado Plateau. Regional low heat flow (Figure 1) suggests that high-temperature hydrothermal systems are not likely to exist in this area.

Basin and Range

Young volcanic areas along the northern, eastern, and western margins of the Basin and Range Province, and the ~~Butte~~ Battle Mountain heat flow high in northern Nevada (Figure 1) are the most favorable sites for the discovery of electric-quality hydrothermal systems in this region. Many hydrothermal systems exist along range-bounding faults throughout this area. Garside and Schilling (1979) list 298 thermal areas in Nevada alone; many more exist in California, Oregon, Utah, and Arizona. Although some of these are very hot, such as at Beowawe, electric-quality resources have not yet been demonstrated outside the area of high heat flow (Figure 1).

Smith and Shaw (1975) suggest that plutonic equivalents of rhyolites less than one ~~m.y.~~ million years old provide the heat for most electric-grade resources in the eastern Basin and Range. Unfortunately, the existence of such hot intrusions has yet to be established. Further, young rhyolites are not always associated with the higher-temperature resources already identified in the central and western Basin and Range of Nevada. These observations ~~lead us and other workers to conclude~~ has led to the conclusion that the key ingredient of many Basin and Range hydrothermal systems is deep circulation (greater than 3 km) along major structures in a region of high heat flow where the crust is thin (less than 25 km). Siliceous melts may or may not be present at depth in proximity to these resources.

The Roosevelt Hot Springs system in Utah has produced small quantities of electricity and further development is planned. This system is discussed in more detail in sections below. High-temperature resources at Beowawe, Steamboat Springs, and Dixie Valley in Nevada and the Alvad desert of Oregon, are other hydrothermal systems that may be suitable for electric power generation.

Most of the resources in the Basin and Range are fault controlled; identification of zones of hydrothermal flow along faults thus becomes a prime task of the regional explorationist (Ward et al., 1981).

Plateaus

Isolated resources suitable for direct heat applications exist in the Plateaus terrain of Oregon and Washington. Unlike the large, young volcanos that serve as resource targets in Alaska, Hawaii, and the Cascades, most of the hydrothermal systems of the Plateaus will probably be discovered using target models that call for either deep circulation of water along faults or deep circulation along volcanic and interbedded sedimentary ~~horizons~~ deposits. Electric-grade resources have not been documented from this terrain.

Snake River Plain

Zones of higher heat flow along the margins of the Snake River plain, and broader areas of resources in the western plain, form the most attractive resource targets (Mitchell et al., 1980). Electric development of a 150°C resource has occurred ~~is proceeding~~ at the Raft River, Idaho site and similar temperatures, without sufficient production, have been encountered in drill holes in both the eastern (Prestwich and Mink, 1979) and western (Austin, 1981) plain. The Yellowstone caldera, at the eastern end of the Snake River Plain, is the single largest concentration of geothermal energy in the U.S.; development in Yellowstone is prohibited, of course ~~however~~, by its status as a National Park.

Rocky Mountains

Water heated in fault or fracture systems and flowing to the surface near drainage bottoms forms most of the known hydrothermal systems in the Rocky Mountains. Big Creek Hot Springs in Idaho and Paradise Hot Springs in Colorado are the only springs identified by Brook et al. (1979) as having potential reservoir temperatures above 150°C in this region. Systems such as Glenwood Springs in Colorado, however, have large flows and are very attractive targets for direct applications (Barrett and Pearl, 1978). Some systems, such as Thermopolis, Wyoming, apparently have water that circulates to depth in stratigraphic horizons and rises to the surface along the limbs of anticlines (Heasler, 1981). Exploration programs in the Rocky Mountains should be targeted to identify favorable deep circulation paths for water because local heat sources, such as very young volcanos, are rare.

Great Plains

Hydrothermal resources in the Great Plains are dominated by the Madison Group of Paleozoic carbonate rocks. These rocks are found predominantly in Montana, Wyoming, and the Dakotas. Temperatures are suitable for direct applications. In North Dakota, much of the water is of poor quality (Harris et al., 1980). In Nebraska, waters are found in the Cretaceous Dakota Formation which, although cooler than the deep waters in the Madison (Gosnold and Eversoll, 1981), typically have much better water quality.

Regional exploration in the Great Plains is based on a deep sedimentary basin model, where exploration is concerned with identification of appropriate stratigraphic horizons at suitable depths for the thermal gradient to provide sufficient temperatures. Areas of anomalously high heat flow within these basins can produce water several tens of degrees hotter than water elsewhere at the same depth.

Ouachita Belt

Hydrothermal resources in central Texas are principally found in Cretaceous sandstone and limestone aquifers along the buried Ouachita fold belt (Woodruff and McBride, 1979). Water heated in the thermal gradient of the earth circulates either downward from the west or upward from the deeper parts of the gulf basin to the east. Downward-circulating waters are typically fresh while upward-circulating waters are more saline. Woodruff and

McBride (1979) have identified a broad region favorable for geothermal resources. Exploration within this region is directed toward identifying suitable stratigraphic horizons at sufficient depth.

Rio Grande Rift

Non-volcanic related-hydrothermal systems in the Rio Grande rift are apparently the result of upward flow of waters from deep rift basins at groundwater constrictions (Harder et al., 1980; Morgan et al., 1981). These geothermal resources have potential for direct applications but one related to volcanic rocks is being explored for generation of electricity in the Valles Caldera in northern New Mexico (Dondanville, 1978). Also, a deep system suitable for electric generation may exist in the vicinity of Soccorro, New Mexico (Chapin et al., 1978).

Exploration for resources suitable for direct heat application should focus on identification of groundwater flow patterns in basins, particularly in areas where constrictions may cause water to flow to the surface, while exploration for resources suited to electric power generation should focus on local regions of young volcanic activity.

Colorado Plateau

Isolated, generally cool, and low-flow rate hydrothermal systems exist in the Colorado Plateau. Regional low heat flow (Figure 1) suggests that high-temperature hydrothermal systems are not likely to exist in this area.

Basin and Range

Young volcanic areas along the northern, eastern, and western margins of the Basin and Range Province, and the ~~Butte~~ Battle Mountain heat flow high in northern Nevada (Figure 1) are the most favorable sites for the discovery of electric-quality hydrothermal systems in this region. Many hydrothermal systems exist along range-bounding faults throughout this area. Garside and Schilling (1979) list 298 thermal areas in Nevada alone; many more exist in California, Oregon, Utah, and Arizona. Although some of these are very hot, such as at Beowawe, electric-quality resources have not yet been demonstrated outside the area of high heat flow (Figure 1).

Smith and Shaw (1975) suggest that plutonic equivalents of rhyolites less than one ~~m.y.~~ million years old provide the heat for most electric-grade resources in the eastern Basin and Range. Unfortunately, the existence of such hot intrusions has yet to be established. Further, young rhyolites are not always associated with the higher-temperature resources already identified in the central and western Basin and Range of Nevada. These observations ~~lead us and other workers to conclude~~ has led to the conclusion that the key ingredient of many Basin and Range hydrothermal systems is deep circulation (greater than 3 km) along major structures in a region of high heat flow where the crust is thin (less than 25 km). Siliceous melts may or may not be present at depth in proximity to these resources.

The Roosevelt Hot Springs system in Utah has produced small quantities of electricity and further development is planned. This system is discussed in more detail in sections below. High-temperature resources at Beowawe, Steamboat Springs, and Dixie Valley in Nevada and the Alvad desert of Oregon, are other hydrothermal systems that may be suitable for electric power generation.

Most of the resources in the Basin and Range are fault controlled; identification of zones of hydrothermal flow along faults thus becomes a prime task of the regional explorationist (Ward et al., 1981).

The Geysers

(See Chapter 14).

The Geysers area, California, is the premier hydrothermal producti in the world Electric power generation began in 1960; nearly 1000 megawatts are now on line. The geology of this region has been summarized in McLaughlin and Donnelly-Nolan (1981). The Geysers is a vapor-dominated system (see discussion in following sections), with production of steam from fractured metamorphic rocks.

Vapor-dominated hydrothermal systems are rare, but they form the most attractive exploration targets, since production and fluid-handling characteristics are relatively simple. Larderello, Italy, is a vapor-dominated system that has been producing electricity since 1904.

The Geysers lies on the margin of the (primarily) Pleistocene Clear Lake volcanic field (Donnelly-Nolan et al., 1981). The heat source for The Geysers system is thought to be partially molten rock related to the young volcanic activity (Iyer and others, 1981). Geophysical studies reported in McLaughlin and Donnelly-Nolan (1981) support this conclusion. Present exploration in The Geysers area is emphasizing both the expansion of development of the steam field and regional identification of surrounding hot water systems.

Imperial Valley

Electricity is being produced from hydrothermal brines in the Imperial Valley and much additional development is planned. High-temperature resources have been identified in the Salton Sea, Westmoreland, Brawley, Heber, and East Mesa KGRAs; intensive industry effort suggests that other resources are likely to be discovered. At Cerro Prieto, just across the Mexican Border, electric power has been generated from a similar geologic environment since 1973.

Brook and others (1979) estimate that slightly less than 500 quads (10^{15} BTUs) may exist in the Salton Trough area. Numerous low- and moderate-temperature wells have also been drilled (Higgins and Martin, 1980). These resources are found in fractured stratigraphic horizons (Elders, 1979). The heat in these systems is apparently derived from local volcanic activity and high heat flow derived from a shallow mantle. Regional exploration in the Imperial Valley is based on identification of geophysical anomalies and extension of known systems.

Eastern United States

Several geological environments have been identified in the East in which geothermal resources may occur. Perhaps the best known to date is on the Atlantic coastal plain, portions of which are underlain by granitic intrusions in which the decay of naturally occurring radionuclides generates heat. Sediments of the plain that typically have low thermal conductivity blanket these intrusions, thereby causing thermal anomalies (Virginia Polytechnical Institute and State University, 1980; Costain et al., 1980). To date only one deep well has been drilled (at Chrisfield, MD) to prove the existence of such thermal anomalies and to determine the producibility of associated aquifers (Svetlichny and Lambiase, 1979). It is expected that water temperatures up to 100°C might be available at depths of perhaps 1500 m over fairly large areas defined by Costain and his co-workers.

Throughout the east there is some potential for occurrence of low-temperature thermal waters in basins and in permeable aquifers such as the basal sandstones. The potential for temperatures above 100°C seems to be quite limited, based on our present understanding of regional heat flow and sediment thermal conductivities.

EXPLORATION STRATEGY FOR REGIONAL RESOURCE ASSESSMENT

By now most major companies involved in geothermal exploration in the United States have addressed regional exploration in the high heat flow regions and in the most easily explored and highly favorable areas delineated in Figures 1 and 2. The USGS has conducted intensive intra- and extra-mural programs aimed at both geothermal resource assessment and the development of exploration technology. A great deal of geological, geochemical, geophysical and hydrological information of both a general nature and a specific geothermal nature has been published over the years. State agencies from at least 25 states, operating with federal (DOE) and state funds, have conducted resource inventories and have produced maps to aid the U.S. Geological Survey in geothermal resource assessment. DOE has funded much needed development of geothermal technology; especially pertinent to this discussion is the technology of geothermal exploration and geothermal reservoir engineering. The net result of all of this effort is that much resource data and data of a more general nature but pertinent to geothermal assessment have been placed in the public domain.

We visualize that the first major effort in regional hydrothermal resource assessment should be an integration of all available data, which we refer to as the "available data base", of box 1, Figure 3. Once these data have been integrated, critical items of missing information should be identified, and the data base should be supplemented and extended as required by acquisition of a first round of supplemental data as illustrated by boxes 2 through 5 of Figure 3. Data integration (box 6) is then performed to define the more promising prospect areas within the region. Thereafter a second round of supplemental data can meaningfully be collected (boxes 7 through 10) to evaluate each project area and to provide information to be used in selection of high priority hydrothermal targets. A final data integration (box 11) follows.

Data acquisition depicted in boxes 2 through 5 and 7 through 10 should allow considerable latitude in order to facilitate exploration of the variety of hydrothermal systems described earlier. Actual techniques shown in these boxes are in the nature of a shopping list, and the applicability of each technique or of others not indicated in Figure 3 must be assessed on a specific basis by experienced exploration personnel.

Data integration and interpretation depicted in boxes 1, 6, and 11 are based on a conceptual geological model of the resource type(s) known or expected to occur in the exploration area. These models are progressively refined with acquisition and incorporation of more data. Eventually firm conceptual models of each resource type within each region should be developed (box 12) and a prioritized inventory of all prospects based on these models should be prepared. Detailed exploration and assessment techniques can then be applied to these prospects to select suitable drill test sites or to make a decision to reject the prospect.

GEOLOGICAL TECHNIQUES

Introduction

Geological techniques for regional assessment of geothermal potential are used in conjunction with geochemical and geophysical techniques to narrow the search to a prospect area of less than 100 km^2 in size. Existing data are important, particularly geologic maps, but there will usually be a requirement for field checks and fill-in work. It is important at this stage to evaluate the tectonic and intrusive history of an area, understand the deposits and alteration produced by hydrothermal systems, and understand the effects of the regional hydrologic environment.

Tectonics

High-quality geothermal systems are located in areas of active tectonism as demonstrated earlier in this paper. This includes both faulting and young extrusive activity, although young intrusive activity may not be required for the occurrence of high-temperature systems. Faulting is necessary to maintain open fractures which are required to convey meteoric fluids to depth and return them to the surface. Permeable zones along these faults are the targets for much of the geothermal exploration that occurs today. Experience has shown that fault zones are not always permeable zones of upwelling hot fluids. Indeed, many faults exhibit lateral and vertical variability from permeable to impermeable. And, within the permeable parts, zones of cold water recharge and of thermal upwelling may be closely associated. These are problems which must be resolved using thermal and hydrological techniques.

Theories of the relationships between magmatic/volcanic areas and geothermal systems have been developed by Smith and Shaw (1975). Basically they have proposed that rhyolitic systems younger than 1 million years have the potential of providing the heat for high-temperature hydrothermal systems. Thus, the presence of these young rocks serves as a regional exploration tool.

The Roosevelt Hot Springs thermal area in Utah is a fault and fracture controlled hydrothermal system associated with young volcanism (Nielson et al., 1978). The geologic map (Fig. 4) shows the relationships which would be important to observe during regional geologic reconnaissance. In Figure 4, the geothermal system is defined by the production wells and dry holes. The Opal Mound fault has served as a fluid pathway as demonstrated by the siliceous sinter deposited along it. Although the Negro Mag fault does not have such an extensive development of siliceous sinter, production wells 54-3 and 14-2 and high heat flow values indicate that it is presently a zone of thermal fluid discharge. The Quaternary rhyolites exposed to the east and southeast of the field range in age from 800,000 to 500,000 years (Lipman et al., 1978). These rocks probably represent extrusive equivalents of an inferred granitic pluton which is sub-solidus and the heat source for the geothermal system.

Stratigraphy

Portions of stratigraphic sequences serve as geothermal reservoirs, as impermeable caps which confine the thermal system, and as cold water aquifers which may serve to mask the geothermal target. The controlling influences of the stratigraphy should be evaluated in the initial stages of an exploration program.

Large, stratigraphically controlled geothermal reservoirs are located in sedimentary basins in areas of normal and elevated heat flow. These resources are generally considered to be in the low- to intermediate-temperature range with the Paris Basin in France and the Pannonian Basin in Hungary being important examples which have been studied intensively. Ottlik et al. (1981) have studied the aquifer rocks of the Pannonian Basin and have found the aquifers to be in both carbonate and clastic rocks. The permeability in the carbonates is largely secondary with tectonic fracturing producing the permeability of some competent units while a solution porosity was developed in other units during a period of subaerial weathering.

In high-temperature environments, permeability is often considered to be principally fracture controlled with fractures at times localized within particular stratigraphic horizons. However, it is evident that the flow of geothermal fluids through stratigraphic units may increase permeability through the solution of mineral and/or glass phases. This is likely to happen where fluids are hot. On the other hand, precipitation of mineral phases and decrease in permeability will occur in areas of rapid temperature decrease.

Surface Deposits of Liquid- and Vapor-Dominated Systems

The surface deposits of both liquid- and vapor-dominated geothermal systems can include hot springs, sinters, and fumeroles. Although hot spring deposits are generally composed of calcium carbonate (travertine) or silica, manganese and iron deposits may also occur.

Hot spring waters that deposit siliceous sinters nearly always have been found to contain SiO_2 concentrations of at least 240 ppm. These concentrations of silica require subsurface temperatures of at least 180°C. Because of the high solubility of amporphous silica, these fluids then must cool to about 70°C to precipitate amorphous silica. These initial amorphous precipitates are very susceptible to weathering and their preservation is dependent on protection by subsequent deposits. Once the siliceous sinters have been deposited and protected, however, they undergo polymorphic transformations to more stable species. This transformation process generally follows the sequence:

opal → christobalite → chalcedony

The sequence is well documented at Roosevelt Hot Springs, Utah, and Steamboat Springs, Nevada, and may eventually be quantified in order to allow determination of the minimum age of hot spring deposits.

Travertine deposits are characteristic of many low- to intermediate-temperature geothermal systems. Although less conspicuous than siliceous sinters, travertine deposits also occur in high-temperature thermal fields. In these systems travertine deposits are most commonly found on the margins of the field or associated with secondary reservoirs.

Acid Alteration

The surface expressions of vapor-dominated reservoirs characteristically include chloride-poor acid sulfate springs with low discharges accompanied by sodium bicarbonate/sulfate springs, fumeroles, mudpots and acid-altered ground (White et al., 1971). These features are formed by steam and other volatile gases such as hydrogen sulfide, ammonia, and carbon dioxide which discharge at the surface or condense in meteoric water. Non-volatile components such as chloride remain in the underlying boiling brine and are not enriched in the surface discharges. Chloride-rich springs typical of hot-water systems are therefore conspicuously absent over the vapor-dominated portions of the reservoir but may occur on its margins in surrounding areas of lower elevation if the reservoir is relatively shallow.

Acid sulphate springs are typically a surficial feature produced by the oxidation of hydrogen sulfide to sulfuric acid. Altered ground surrounding the acid springs and fumeroles provide striking examples of reactivity of the waters. The altered areas are typically bleached and converted to a siliceous residue containing native sulfur, cinnabar, yellow sulfate minerals, and clay minerals including kaolinite and alunite. Similar acid alteration can also be formed at depth where steam heating of groundwaters occurs. At Matsukawa, Japan, alunite, quartz and pyrite appear to have formed from 250° to 280°C fluids with a pH near 3 (Sumi, 1969). Thus, mineral assemblages in acid-altered rocks may occur at both high and low temperatures.

Regional Hydrologic Considerations

Regional hydrologic data are viewed as being important from several standpoints. First, a sufficient amount of available water is necessary to ensure the life of a geothermal reservoir. Second, regional water quality data have been shown to be useful in pinpointing buried hydrothermal systems, and third, near-surface cold aquifers are able to distort or mask altogether the thermal signatures of underlying hydrothermal systems.

The quantity of water necessary to guarantee the recharge of the system is not generally regarded as a principal exploration factor but can be a supporting factor when combined with the probable presence of heat and fractures. In addition, it is often difficult to evaluate the recharge portion of the system until extensive exploration work has been completed, and often this remains a mystery even in fully developed fields.

Even in systems which crop out at the surface to form hot springs, it is thought that a large percentage of the thermal waters are lost to the near-surface hydrologic environment. For many buried systems, all the discharged water is thought to be lost to near-surface groundwater systems. Data from Roosevelt Hot Springs, Utah (Figure 5) have demonstrated that the system can be identified by using regional water quality data published by the USGS. Certain components such as boron, chlorine, and total dissolved solids define the discharge zones of the systems. Thus the analysis of available water quality data is a powerful and inexpensive geothermal exploration tool.

In addition to aiding in the exploration effort as described in the paragraph above, regional hydrologic systems often tend to distort or obscure entirely the discharge zones of active hydrothermal systems. Studies at Cerro Prieto, Mexico have shown that the flow of groundwater from the northeast has distorted the thermal plume rising from the system. Cold water overflow reaches an extreme condition in the Cascades Province of the U.S. where it is able to mask completely the near-surface thermal manifestations of buried systems (Sammel, 1981).

GEOCHEMICAL TECHNIQUES

Introduction

Geochemical investigations frequently play an important role in the regional evaluation of geothermal resources by providing information on sites of upwelling, the temperature and quality of the resource, and the type of resource present. This information can be obtained from careful evaluation of the chemical compositions of fluids discharged from springs and fumeroles, and from the mineral and trace element distributions in the altered rocks found at the surface and in the thermal gradient and deeper test wells. Geochemical data can also prove useful in that hydrothermal alteration effects may substantially affect the geophysical response of the rocks at depth, and the interpretation as a result.

Fluid Chemistry

Fluids discharged at the surface may differ chemically from the deeper reservoir fluids as a result of changes accompanying mixing, dilution, boiling, or conductive cooling. In addition, the chemistry of the fluid may be further modified as constituents partially or completely reequilibrate with the reservoir rocks during ascent of the fluids to the surface. The actual paths taken by the fluids may be complex and the fluid chemistry may be modified by more than one process. Despite this complexity, careful evaluation of fluid chemistry frequently provides diagnostic information about the subsurface characteristics of the geothermal system. As mentioned previously, geochemical and basic hydrologic data from springs and wells are an important source of information which can be used at an early stage in the exploration program to predict the kind of fluid that will be produced. Chemical analyses of many of the hot spring systems in the U.S. are tabulated in the literature and elsewhere (for example, U.S.G.S. computer file GEOTHERM; Teshin et al., 1979) and can be supplemented at relatively low cost during reconnaissance investigations.

The geothermal fluids of explored high-temperature liquid-dominated systems are sodium chloride brines which vary greatly in composition from field to field. These solutions may be as dilute as potable water or as concentrated as the 25 weight percent solutions characterizing some of the systems in the Imperial Valley. Systems with such extreme salinities are, however, rare. Most systems currently under evaluation in the Basin and Range Province contain less than 10,000 ppm total dissolved solids.

Bicarbonate-rich waters are commonly found in low-temperature geothermal systems and in secondary reservoirs in the shallow portions and margins of high-temperature fields. The origin of bicarbonate-rich fluids found in the secondary reservoirs of high-temperature systems was discussed by Mahon et al. (1980), who concluded that the fluids form by gas and steam heating of meteoric water. The final composition of the fluids is determined by the composition and volume of the gases and ground-water and the extent of water-rock interactions.

Subsurface Temperature -- Geothermometers

An understanding of the temperatures at depth in the geothermal reservoir rocks is crucial to the development and exploitation of the resource. Temperatures can be determined directly through downhole measurements or estimated indirectly from chemical and stable isotopic (O, H, S, C) analyses of the water, steam, gas and reservoir rocks themselves. Direct and indirect methods provide, however, different information about the reservoir.

The application of indirect methods plays a critical role in regional geothermal exploration. Indirect methods based on the chemistry of the thermal fluids can provide information on deep thermal regimes that are otherwise are inaccessible to shallow and even moderate-depth thermal gradient holes. Thus, indirect methods can be used to prioritize drilling targets and, when compared with thermal measurements made in shallow gradient wells, can be used to establish depth requirements for the deeper drilling program.

The quantitative geothermometer techniques currently available require chemical or isotopic analyses of thermal waters, steam and gas from wells and springs. These techniques can be categorized as follows: major element geothermometers, mixing geothermometers, and isotope geothermometers. The underlying premise for all three categories is that temperature-dependent reactions between either the reservoir rock and fluid or evolving gases and the fluid attain equilibrium. Furthermore, it is assumed that no reequilibration occurs after the fluid leaves the reservoir (see Fournier et al., 1974; Truesdell, 1976; Fournier, 1977; Ellis, 1979 for further details).

Several major element geothermometers have been proposed and have proven to be extremely valuable in accurately estimating subsurface temperatures. An extensive review of the use of these geothermometers was recently published by Fournier (1981).

Qualitative fluid geothermometers are used extensively during preliminary chemical surveys to locate zones of upwelling and determine the distribution of thermal waters and directions of groundwater flow. Fluid constituents that have proven to be particularly useful during these surveys include the soluble elements chlorine, boron, arsenic, cesium and bromine. Ellis and Mahon (1964, 1967) showed that the solubilities of these elements are controlled mainly by diffusion and extraction processes, and that once liberated they do not form stable secondary minerals. Changes in the concentrations of these elements as the fluids migrate from depth occur mainly from dilution or boiling. The use of atomic ratios (i.e., chloride/boron) can eliminate these effects. Other fluid constituents that are frequently used as qualitative geothermometers include lithum, trace metals (antimony, zinc, copper, uranium mercury), ammonia, hydrogen sulfide, and the ratios chloride/fluoride, chloride/sulfate, sodium/calcium, sodium/magnesium and chloride/(bicarbonate+carbonate). In general, the concentrations and ratios increase with increasing temperature, reflecting changes in constituent concentrations as a result of contamination with cold surface water, interaction between the fluids and rock at depth, and steam heating of waters (Mahon, 1970).

A map of the distribution of boron and chloride in waters in the region that includes Roosevelt Hot Springs is presented in Figure 5. It illustrates the use of one of these qualitative geothermometers. The data were compiled from published analyses of well and spring waters. The distribution suggests that the Roosevelt Hot Springs area is indeed a major center of upwelling thermal fluids and that exploration activities should be directed there. Changes in the concentration of boron and chloride occur as the thermal fluids are diluted with local groundwaters. Movement of the fluids appears to be first westward and then northward. A second source of thermal fluids is located at Thermo Hot Springs in the southwestern portion of the map and is marked by boron concentrations greater than 0.5 ppm.

The ratios of gases discharged from fumeroles have also been used as qualitative geothermometers. Mahon (1970) showed that fumeroles with the lowest ratios of carbon dioxide/hydrogen sulfide, carbon dioxide/ammonia and carbon dioxide/hydrogen were the most directly connected to the deep aquifers. The concentrations of these constituents are controlled by steam-rock reactions which can rapidly deplete the hydrogen sulfide, ammonia and hydrogen in the steam. The longer the steam path to the surface, the greater these depletions are likely to be.

Trace Element Analysis

Trace element analyses of hot spring deposits and altered rocks can supplement other data and help prioritize target areas. For example, mercury and sulphur are frequently enriched in rocks and altered ground over high-temperature thermal systems, (Matlick and Buseck, 1976; Capuano and Bamford, 1978).

GEOPHYSICAL TECHNIQUES

Introduction

Geophysics typically, and appropriately, plays a major role in the exploration for and delineation of geothermal systems by: 1) the identification of thermal provinces, and 2) geologic characterization on a regional or crustal scale. Several techniques have been applied in the geologic study and problem solving phases of detailed site-specific exploration (for example, Ward et al., 1981).

Thermal Methods

Regional heat flow characteristics on a province scale have been described in an earlier section. A prudent exploration program or regional assessment utilizes the existing heat flow or thermal gradient data base compiled by government agencies and academic workers over the years. It is often cost-effective to supplement this compilation with a regional-scale thermal gradient program which includes temperature measurement on all existing wells for which access can be gained. Several papers and texts describe details and refinements of the method and the results of regional or detailed heat flow studies (Lachenbruch, 1978; Sass et al., 1971; Chapman and Pollack, 1977; Sass et al., 1980; Ryback and Muffler, 1981).

The limitations on the use of the thermal methods are generally imposed by the drilling program. The main factor is drilling cost, but environmental restrictions, land control, permitting, and time involved are other considerations. One reconnaissance method to determine near-surface temperatures is a shallow temperature survey. With a hand-held or truck-mounted power auger a large number of holes are bored to depths of 1 to 2 meters (LeShack, 1977; Olmsted, 1977). Plastic (PVC) pipe with a sealed bottom is inserted, the hole is back filled, and temperature measurements are made after the hole temperature has stabilized. The advantage of the method is that a large number of holes can be drilled to cover a fairly large area at low or moderate cost.

The use of shallow temperature surveys has been limited because of the uncertainty that these temperatures are related to the temperature distribution at depth. The principal unknowns and disturbing factors are near-surface hydrology, soil thermal properties, topographic and slope corrections, and short-term variations. At Long Valley and Coso Hot Springs areas in California, and Soda Lakes in Nevada, however, shallow temperature measurements (Olmstead, 1977; LeShack, 1977) seem to delineate the area of anomalous heat flow in a low-cost manner. In the absence of substantial surface thermal manifestations and without obvious near-surface cold-water flow, a shallow temperature survey could be the best basis on which to plan a shallow (30-200 m) thermal gradient program. There does seem to be a limited acceptance by industry of this technique (Ward et al., 1981).

Aeromagnetic Methods

Aeromagnetic data can play a major role in the regional assessment of geothermal resources. Two major areas in which the magnetic data contribute are Curie point isotherm determinations and interpretation for subsurface geologic information.

Curie point isotherm interpretations have been reported in the literature by Bhattacharyya and Leu (1975), Shuey et al. (1977), Aiken et al. (1981) and many others. These interpretations are dependent on many assumptions and limitations. It is assumed that long wavelength negative anomalies due to lithologic changes, e.g., alluvial basins in the Basin and Range, do not significantly perturb the interpretation, and that the bottom determination of a magnetized crustal block is due to temperatures above Curie point rather than to deep-seated lithologic changes. Numerous other limitations apply to the interpretational algorithms and the data themselves. Our present judgment is that a) Curie point depth anomalies have been determined with unknown accuracy in some cases, b) Curie point studies can be a regional exploration guide especially in active volcanic provinces, c) many interpreted Curie point highs may, in fact, be due to lithologic changes at depth or lateral geologic changes, and d) because the bottom of a magnetized prism is not accurately determined from magnetic data, accuracy of Curie point depth as determined by these techniques can be poor.

Aeromagnetic surveys are widely used by industry in petroleum and mineral exploration in attempting to map subsurface structure and lithologic changes. The use in geothermal exploration should closely follow that of mineral exploration, for most geothermal resources are located in active tectonic environments characterized by a broad range of volcanic and intrusive rocks and often by active structural movement. Magnetic susceptibility often varies substantially in these rock types and provides major magnetization changes which delineate geologic units. The scale of many geothermal systems is also similar to porphyry-type mineral occurrences.

Regional aeromagnetic data are often available as part of State, (Cook et al., 1975) USGS, (Zietz et al., 1976) or NURE (Tinnel and Hinze, 1981) magnetic survey programs. These data, as at the Baltazor and Carson Sink areas in Nevada, often show major structural features and aid in forming a generalized geologic model for areas otherwise covered. These regional data are generally too widely spaced and/or too high to warrant detailed quantitative model interpretation.

The locations of faults, fracture zones, intrusives, silicic domes and possibly major alteration areas (speculative) are apparent on data we have examined from the Coso Hot Springs KGRA in California, from Baltazor, Tuscarora, McCoy, and Beowawe in Nevada, from Cove Fort-Sulphurdale and Roosevelt Hot Springs, in Utah, and from a moderate-temperature prospect near Alamosa, Colorado along the northern extension of the Rio Grande Rift. Figure 6 shows a portion of the Aeromagnetic Map of Utah (Zietz et al., 1976). The Monroe Hot Springs, Chief Joseph, Cove Fort-Sulphurdale, and Roosevelt Hot Springs KGRAs are all located in close proximity to a major magnetic discontinuity which trends east-west for a distance exceeding 150 km. This trend reflects the northern margin of the Pioche-Beaver-Tushar mineral trend with many intrusive and volcanic rocks to the south, and thin volcanics overlying thick Paleozoic through Tertiary sediments and few intrusions to the north. The magnetic trend clearly indicates a major tectonic-geologic feature important to geothermal resource localization.

Mabey (1980) has reported on the use of aeromagnetic data for the Raft River area of the Snake River Plain. Bacon (1981) interprets major structural trends and fault zones from aeromagnetic data in the Cascades. Couch et al. (1981) report Curie point isotherm minima of 5 to 9 km for several areas within the Cascade Mountains area. Costain et al. (1977;1980) have used aeromagnetic data to search for radiogenic granitic rocks beneath the insulating sediments of the Atlantic coastal plain.

The general utility of the method, the applicability to numerical modeling, the low unit costs, all argue strongly for inclusion of aeromagnetic studies in the regional assessment of geothermal resources.

Gravity Methods

Regional gravity data, with station densities of 1 station per sq km to 1 station per 25 sq km, may be available as the result of USGS studies, the Department of Defense (DOD) regional data compilation, or of university or state geophysical studies. These data are often suitable for regional-scale interpretations and are often the starting point for detailed survey design rather than the basis for detailed interpretation.

The contribution from gravity data is much the same as from aeromagnetics, that is, structural and lithologic information. The location of Basin and Range faults, thickness of alluvial fill and thickness of volcanic cover are problems addressed by gravity surveys for both the mining and geothermal industry. The delineation of low-density silicic intrusives, magma chambers in the Cascades, or major structural zones of crustal significance are other applications of the method. Gravity data may also contribute to the definition of deep sedimentary basins which are a different geothermal resource type. Costain et al. (1977;1980) have made extensive use of regional gravity data in defining radioactive granitic rocks, generally expressed as negative Bouguer anomalies, beneath the Atlantic Coastal Plain.

Regional gravity data (Cook et al., 1975) provide evidence for some of the major tectonic elements present in the main geothermal province of southwestern Utah (Fig. 7). A prominent north-trending 35-50 milligal gradient links these areas, bending eastward at Cove Fort, then trending northeast along the margin of the Colorado Plateau. Using detailed gravity data, Cook et al. (1980) mapped the many faults which define the Beaver-Cove Fort graben and add substantially to the geologic model for the Cove Fort area. In a similar manner the gravity data have delineated major faults that probably control the geothermal fluid flow at Alamosa, Colorado (Mackelprang, in prep.) and at Baltazor Hot Springs in Nevada (Edquist, 1981).

Regional gravity studies and their interpretation play a major role in understanding the tectonic framework of geothermal systems in the Cascade Range. Bacon (1981) reports a contiguous zone of gravity lows west of the High Cascades in central Oregon and notes that these define major structural trends and delineate fault zones which may localize the movement of geothermal fluids. The zone of gravity lows coincides with (1) an abrupt east-to-west decrease in heat flow from High Cascades values of 100 to 40 mW/m^2, and (2) a substantial east-to-west increase in depth to the lower crustal conductor defined by magnetotelluric soundings. Couch et al. (1981) report similar interpretations. Williams and Finn (1981) have described complexities in reduction of gravity data especially important to the Cascade Province. They report that the large silicic volcanos, calderas exceeding 10 km diameter, produce gravity lows when proper densities of 2.15 to 2.35 g/cm^3 are used for the Bouguer reduction. All other volcanos produce gravity highs as a result of higher-density subvolcanic intrusive complexes.

It would appear that gravity data contribute to a regional exploration program in most geothermal environments.

Passive Seismic Methods

Passive seismic data, which can contribute to a regional geothermal assessment, include long-term historical records of major earthquake activity and microearthquake surveys. On a regional scale, areas of high seismicity, as indicated by earthquake recording networks, define active tectonic provinces which include most areas of geothermal potential in the western

United States. Unfortunately many seismic zones have little geothermal potential.

Microearthquake surveys have been completed in several geothermal areas including Coso Hot Springs and The Geysers, California; Tuscarora and McCoy, Nevada; Roosevelt Hot Springs and Cove Fort-Sulphurdale, Utah and Raft River, Idaho. Some general observations may apply to the seismic behavior of these systems. Earthquake activity is generally episodic rather than continuous. Earthquake swarms, sometimes including tens to hundreds of events over a few days, may be typical. Earthquake magnitudes are small, generally $-0.5 < M < 2.0$, with shallow focal depths generally less than 5 km. The data are interpreted in terms of P-wave delay, S-wave attenuation, and position and alignment of epicenters.

Microearthquake surveys may play a more important role in exploration for deeper, blind geothermal systems where cold water overflow masks near-surface thermal and electrical characteristics, such as the Snake River Plain and the Cascade Province.

Seismic Refraction

Seismic refraction profiles have been recorded at The Geysers, Yellowstone National Park, Roosevelt Hot Springs, and other geothermal areas. These studies may be appropriate for regional-scale structural or crustal studies (attenuation by magma chambers, etc.), but they do not have the spatial resolution or signal averaging appropriate for prospect-scale delineation. Hill et al. (1981) recently reported on a 270-km profile from Mount Hood to Crater Lake in the Cascades and presented results in terms of crustal velocity structure. These data contribute to a better understanding of regional geology and are indirectly used in geothermal exploration.

Regional seismic reflection data such as the COCORP profiles may be useful in the same sense but are rarely available.

Electrical Methods

Thermal waters become increasingly conductive with increasing salinity and with increasing temperature up to 300°C above which conductivity decreases, and the long-term interaction between thermal fluids and the subsurface environment gives rise to extensive wall rock alteration (Moskowitz and Norton, 1977). The alteration produces conductive mineral assemblages such as clays and may develop additional porosity. This environment of low-resistivity pore fluids and conductive mineral assemblages is often a good target for the electrical exploration techniques.

The magnetotelluric (MT) method is routinely used in both the reconnaissance and detailed stages of geothermal exploration. Through precise measurements of the frequency-dependent electric and magnetic field components made at the earth's surface, one may obtain information relating to the impedance distribution (i.e., electrical resistivity) to depths greater than 100 km within the earth's crust, although reliable interpretations to these depths are rarely achieved in routine contract surveys.

Ward et al. (1981) noted that MT was used in most of the Basin and Range exploration programs which they reviewed. They attribute this to its advertised great depth of exploration and a common assumption that it is able to detect the hot rock source of heat at depths on the order of tens of kilometers. Neither of these attributes is necessarily correct. Only if a carefully selected two- or three-dimensional modeling of the earth is used in interpreting the survey results may one predict accurately the distribution of resistivities at depths of several to several tens of kilometers. Predictions of resistivities at depth are limited by the influence of surficial conductors such as alluvial fill or shallow alteration zones unless these are included in the model (Wannamaker et al., 1980). In addition the conductivity of magma at elevated temperatures is strongly dependent upon the partial pressure of water (Lebedev and Khitarov, 1964) and so hot, dry partial melt is more difficult to detect by MT than hot, wet partial melt.

Stanley (1981) described a regional, 97 station MT survey for the Cascades volcanos region. In addition to generalizing the resistivity structure for 0 to 10 km depth, he interpreted a lower crustal conductor ($\rho <$ 5 ohm-m) at 10-22 km depth which he suggests may be due in part to a partial melt associated with Cascade volcanoes. Perhaps the most important application of MT in regional geothermal exploration will lie in detecting regions of partial melt in the deep crust or upper mantle (Wannamaker et al., 1980).

Electrical resistivity data are routinely acquired in geothermal exploration on the detailed, site-specific scale but are less frequently used in regional or reconnaissance exploration. Schlumberger soundings are often conducted at many scattered sites within a large region, and depth to a given conductive horizon is contoured from these data. Although the array is efficient for data acquisition, the assumption of one-dimensional environments must be evaluated, particularly as current and potential electrodes expand across structures or other lateral resistivity contrasts in complex geologic environments. Thus the results are often misleading even for a regional assessment.

The USGS and some survey contractors have promoted the bipole-dipole or roving dipole array for reconnaissance resistivity surveys. In this array current is introduced through a long (one- to two-km) transmitting dipole and voltage drops are observed at two short (0.2 to 0.5 km) orthogonal receiving dipoles two to ten km distant. The reduced resistivity values are contoured and then considered to represent large-scale resistivity variations at substantial (one to five km) depths. Although the generalization is often valid, the reduced resistivity values are strongly dependent on the local resistivity distribution in the vicinity of the transmitting dipole (Frangos and Ward, 1980). The data are difficult to interpret accurately and are, in general, only appropriate for regional-scale interpretation. In view of these complications for reconnaissance resistivity arrays, the resistivity method plays a relatively minor role in regional assessment in contrast to a key role in detailed site-specific exploration.

CONCLUSIONS

Hydrothermal resources occur worldwide in specific geologic environments. In the U.S., the bulk of known resources and all known high-temperature resources occur in the west, including Hawaii and Alaska. Although the hydrothermal potential of a given region can often be qualitatively assessed based on knowledge of the geologic environment, reconnaissance exploration is usually required to locate and assess sub-regions having higher potential for resource occurrence. Thereafter more detailed exploration is performed in these sub-regions for the purpose of locating sites to drill test wells. Geological, geochemical, geophysical and hydrological surveys and evaluations are used jointly for all of this work.

The specific techniques used in a given region are ideally selected by considering the expected causes and manifestations of hydrothermal resources in that region. This requires an understanding of the regional geology including stratigraphy and structure. For example, in the mid-continental U.S. one would use techniques capable of providing information on occurrence of aquifers at depth, areas of highest heat flow and areas of best water quality, whereas in the Basin and Range Province one would select techniques that reveal previously overlooked surface manifestation of once-active thermal springs and techniques that detect faults carrying circulating thermal waters. Compilation of available relevant data allows gaps in the data base to be identified, and these gaps are filled in by field and laboratory work.

Several successive cycles of field surveys and data integration are used with progressive refinement of understanding and elimination of areas having low potential for occurrence of a resource. An attempt is made to use techniques having lower unit cost during early stages of the regional assessment program, when a large area is being considered, and to apply those techniques that have higher unit cost to a small, carefully selected portion of the whole region.

Geological techniques are used to study the stratigraphy, structure and tectonics of a region and to locate and evaluate surface manifestations such as thermal springs and spring deposits. The objective at this stage is to identify geologic environments where the three basic elements of a geothermal resource, a source of heat, permeable rocks and water to carry the heat to the surface where it can be extracted, are present together. The geologist may thus search for volcanic rocks that are young enough to indicate a still-cooling intrusion at depth to provide heat, and for indications that thermal waters are now or recently were active in the area. Alternatively an environment may be found that has high regional heat flow and active faulting that may provide pathways for deep water circulation and subsequent heating in the geothermal gradient. In any case, it is required that the geologist be experienced enough to recognize potential hydrothermal environments and to recognize in the field even subtle indications of present or former geothermal activity.

Hydrologic considerations can be important at this stage both in assessing potential of specific areas and in helping to guide other exploration work. For example, flow of cold water in near-surface aquifers invalidates use of heat flow and thermal gradient studies in holes that do not penetrate the cold aquifers.

Geochemistry finds important application to regional hydrothermal assessment. Traditionally in geothermal work geochemistry as a term has meant fluid geochemistry, mostly because of excellent, pioneering work by the U.S.

Geological Survey and others. In recent years it has become evident that geochemical study of rocks and of hydrothermal alteration products is equally important.

Upwelling thermal water can sometimes be detected by rather simple studies of the chemistry of springs and wells. Often these chemical data are available and, whenever they are not, they are relatively fast and inexpensive to collect. Potential chemical indications of hydrothermal waters include anomalous boron or chloride concentrations in ground water. Another important application of fluid geochemistry is in predicting reservoir temperatures. Although chemical geothermometry is based on many seemingly limiting assumptions, it has proven successful in predicting subsurface temperature in most high-temperature systems. Application to resources whose temperature is below, perhaps, 130°C is questionable at present. New work in application of light-stable isotope studies to geochemistry shows promise but needs further development.

Once a potential resource area is located, certain predictions about the reservoir fluids are possible through geochemistry. For example, chloride-rich waters are usually associated with water-dominated hydrothermal systems whereas acid-sulfate waters often occur in association with vapor-dominated systems or portions of systems. In addition, it is practical to study trace element distribution in rocks and hydrothermal alteration products over specific areas but probably not over the whole region under assessment. Mercury, sulfur, antimony, arsenic, gold, silver and thallium are often concentrated in rocks and/or soil over high-temperature geothermal systems.

Geophysical methods can be used directly to detect a hydrothermal resource through heat flow and downhole temperature surveys and also to contribute to a better geological picture of the reconnaissance area and of selected subareas of high resource potential. It is important to remember that one is looking not for high heat flow per se but for anomalous temperature. Temperature and, if possible, heat flow measurements should be made in available wells in the region, especially those within subareas of expected high resource potential. Great care must be exercised in interpreting these data, and, if cold water overflow is suspected, an area should not be casually downgraded simply because of low temperature or heat flow values. Data from available wells is usually fairly inexpensive but collection of gradient and heat flow information can become very expensive when new drilling is required. Techniques under development for using temperature measurements made in shallow (3 m) holes promise to cut costs substantially in specific areas where this technique may apply.

Aeromagnetic and gravity methods are primarily valuable in helping to determine regional and detailed structure, locating deeply buried radiogenic granites, and in extending geologic information into areas of little or no outcrop. Gravity surveys have been used to detect silicified zones over hydrothermal systems in the less dense sediments of the Imperial Valley, California, but this use probably has restricted application. Curie point isotherm interpretations may make a major contribution to regional evaluation in certain geological provinces.

Electrical geophysical methods are often used to detect the greater electrical conductivity, or lower resistivity, often associated with hydrothermal systems due to elevated temperature, fracturing and deposition of clay and other minerals through alteration of reservoir rocks by thermal fluids. Magnetotelluric (MT) surveys are also used to attempt to detect hot rocks or magma at depth. Some recent studies (Wannamaker et al., 1980) indicate that MT has the potential to detect partial melts in the lower crust or upper mantle. To do so, however, requires an extensive MT data base throughout the region. Galvanic resistivity is not often used in a reconnaissance mode, but has proved to be very helpful in evaluation of specific prospect sites.

Experience with regional hydrothermal assessment will convince even the most optimistic that a great deal of improvement is needed in many techniques. Geothermal exploration has not had the benefit of the decades of development and refinement in techniques common to petroleum and mining exploration. Wildcat well success ratios are very low in hydrothermal exploration, and this has had a negative impact on the economics of hydrothermal development for a fledgling industry struggling with economics for even the highest grade resources.

REFERENCES

Aiken, C. L. V., Hong, M. R., and Peeples, W. J., 1981, Aeromagnetic anomaly inversion and analysis of the depth-to-Curie isotherm: abs., 51st Annual International Meeting, Soc. Expl. Geophys., October 11-15, 1981, in Los Angeles.

Austin, John, 1981, Direct utilization of geothermal energy for food processing at Ore-Ida Foods, Inc., in Geothermal Direct Heat Applications Program Summary: U. S. Department of Energy, Idaho Falls, ID, p. 29-37.

Bacon, C. R., 1981, Geology and geophysics of the Cascade Range: abs., 51st Annual International Meeting, Soc. Expl. Geophys., October 11-15, 1981, in Los Angeles.

Barrett, J. K., and Pearl, R. H., 1978, An appraisal of Colorado's geothermal resources: Colorado Geological Survey Bull. 39, 224 p.

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

Bloomquist, R. G., 1981, Geothermal energy policy in Washington--an overview: Geothermal Resources Council Special Report 10, p. 65-67.

Bowen, R. G., 1981, Mt. Hood exploration, Oregon--a case history: Geothermal Resources Council Special Report 10, p. 21-23.

Brook, C. A., Mariner, R. H., Mabey, D. R., Swanson, J. R., Goffanti, M., and Muffler, L. J. P., 1979, Hydrothermal convection systems with reservoir temperatures $\geq$ 90°C, in Muffler, L. J. P., ed., Assessment of Geothermal Resources of the United States-1978: U. S. Geological Survey Circular 790, p. 18-85.

Capuano, R. M., and Bamford, R. W., 1978, Initial investigation of soil mercury geochemistry as an aid to drill site selection in geothermal systems: University of Utah Research Institute, Earth Science Laboratory Report 13, 32 p.

Chapin, C. E., Chamberlin, R. M., Osborn, G. R., White, D. W., and Santad, A. R., 1978, Exploration framework of the Socorro geothermal area, in Field Guide to Selected Cauldrons and Mining Districts of the Datil-Moyollon Volcanic Field, New Mexico: New Mexico Geological Society Special Publication 7, p. 115-129.

Chapman, D. S., and Pollack, H. N., 1977, Regional geotherms and lithospheric thickness: Geology, v. 5, p. 265-268.

Chen, B. H., Lopez, L. P., Kuwada, J. T., and Farrington, R. J., 1980, Progress report on HGP-A wellhead generator feasibility project: Geothermal Resources Council Transactions, v. 4, p. 491-494.

Cook, K. L., Montgomery, J. R., Smith, J. T., and Gray, E. F., 1975, Simple Bouguer gravity anomaly map of Utah: Utah Geological and Mineral Survey, Map 37.

Cook, K. L., Serpa, L. F., and Pe, W., 1980, Detailed gravity and aeromagnetic surveys of the Cove Fort-Sulphurdale KGRA and vicinity, Millard and Beaver Counties, Utah: Univ. of Utah Dept. Geol. and Geophys. Report 78-1701.a.5.2., DOE contract no. DE-AC07-78ET28392.

Costain, J. K., Glover, L., and Siuha, A. K., 1980, Low-temperature geothermal resources in the eastern United States: EOS Trans., Amer. Geophys. Union, v. 61, no. 1.

Costain, J. K., Glover, L. III, and Sinha, A. K., 1980, Low temperature geotheraml resources in the Eastern United States: EOS, v. 61, n. 1, p. 1-13.

Costain, J. K., Glover, L. III, and Sinha, A. K., 1977, Evaluation and targeting of geothermal energy resources in the southeastern United States, progress reports: VPI & SU-5648, contract ET-78-C-05-5648, Dept. of Energy, Washington, D.C.

Couch, R., Gemperle, M., Connard, G., and Pitts, G. S., 1981, Structural and thermal implications of gravity and aeromagnetic measurements made in the Cascade volcanic arc: abs., 51st Annual International Meeting, Soc. Expl. Geophys., October 11-15, 1981, in Los Angeles.

Dondanville, R. F., 1978, Geologic characteristics of the Valles Caldera geothermal system, New Mexico: Geothermal Resources Council Transactions, vol. 2, p. 157-160.

Donnelly-Nolan, J. M., Hearn, B. C. Jr., Curtis, G. H., and Drake, R. E., 1981, Geochronology and evolution of the Clear Lake Volcanics, in R. J. McLaughlin and J. M. Donnelly-Nolan, eds., Research in The Geysers-Clear Lake Geothermal Area, Northern California: U. S. Geological Survey Professional Paper 1141, p. 47-60.

Edquist, R. K., 1981, Geophysical investigations of the Baltazor Hot Springs known geothermal resource area and the Painted Hills thermal area, Humboldt County, Nevada: University of Utah Research Insitute, Earth Science Lab Report No. 54, 89 p.

Ehni, W. J., 1981, Summary of 1980 geothermal drilling -- western United States: Geothermal Energy, v. 9, no. 8, p. 4-150.

Elders, W. A., ed., 1979, Geology and geothermics of the Salton trough: Geological Society of America, 92nd Annual Meeting, Field Trip No. 7, 109 p.

Ellis, A. J., 1979, Chemical geothermometry in geothermal systems: Chemical Geology, v. 25, p. 219-226.

Ellis, A. J., and Mahon, W. A. J., 1964, Natural hydrothermal systems and experimental hot water/rock interactions: Geochem. et Cosmochim. Acta, v. 28, p. 1323-1357.

Ellis, A. J., and Mahon, W. A. J., 1967, Natural hydrothermal systems and experimental hot water/rock interactions, Part 2: Geochem. et Cosmochim. Acta, v. 31, p. 519-538.

Fairbanks, B. D., Openshaw, R. E., Souther, J. G., and Stauder, J. J., 1981, Meager Creek Geothermal Project - an exploration case history: Geothermal Resources Council Bulletin, v. 10, no. 6, p. 3-7.

Fournier, R. O., 1981, Application of water geochemistry to geothermal exploration and reservoir engineering, in Rybach, L. and Muffler, L.J.P. (eds.), Geothermal Systems, Principles and Case Histories: New York, Wiley & Sons, p. 109-143.

Fournier, R. O., 1977, Chemical geothermometers and mixing models for geothermal systems, in Proceedings of the International Atomic Energy Agency advisory group on the application of nuclear techniques to geothermal studies, Pisa, 1975: Geothermics, Special Issue 5, p. 41-50.

Fournier, R. O., White, D. E. and Truesdell, A. H., 1974, Geochemical indicators of subsurface temperature-Part I, Basic assumptions: Journal of Research, U. S. Geological Survey, v. 2, p. 259-262.

Frangos, W., and Ward, S. H., 1980, Bipole-dipole survey at Roosevelt Hot Springs KGRA, Beaver County, Utah: University of Utah Research Institute, Earth Science Laboratory report no. ESL-43, 51 p.

Furumoto, A. S., 1978, The relationship of a geothermal reservoir to the geological structure of the east rift of Kilauea volcano, Hawaii: Geothermal Resources Council Transactions, v. 2, p. 199-201.

Garside, L. J., and Schilling, J. H., 1979, Thermal Waters of Nevada: Nevada Bureau of Mines and Geology, Bulletin 91, 163 p.

Gosnold, W. D., and Eversoll, D. A., 1981, Usefulness of heat flow data in regional assessment of low temperature geothermal resources with special reference to Nebraska: Geothermal Resources Council Transactions, vol. 5, p. 79-82.

Harder, V., Morgan, P., and Swanberg, C. A., 1980, Geothermal resources in the Rio Grande rift, Origins and potential: Geothermal Resources Council Transactions, vol. 4, p. 61-64.

Harris, K. L., Winczewski, L. M., Umphrey, H. R., and Anderson, S. B., 1980, An evaluation of hydrothermal resources of North Dakota: Univ. North Dakota Engineering Equipment Station Bull. 80-03-EES-02, 82 p.

Heasler, H. P., 1981, Conductive thermal modeling of Wyoming geothermal systems: in C. A. Ruscetta, D. Foley, eds., Geothermal Direct Heat Program Glenwood Springs Technical Conference Proceedings: Earth Science Laboratory, Univ. Utah Research Institute, ESL-59, p. 301-313.

Higgins, C. T., and Martin, R. C., compilers, 1980, Geothermal Resources of California: California Division of Mines and Geology, Geologic Data Map Series, Map No. 4.

Hill, D. P., Mooney, W. D., Fuis, G. W., and Healy, J. H., 1981, Evidence on the structure and tectonic movements of the volcanoes in the Cascade Range, Oregon and Washington from seismic refraction/reflection measurements: abs. 51st Annual International Meeting, Soc. Expl. Geophys., October 11-15, 1981, in Los Angeles.

Holland, H. D. and Malinin, S. D., 1979, The solubility and occurrence of non-ore minerals, in Barnes, H. L. (ed.), Geochemistry of hydrothermal ore deposits, 2nd edition: New York, Wiley, p. 461-508.

Iyer, H. M., Oppenheimer, D. H., Hitchcock, Tim, Roloff, J. N., and Coakley, J. M., 1981, Large teleseismic P-wave delays in the Geysers-Clear Lake geothermal area, in R. J. McLaughlin, J. M. Donnelly-Nolan, eds., Research in the Geysers-Clear Lake Geothermal Area, Northern California: U. S. Geological Survey Professional Paper 1141, p. 97-116.

Lachenbrach, A. H., 1978, Heat flow in the Basin and Range Province and thermal effects of tectonic extension: Pure and Appl. Geophys., v. 117, p. 34-50.

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

LeShack, L. A., 1977, Rapid reconnaissance of geothermal prospects using shallow temperature surveys: Development and Resources Transportation Co., Rept. DOE contract EG-77-C-01-4021, Silver Springs, MD.

Lipman, P. W., Rowley, P. D., Mehnert, H. H., Evans, S. H. Jr., Nash, W. P., and Brown, F. H., 1978, Pleistocene rhyolite of the Mineral Mountains, Utah - geothermal and archeological significance: J. Research U. S. Geol. Survey, v. 6, p. 133-147.

Mabey, D. R., Hoover, D. B., O'Donnell, J. E., and Wilson, C. W., 1978, Reconnaissance geophysical studies of the geothermal system in southern Raft River Valley, Idaho: Geophys., v. 43, no. 7, p. 1470-1484.

Mackelprang, C. E., (1982), Results of a detailed gravity survey in the Alamosa area, Alamosa County, Colorado: University of Utah Research Institute, Earth Science Laboratory Report, (in prep).

Mahon, W. A. J., McDowell, G. D., and Finlayson, J. B., 1980, Carbon dioxide, its role in geothermal systems: New Zealand Journal of Science, v. 23, p. 133-148.

Mahon, W. A. J., 1970, Chemistry in the exploration and exploitation of hydrothermal systems, United Nations Symposium on the Development and Utilization of Geothermal Resources, Pisa, 1970: Geothermics Special Issue 2, p. 1310-1322.

Matlick, J. S. III, and Buseck, P. R., 1976, Exploration for geothermal areas using mercury: a new geochemical technique, in C. Pezzotti, ed., Proc. 2nd U.N. Symposium on Development and Use of Geothermal Resources, v. 1: U. S. Govt. Printing Office, p. 785-792.

McLaughlin, R. J. and Donnelly-Nolan, J. M., eds., 1981, Research in the Geysers-Clear Lake Geothermal Area, Northern California: U. S. Geological Survey Professional Paper 1141, 259 p.

Mitchell, J. C., Johnson, L .L., and Anderson, J. E., 1980, Potential for direct heat application of geothermal resources: Idaho Dept. of Water Resources, Water Information Bull. No. 30, Part 9, 396 p.

Morgan, Paul, Harder, Vicki, Swanberg, C. A., and Daggett, P. H., 1981, A groundwater convection model for Rio Grande Rift geothermal resources: Geothermal Resources Council Transactions, vol. 5, p. 193-196.

Moskowitz, B. and Norton, D., 1977, A preliminary analysis of intrinsic fluid and rock resistivity in active hydrothermal systems: Jour. Geophys., Res. v. 82, n. 36, p. 5787-5795.

Motyka, R. J., Moorman, M. A., and Reeder, J. W., 1980, Assessment of thermal springs sites in southern southeastern Alaska--preliminary results and evaluation: Alaska Div. Geological and Geophysical Surveys Open File Report 127, 72 p.

Motyka, R. J. and Moorman, M. A., 1981, Reconnaissance of thermal spring sites in the Aleutian Arc, Atka Island to Becherot Lake: Geothermal Resources Council Transactions, v. 5, p. 111-114.

Muffler, L. J. P., 1979, Assessment of geothermal resources of the United States-1978: U. S. Geological Survey Circular 790, 63 p.

Nielson, D. L., Sibbett, B. S., McKinney, D. B., Hulen, J. B., Moore, J. N., and Samberg, S. M., 1978, Geology of Roosevelt Hot Springs KGRA, Beaver County, Utah: Univ. of Utah Research Institute, Earth Science Lab. Rept. No. 12, 121 p.

Olmsted, F. H., 1977, Use of temperature surveys at a depth of 1 meter in geothermal exploration in Nevada: U. S. Geol. Survey Prof. Paper, 1044-B, 25 p.

Ottlik, P., Galfi, J., Horvath, F., Korim, K., and Stegena, L., 1981, The low enthalpy geothermal resource of the Pannonian Basin, Hungary, in Rybach, L. and Muffler, L. J. P. (eds.), Geothermal Systems, Principles and Case Histories: New York, Wiley & Sons, p. 221-245.

Press, Frank, and Siever, Raymond, 1974, Earth: San Francisco, W. H. Freeman and Co., 945 p.

Prestwich, S. M. and Mink, L. L., 1979, Snake River Plain, Idaho geothermal exploration well: Geothermal Resources Council Transactions, v. 3, p. 549-552.

Renner, J. L. and Vaught, T. L., 1979, Geothermal resources of the eastern United States: Arlington, Gruy Federal, Report DOE/NVO/1558-7, 59 p.

Riccio, J. F., ed., 1979, Geothermal Resource Assessment of Mount Hood: Oregon Dept. Geology and Mineral Industries Open-File Report 0-79-8, 273 p.

Ross, H. P., Nielson, D. L., and Moore, J. N., (1982), An exploration case study of the Roosevelt Hot Springs geothermal system, Utah: American Association of Petroleum Geologists Bulletin, (in press).

Rybach and Muffler, L. J. P., eds., 1981, Geothermal Systems, Principles and Case Histories: New York, Wiley & Sons, 359 p.

Sammel, E. A., 1981, Results of test drilling at Newberry volcano, Oregon: Geothermal Resources Council Bulletin, vol. 10, no. 11, p. 3-8.

Sass, J. H., Blackwell, D. D., Chapman, D. S., Costain, J. K., Decker, E. R., Lawver, L. A., and Swanberg, C. A., 1981, Heat flow from the crust of the United States, in Y. S. Tourlovkian, W. R. Judd, R. F. Roy, eds., Physical Properties of Rocks and Minerals: New York, McGraw-Hill, p. 503-548.

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

Shuey, R. T. et al., 1977, Curie depth determination from aeromagnetic data: Royal Astronomical Society Geophys. Jour., v. 50, p. 75-101.

Simmons, Gene, and Roy, R. F., 1969, Heat flow in North America, in P. J. Hart, ed., The Earth's Crust and Upper Mantle: American Geophysical Union Monograph 13, p. 78-81.

Smith, R. L., and Shaw, H. R., 1975, Igneous-related geothermal system, in Assessment of geothermal resources of the United States, 1975: U. S. Geol. Survey Circ. 726, p. 58-83.

Stanley, W. D., 1981, Magnetotelluric survey of the Cascade volcanoes region, Pacific Northwest: abs., 51st Annual International Meeting, Soc. Expl. Geophys., October 11-15, 1981, in Los Angeles.

Sumi, K., 1969, Zonal distributions of clay minerals in the Matsukawa geothermal area, Japan: Proceedings, International Clay Conference, Tokyo, p. 501-512.

Svetlichny, M., and Lambiase, J. J., 1979, Coastal plain stratigraphy at DGT-1, Chrisfield, Maryland, in J. K. Costain and L. Glover III, eds., Evolution and Targeting of Geothermal Resources in the southeastern United States: Virginia Poly. Inst. and State Univ. Progress Report 78ET27001-7, 134 p.

Teshin, V. N., Swanson, J. R., and Orris, G. J., 1979, GEOTHERM-Geothermal Resources file: Trans. Geoth. Resources Council, v. 3, p. 721-724.

Thomas, D. M., Cox, M. E., Kavahikava, P., and Mattice, M. D., 1980, Hawaii geothermal resource assessment program, direct heat resource assessment, Phase II: Hawaii Institute of Geophysics Report DOE/ET/27023-4, 80 p.

Tinnel, E. P., and Hinze, W. J., 1981, Preparation of magnetic anomaly profile and contour maps from DOE-NURE aerial survey data, Vol. I, Processing procedures: Union Carbide Corp. Nuclear Division report ORNL/CSD/TM-155, 131 p.

Truesdell, A. H., 1976, Summary of section III; Geochemical techniques in exploration, in Proceedings of the 2nd United Nations symposium on the development and use of geothermal resources, p. liii-lxxiv.

Turner, D. L., Forbes, R. B., Albanese, M., Macbeth, J., Lockhart, A. B., and Seed, S. M., 1980, Geothermal energy resources of Alaska: Univ. Alaska Geophysical Institute Rept. UAG R-279, 19 p.

Virginia Polytechnic Institute and State University, 1980, Geothermal energy for the eastern U.S.: Bull. Geoth. Resources Council, v. 10, no. 10, p. 3-12.

Vozoff, K., 1972, The magnetotelluric method in the exploration of sedimentary basins: Geophysics, v. 37, no. 1, p. 38-141.

Wannamaker, P. E., Ward, S. H., Hohmann, G. W., and Sill, W. R., 1980, Magnetotelluric models of the Roosevelt Hot Springs thermal area, Utah: University of Utah Department Geological and Geophysics Tech. Report, DOE/ET/27002-8, 213 p.

Ward, S. H., Ross, H. P., and Nielson, D. L., 1981, Exploration strategy for high-temperature hydrothermal systems in the Basin and Range Province: Am. Assoc. Petroleum Geologists Bull., v. 65/1, p. 86-102.

White, D. E., Muffler, L. J. P., and Truesdell, A. H., 1971, Vapor dominated hydrothermal systems compared with hot water systems: Econ. Geology, v. 66, p. 75-97.

White, D. E., and Williams, D. L., 1975, Summary and conclusions, in Assessment of Geothermal Resources of the United States-1975: U. S. Geological Survey Circular 726, 155 p.

Williams, D. L., and Finn, C., 1981, Evidence from gravity data on the location and size of subvolcanic intrusions, preliminary results: 51st Annual International Meeting, Soc. Expl. Geophys., October 11-15, 1981, in Los Angeles.

Woodruff, C. M., and McBride, M. W., 1979, Regional assessment of geothermal potential along the Balcones and Luling-Mexia-Tulco fault zones, Central Texas: Texas Bureau of Economic Geology Report DOE/ET/28375-1, 145 p.

Youngquist, Walter, 1981, Geothermal potential of the Cascades: Geothermal Resources Council Special Report 10, p. 25-29.

Zietz, I., Shuey, R., and Kirby, J. R., Jr., 1976, Aeromagnetic map of Utah: U.S. Geological Survey Map GO-907.

FIGURE 1
HEAT FLOW IN THE U.S.

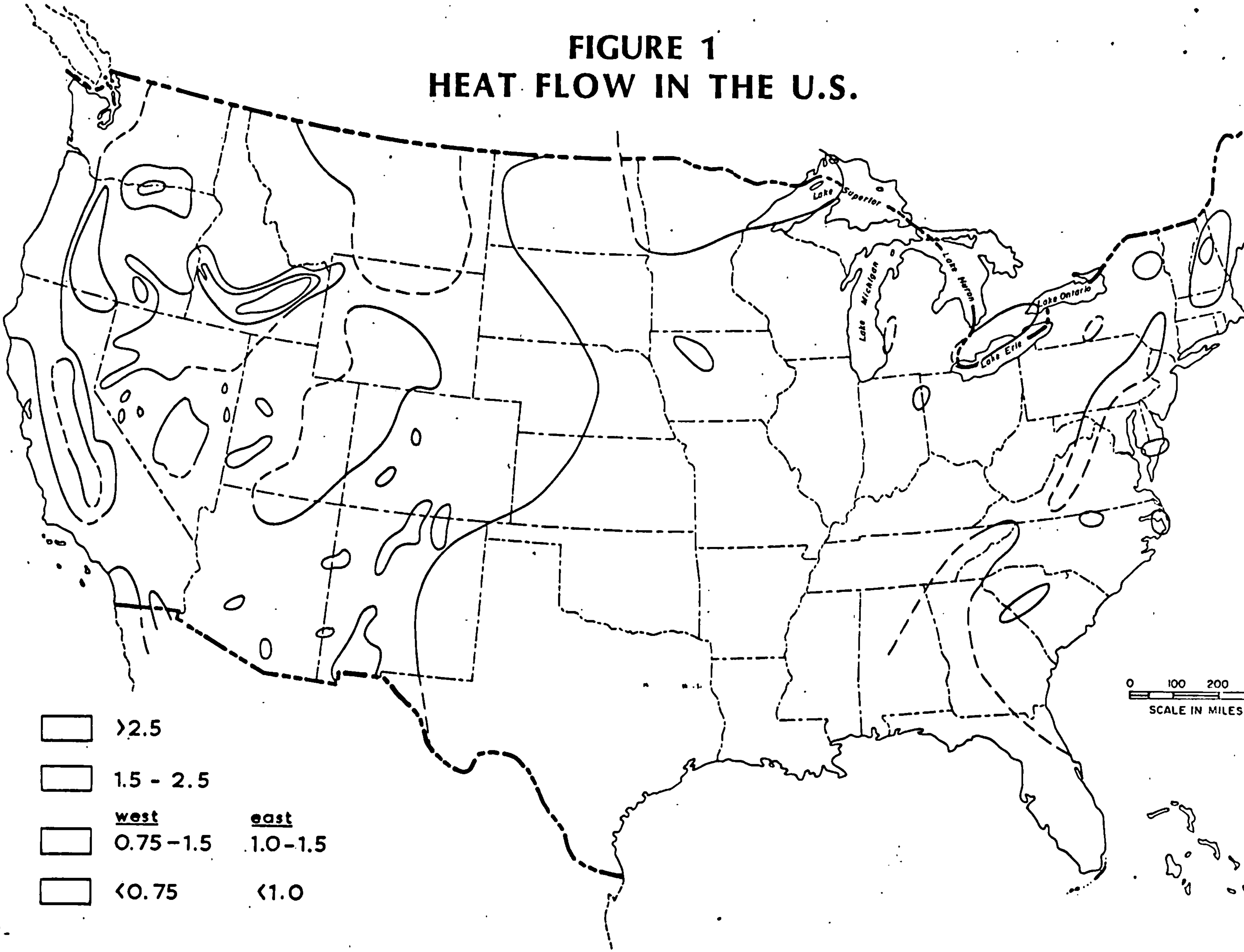
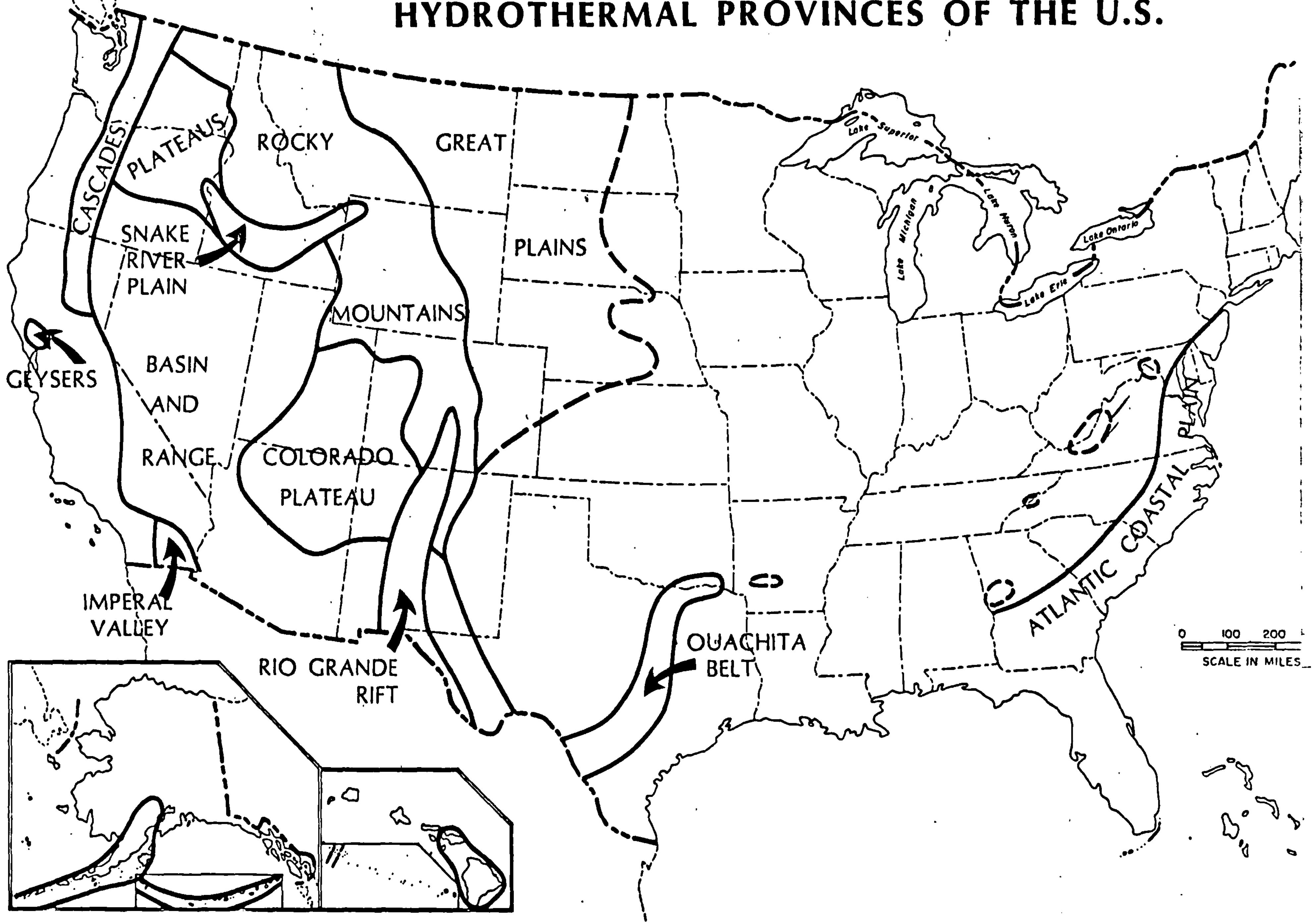


FIGURE 2
HYDROTHERMAL PROVINCES OF THE U.S.



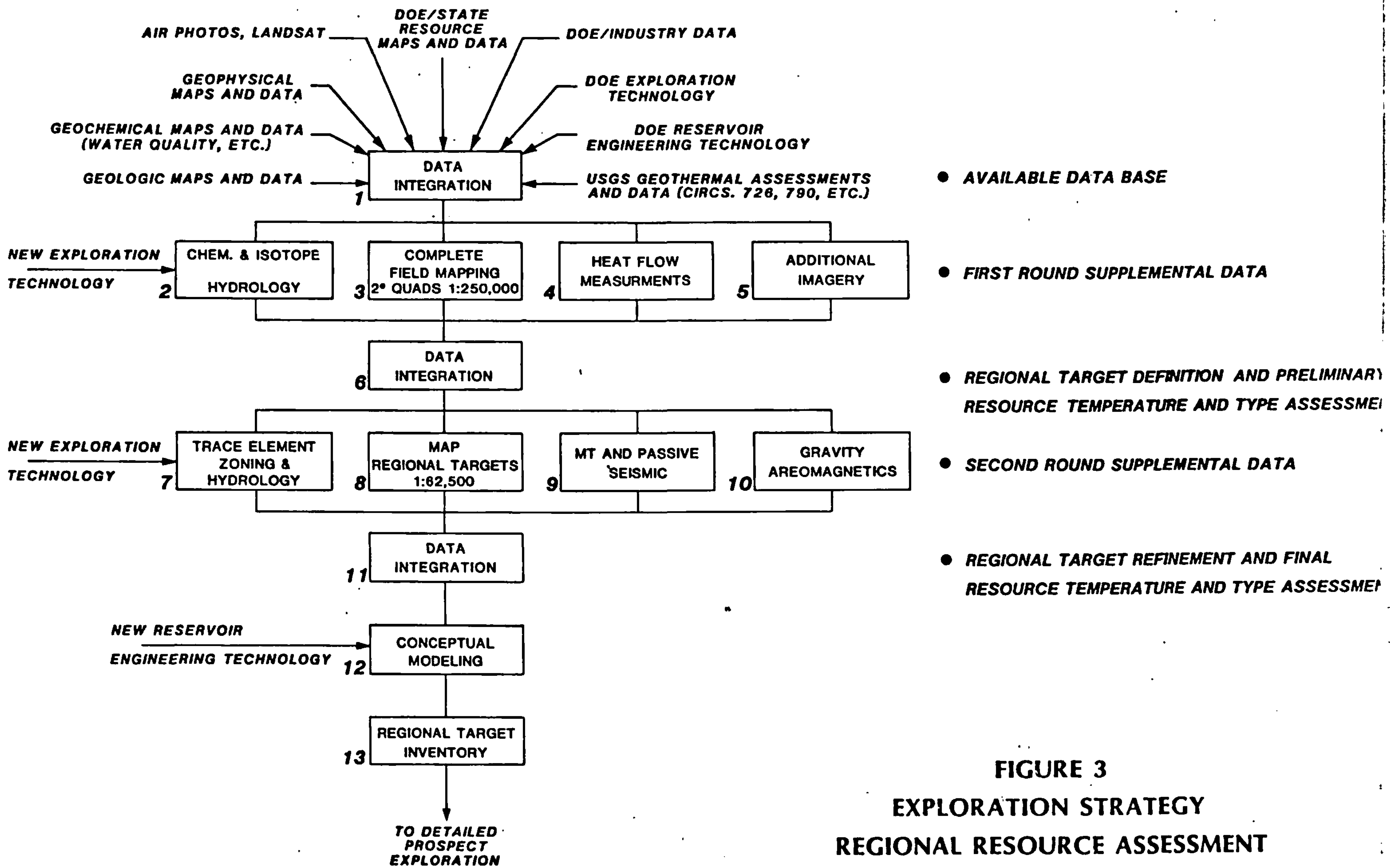


FIGURE 3
EXPLORATION STRATEGY
REGIONAL RESOURCE ASSESSMENT

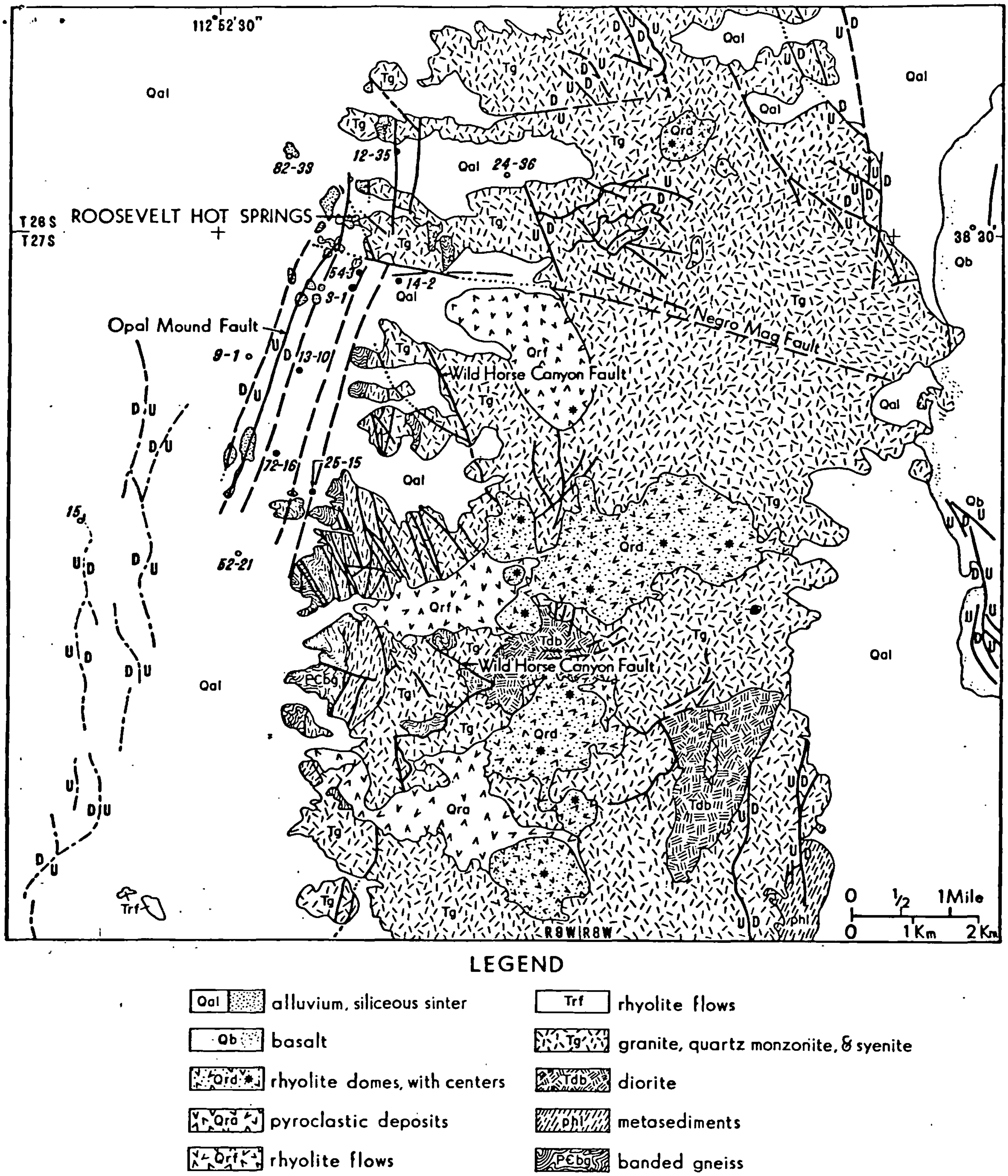


Figure 4. Geologic map of the Roosevelt Hot Springs geothermal area and the adjacent Mineral Mountains. Closed circles indicate producing geothermal wells and dry holes are shown by the open circles. (Ross et al., 1982)

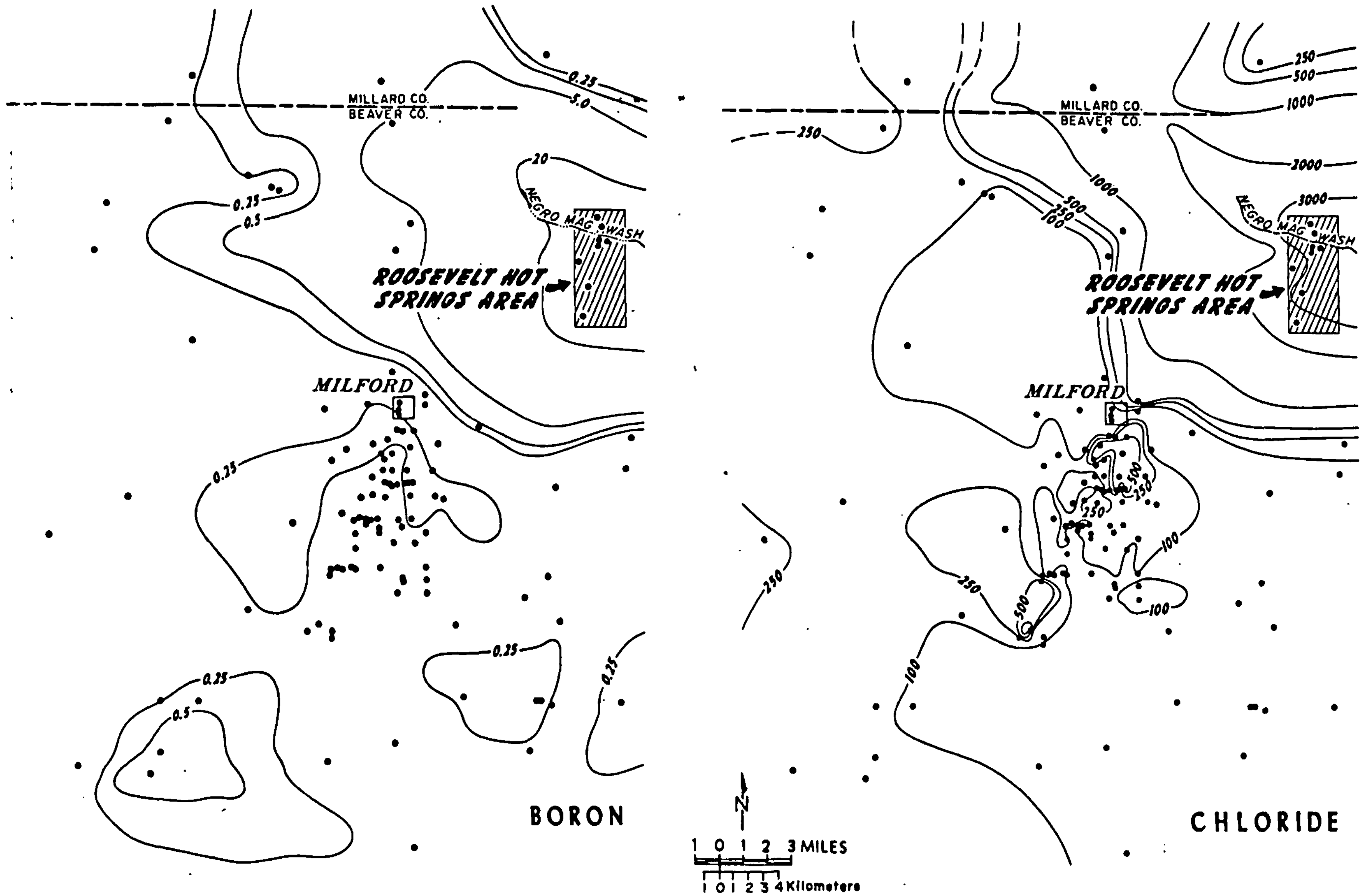


Figure 5 Boron and cloride in wells and sqrings in the Milford, Utah area.

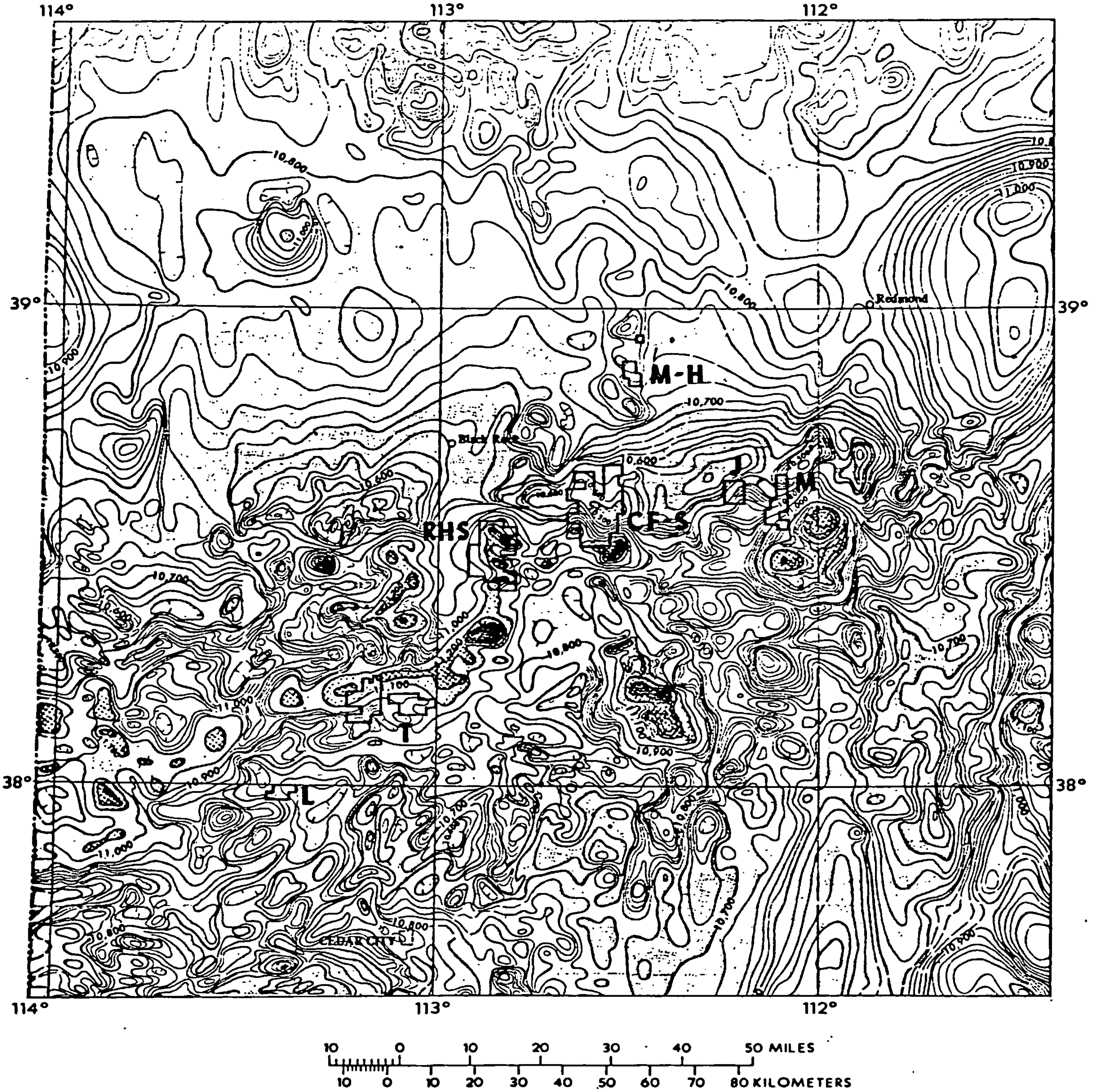


Figure 6. A portion of the Aeromagnetic Map of Utah
(after Zietz, Shuey, Kirby)

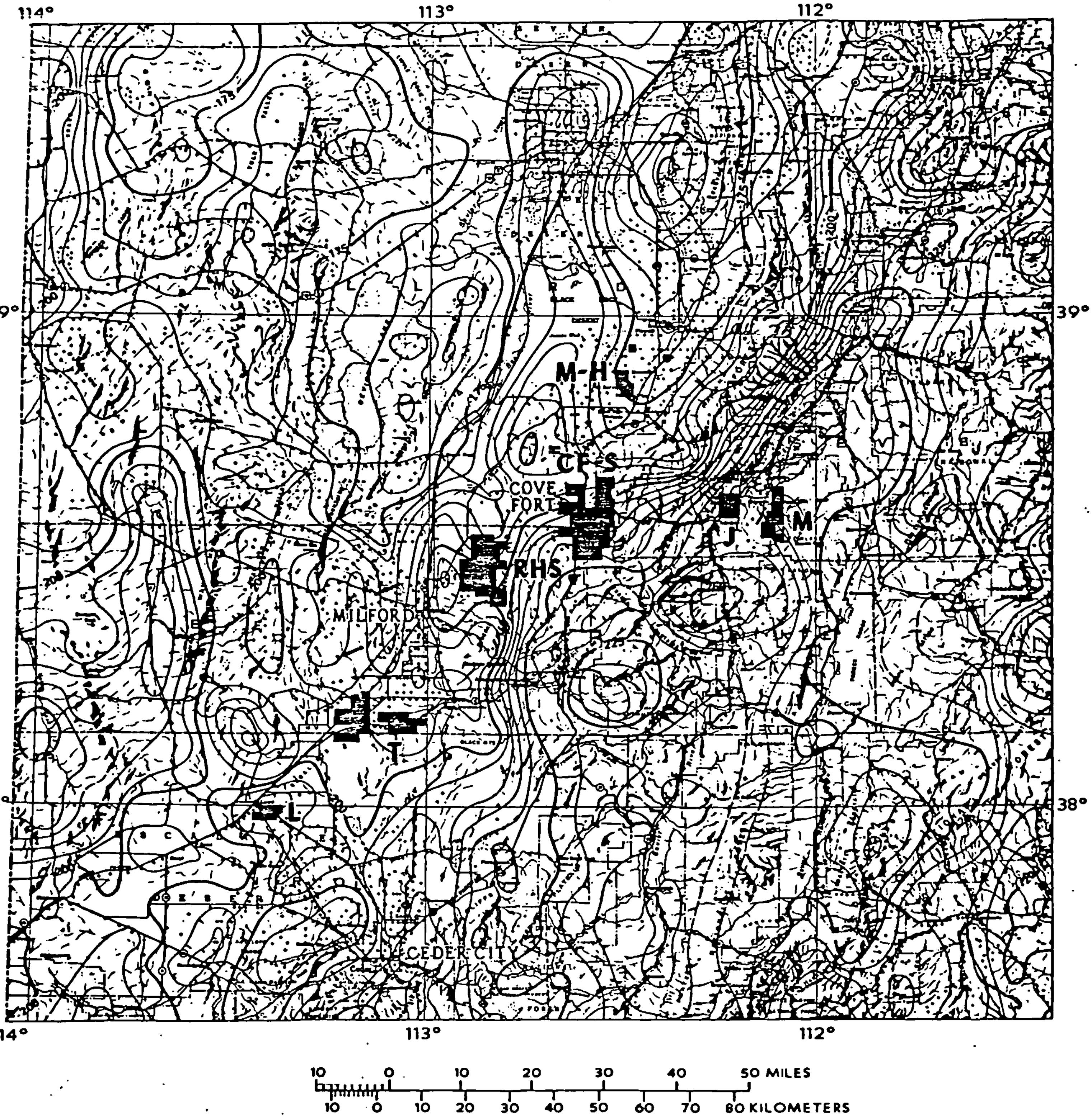


Figure 7. A portion of the Simple Bouguer Gravity Anomaly Map of Utah (after Cook et al.)

ESL-88043-JP
DOE/ID/12489.36

REGIONAL EXPLORATION

FOR CONVECTIVE-HYDROTHERMAL RESOURCES

by

Phillip M. Wright
Dennis L. Nielson
Howard P. Ross
Joseph N. Moore
Michael C. Adams
Stanley H. Ward

Earth Science Laboratory
University of Utah Research Institute
391 Chipeta Way, Suite C
Salt Lake City, Utah 84108

ABSTRACT

Intrusion of magmas into the crust is controlled on a regional scale by global tectonic and magmatic processes. Magmas generated in the mantle are mafic in composition and are relatively mobile because of their low viscosity. Felsic magmas, which have much higher viscosity, are generated by progressive differentiation from mafic magmas or by fusion of crustal material. They are less mobile than mafic magmas, and tend to accumulate in magma chambers at depths of a few to a few tens of kilometers. With continued magmatic input, felsic magma bodies of large enough size have significant thermal impact on the crust, and are believed to be responsible for most of the known high-temperature hydrothermal convection systems.

The purpose of regional exploration is to locate and prioritize geothermal prospects within large reconnaissance areas of 10,000 to 1,000,000 km^2. Regional exploration for high-temperature hydrothermal systems should concentrate in areas where there are one or more of the following indicators: (1) active or fossil surface manifestations such as hot springs, fumaroles, spring deposits, or hydrothermally altered ground; (2) volcanic rocks less than about 1 million years old, with relatively higher priority assigned if there is evidence for silicic volcanism; (3) high regional heat flow and high thermal gradients; and, (4) active tectonism and seismicity. Indicators (2) and (3) relate to the

possible presence of heat sources whereas indicator (4) gives evidence that permeability may be periodically rejuvenated.

Most regional exploration techniques are also used for subregional and detailed exploration of hydrothermal systems. Regional exploration programs should be guided by strategies specifically designed for the area of interest. Such strategies allow systematic application of techniques and promote decision-making at specific stages. A systematic approach helps reduce both the risk of failure and the cost of exploration.

NATURE AND OCCURRENCE OF GEOTHERMAL RESOURCES

Geothermal energy is thermal energy from the earth. Because the earth is hot inside, heat flows steadily to the surface at a mean rate of 82 E-3 W/m^2 and is permanently lost by radiation into space. Since the surface area of the earth is 5.1 E+14 m^2, the steady rate of heat loss is about 42 million megawatts (Williams and Von Herzen, 1974). White (1965) estimated the total thermal energy above surface temperature to a depth of 10 km to be 1.3 E+27 J, equivalent to burning 2.3 E+17 barrels of oil. The outward heat flux is about 5,000 times smaller than the solar flux, and the earth's surface temperature is, thus, controlled by the sun, not by internal heat (Bott, 1982).

Two sources of internal heat are most important among several contributing alternatives: 1) heat released throughout the earth's 4.7 billion-year history by decay of radioactive isotopes of uranium, thorium, potassium and other elements; and, 2) heat released from impacts during formation of the earth by accretion and from gravitational potential energy during subsequent mass redistribution when heavier material sank to form the earth's core. The relative contribution of these two mechanisms to the surface heat flux is not resolved.

The genesis of geothermal resources lies in the geological transport of anomalous amounts of heat near enough to the surface for access. As a result, the distribution of geothermal occurrences is not random but is governed by geological processes of global, regional and local scale.

Geothermal resources commonly have three components: 1) a heat source, i.e., an anomalous concentration of heat; 2) interconnected permeability in the rock to form a plumbing system; and, 3) fluid to transport the energy from the rock to the surface. We will discuss these elements in turn.

Heat Sources

In geothermal areas, higher temperatures are found at shallower depths than is normal. This condition usually results from one or a combination of the following five mechanisms: 1) intrusion of molten rock into high levels of the earth's crust; 2) thin crust and high surface heat flow, with an attendant high temperature gradient; 3) ascent of groundwater that has circulated to depths of 2 to 5 km; 4) blanketing of aquifers by rocks of low thermal conductivity, such as shale; or, 5) anomalous heating of shallow plutons by radioactive decay. Most high-temperature resources appear to be caused by the first mechanism. In subsequent paragraphs, we will discuss magmatism as a heat source at some length and touch upon mechanisms (2) and (3). The fourth, blanketing, is one aspect of some basin-type resources, which are usually not of sufficient temperature for generation of electricity, whereas the fifth, radioactive heating (Costain et al., 1980) has not been explored sufficiently.

Plate-Tectonic Processes and Geothermal Resources

The global-scale geological process that generates crustal magmas through several mechanisms is plate tectonics (see Bird and Isacks, 1980, for a compilation of original papers on the plate-tectonic revolution in geological thinking). High temperature at depth causes convection in the mantle by which hotter material slowly rises, spreads horizontally beneath the solid lithosphere, cools and descends again. Horizontal stresses at spreading centers (rift zones) break the lithosphere into plates which are set in relative motion. Plates on opposite sides of a rift separate a few centimeters per year while mantle material rises and solidifies in the crack to form new crust. Spreading or divergent plate boundaries are typically thousands of kilometers long, several hundred kilometers wide and cause the world's mid-oceanic mountain systems. In places, the ridge crest is offset by transform faults caused by variations along the ridge in the rate of spreading.

Plates also converge along arcuate zones and, in most locations, oceanic plates are thrust beneath continental plates. As the under-thrust, or subducted, plate descends into the mantle, it is heated by surrounding higher-temperature material, by friction and by exothermic reactions. At the descending plate's upper boundary, rich in water from incorporated oceanic sediments, temperatures become high enough in places to cause partial melting. The melts ascend buoyantly through the crust, presumably along zones of structural weakness, carrying their heat to within 1.5 to 20 km of the surface. Since the subducted plate normally descends at an angle of about 45 degrees, intrusion and andesitic volcanism occur 50 to 200 km inland from the convergent margin. Downward dragging of the crust causes oceanic trenches, which typically mark subduction zones.

Oceanic rises occur in all major oceans (Figure 1). The East Pacific Rise, the Mid-Atlantic Ridge and the Indian Ridges are examples. Geothermal resources at divergent plate boundaries include those in Iceland (Reykjavik, Namafjall, Krafla, Svartsengi), the systems in the East African rift (Olkaria) and the systems in the Salton trough in southern California and adjacent Mexico (Salton Sea, Heber, East Mesa, Cerro Prieto). Submarine hydrothermal systems occur along submerged portions of the oceanic rises (Rona and Lowell, 1980). None are being explored for geothermal development, although they represent a considerable source of energy.

Resources associated with the intrusion and andesitic volcanism of convergent plate boundaries with subduction include fields in Italy (Lardarello, Monte Amiata, Travale), New Zealand (Wairakei, Broadlands, Kawerau), Japan (Otake, Matsukawa, Hatchobaru, Kakkonda, Onuma, Mori), the Philippines (Tiwi, Mak-Ban, Tongonan, Palimpinon), Indonesia (Kamomang), Mexico (Los Azufres, La Primavera, Los Humeros), Central and South America (Momotombo, Nicaragua; Ahuachapan, El Salvador) and the Pacific Northwest (Meager Creek, British Columbia; Newberry volcano, Oregon; Medicine Lake and Mt. Lassen, California).

Back-arc basins associated with subduction often have high heat flow, diaperic magmatism and geothermal occurrences. The extensional environment of the Basin and Range province in the western U.S. and associated magmatism have resulted in such resources at Coso, California; Steamboat, Nevada; and Roosevelt Hot Springs and Cove Fort, Utah. Other resources, such as Beowawe, Nevada do not appear to be related to recent volcanism, and may be due to deep circulation along faults kept open by ongoing tectonism. Crustal thinning in the Basin and Range has resulted in an average heat flow of 80 mW/m^2 and an average geothermal gradient of 60 °C/km. By contrast, the typical geothermal gradient in the continental interior is 20 to 30 °C/km.

Transform plate boundaries are also associated with geothermal resources. The Geysers, CA is possibly one example. Leaky transforms, where open space for magmatism is created by strike-slip at a change in fault strike, may also play a role in the formation of resources such as those in the Salton trough.

Intraplate Resources -- Mantle Plumes

Another mechanism for generation of magmas is point sources of heat in the mantle. It is hypothesized that the upper mantle contains local, plume-like areas of upwelling. As crustal plates move over these hot spots, a linear or arcuate sequence of volcanos is developed. The Hawaiian Island chain is an example. The youngest volcanic rocks on the island of Kauai on the northwest end have been dated through radioactive means at about 4 million years

(Ma), whereas the volcanos Mauna Loa and Mauna Kea on the island of Hawaii at the southeast end of the chain are forming today. To the northwest, the Hawaiian chain continues as submarine volcanic mountains more than 2,000 km beyond Kauai to Midway Island, where the last volcanic activity took place about 20 Ma ago (Burke and Wilson, 1976). The Hawaiian seamounts then continue on trend beyond Midway to a point where they meet the Emperor seamount chain, which trends more northerly. The age of seamounts at the junction is about 40 Ma. From there, the Emperor chain continues to a point off Kamchatka, where the oldest seamounts are about 80 Ma. The trace of the Hawaiian and Emperor chains is consistent with the motions of the Pacific plate as determined from other data.

In the continental U. S., the trace of the Snake River Plain in Idaho is believed to document movement of the North American plate over a mantle hot spot (Christiansen and McKee, 1978). Yellowstone National Park is situated at the northeast end of the Plain, and represents the youngest locus of volcanic activity along the Plain.

Intrusion of magma to shallow depths in the Earth's crust is the most important mechanism in the creation of heat sources for geothermal energy. We have seen how processes occurring in the mantle result in magmas. There are several possible modes of interaction between rising magmas and the rocks of the crust, and we now turn to a discussion of these to develop a better understanding of the geothermal environment.

Plutonic-Volcanic Processes

In a provocative article, Smith and Shaw (1975) developed the idea of using volcanic rocks as a regional guide to high-temperature geothermal systems by estimating the heat residing in a pluton from determinations of the volume and age(s) of igneous activity. Available hydrothermal energy was postulated to depend directly on the heat provided by the magma chamber. They adopted the premise that magmas contribute little heat to the upper crust unless they form chambers. They assumed conductive cooling of the magmas but recognized that hydrothermal convection would speed cooling while continued magmatic input would extend system life. Their work forms not only a rational basis for regional exploration but also a means of assessing geothermal resources on a national basis (White and Williams, 1975; Muffler, 1979).

Magmas of different compositions exhibit different physical behaviors that influence the geometries of plutons. Viscosity is one important, composition-dependent parameter in the movement of magma. In general, silicic melts are more viscous than mafic melts as a result of polymerization of SiO_4^{-4} tetrahedra in the silicic melts, in which cation concentrations are relatively low. All melts show increasing viscosity as temperature decreases, but the viscosity of rhyolite increases faster with decreasing temperature than does the viscosity of basalt. The viscosity of silicic melts also decreases with increasing water content. Loss of fluid may be sufficient to stop an ascending melt.

Figure 2 shows phase diagrams for the basalt and granite systems (Harris et al., 1970). Consider a liquid basalt on the saturated liquidus at a depth of about 12 km, equivalent to about 4 kb pressure. This melt will be starting to crystallize. If it is intruded into the upper crust with little heat loss to the wall rocks, it will reach the surface before it intersects the solidus curve and becomes completely crystallized. The effect of water in the basalt is shown by liquidus curves for 0, 2, and 4 percent H_2O (Figure 2a). Lower water content favors ascent to the surface prior to crystallizing. The granite system (Figure 2b) behaves somewhat differently. If a melt at 4 kb along the saturated liquidus ascends without losing heat, it will intersect the solidus curve and crystallize at somewhat less than 2 kb pressure. The surface eruption of a granitic melt is favored by either an initial temperature above the liquidus or a lower water content.

From the above observations, one concludes that because of viscosity and crystallization differences, basaltic melts can move through relatively narrow conduits and tend to flow to the surface while granitic melts require larger conduits and tend to crystallize before reaching the surface. Silicic melts may form plutons of considerable dimension and heat content. Smith and Shaw (1975) pointed this out, and emphasized that silicic volcanic rocks of age less than about 1 Ma are strong evidence for a subsurface magma chamber whose heat content may be sufficient to sustain substantial hydrothermal convection.

Hildreth (1981) also studied magmatism and made contributions pertinent here. He considered magmatic systems to be fundamentally basaltic since the heat for operation of the system is derived from basalts originating in the mantle. Rocks of granitic composition are formed through either the process of fractional crystallization or fusion of pre-existing crustal material. Smith and Shaw (1975), Lachenbruch et al. (1976) and Hildreth (1981) all postulated that major silicic volcanic centers require continued thermal input from mantle-derived basaltic liquids to sustain high temperature and convective hydrothermal circulation. Basaltic liquids cannot ascend through granitic liquids because basaltic liquids are denser. Basalts appear to pond at the base granitic melts and transfer heat to the granitic melt through conduction. A shadow zone results on the surface where basaltic volcanism surrounds, but does not occur within, a felsic volcanic center.

Models given by Hildreth (1981) relate different styles of magmatism to various amounts of thermal input and crustal extension. Systems that produce magmas of intermediate composition occur in non-extensional areas, with andesitic stratovolcanos such as those in the Cascades representing an early stage of development. These convergent environments produce hydrothermal systems around the Pacific ring of fire and elsewhere world wide. High basaltic flux from the mantle in areas of extension appears to produce large amounts of rhyolitic melt, leading to explosive eruptions and the formation of calderas. The best known examples in the U. S., all of which support high-temperature convection, are

Long Valley, California (Sorey et al., 1978), the Valles caldera, New Mexico (Hulen and Nielson, 1986), and the Yellowstone caldera, Wyoming (Keith et al., 1978). Rhyolitic dome fields appear to occur in areas of crustal extension and modest amounts of heat input. Examples of dome fields that contain hydrothermal systems are Roosevelt Hot Springs, Utah (Ross et al., 1982; Nielson et al., 1978) and Coso Hot Springs, California (Bacon et al., 1980).

In summary, igneous activity in the last one million years and the occurrence of silicic volcanic rocks form important regional guides to geothermal resources. Magmatic processes produce various kinds of volcanism, from the quiet eruptions of basalt directly from the mantle at shield volcanos such as those in Hawaii and Iceland to violently explosive eruptions that create calderas such as those at Long Valley, California and Valles, New Mexico. Calc-alkaline volcanism at stratovolcanos and the formation of silicic dome fields are processes that fit between these two extremes. All of these magmatic processes can and do host hydrothermal convection systems.

Permeability

Porosity is the fraction of void space in a rock, whereas permeability is a measure of a rock's capacity to transmit fluid as a result of pressure differences. Interconnected porosity creates permeability by providing pathways for fluid flow, although there is no simple relationship between porosity and permeability for the fractured rocks common in hydrothermal convection systems.

Permeability and porosity can be primary or secondary, i.e., formed with the rock or subsequently. Primary permeability in sedimentary rocks originates from intergranular pore spaces and it usually decreases with depth and age due to compaction and cementation. In volcanic sequences, intergranular porosity and permeability are low, but higher primary permeability occurs in flow breccias at contacts between individual flows. Crystalline rocks usually have very low primary permeability. Secondary permeability in all rock types can occur in fault zones, fractures and fracture intersections, along dikes and in breccia zones produced by hydraulic fracturing.

Models of permeability in rocks have been investigated by Wyllie and Spangler (1952), Amyx et al. (1960) and Brace (1980), among others. Present permeability and porosity can sometimes be empirically determined from geophysical well logs (Hearst and Nelson, 1985). Permeabilities in rocks range over 13 orders of magnitude (Figure 3). In pristine, unfractured crystalline rock, intergranular permeability is commonly on the order of E-6 darcy or less. However, in-situ measurements at individual sites may vary 4 to 6 orders of magnitude, and zones of >100 millidarcy are common due to increased fracture density. Permeabilities of E-6 darcy or more are required for significant amounts of fluid flow (Norton, 1979). Fractures sufficient to make a geothermal well a good producer need be only a few millimeters in width, but they must be connected to the general fracture network to sustain large fluid volumes.

Norton and Knapp (1977) and Villas and Norton (1977) have studied permeability in hydrothermal systems. They conclude that the permeability history of hydrothermal systems is one of the most poorly understood parameters. Recently, Norton (1988) attempted theoretically to relate permeability and hydrothermal alteration. He observes that permeability controls the distribution and abundance of the chemical and mineral alteration in hydrothermal systems, and is in turn changed by the metasomatism. Numerical modeling reveals that the buoyancy forces generated by purely thermal effects are similar for all plutonic environments, and only the permeability of the percolation network varies significantly from system to system. Application of Norton's (1988) approach poses many challenges to modern field mapping and to integration of field data into numerical analyses.

Structure and Permeability

In most geothermal environments, overlapping structural styles of different origin occur. Manifestations are found of both the regional tectonic forces and the local stresses, related to the volcanic-plutonic system. Regional and local structure can both serve as conduits for geothermal fluids. However, formation of fractures does not guarantee permeability -- fractures may be impermeable due to dike emplacement or vein filling. The three magmatic models discussed above, mature basaltic to andesitic volcanoes, calderas and dome complexes, demonstrate different structural styles that merit consideration in our discussion or regional hydrothermal exploration.

Structural development around basaltic to andesitic volcanos, including faulting and fracturing, results from at least two stress fields, that generated by the volcano itself and regional tectonic stresses. In a paper on dike trends in the Spanish Peaks area of Colorado, Ode (1957) demonstrated the effects of a local stress system, produced by forces associated with magma emplacement, superimposed on a regional stress system. Muller and Pollard (1977) quantified this approach and were able to duplicate the dike pattern by varying the intensity and orientation of the local and regional stresses. Nakamura (1977) discussed the use of surface volcanic features to determine the orientation of dike trends and deduce the orientation of the regional stress system. Recent work (cf. Ryan, 1988 among many examples) has emphasized the application of finite-element modeling to analyze the structure of active volcanoes. Geologic mapping for the purpose of understanding the evolution of structures associated with volcanos is important in predicting where permeability development might be expected to take place. As our ability to model structural development improves, we can expect to make more accurate predictions even in the absence of surface manifestations.

The formation of calderas above silicic magma chambers is a much more energetic process than is involved in the evolution of dome complexes or mature volcanoes. It involves the explosive eruption of large volumes of magma and subsequent collapse of the roof zone into the magma chamber. Continued magma pressure after collapse may result in late-stage rhyolite domes, resurgent doming of the central portion of the caldera, or additional caldera-forming eruptions (Smith and Bailey, 1968). Structures evolve primarily in two different areas, the ring fracture system and the resurgent dome or central uplift. Mapping at Yellowstone caldera (United States Geological Survey, 1972) demonstrated strong control by ring fractures on the hot-spring system. At the Valles caldera, discussed further below, graben faults in the resurgent dome host a high-temperature hydrothermal system (Nielson and Hulen, 1984).

Caldera processes also interact with and are affected by the regional stress system. Cummings (1968) studied the interaction of the Timber Mountain caldera with Basin and Range structure and found that the caldera caused regional normal faults to bend toward its center. Regional faults host fossil hydrothermal systems in the Creede caldera in Colorado (Steven and Eaton, 1975) and present-day convection in the Valles caldera in New Mexico (Goff and Shevenell, 1987).

Rhyolite dome complexes are typically emplaced along regional structures. The setting is often complex and may result from several periods of tectonic activity with differing stress directions. There is little evidence that the process itself of dome formation produces major new faults to serve as conduits for geothermal fluids. An exception is the occurrence of phreatic, phreato-magmatic or magmatic brecciation near a rhyolite conduit.

Lithologic Controls on Permeability

Lithology also controls permeability. In the Salton trough, for example, horizontal permeabilities in porous sandstones may be orders of magnitude higher than vertical permeabilities, in which fluid flow is impeded by shales (Morse and Thorsen, 1978). Large-scale fluid flow may be controlled by major faults, but in detail the sedimentary rocks are better able to sustain a fracture network in areas where they have been embrittled by metasomatism. Permeable lithologies also occur in volcanic environments. In the Valles caldera, fluid flow is controlled both by structure and by lithology (Hulen and Nielson, 1986).

Clearly, an understanding of the geologic structure and lithology in a prospect area can lead not only to evidence for the location of a subsurface magma chamber, but also to inferences about areas of higher permeability at depth.

Fluids

For geothermal resources to be developed economically, an efficient means of bringing large quantities of heat to the surface is needed. Water normally pervades the open spaces in rocks, has a high heat capacity and a high latent heat of vaporization and is, thus, an ideal heat-transfer fluid. The density and viscosity of water both decrease as temperature increases. If buoyancy due to heating creates enough hydraulic drive and if permeability is high enough, heated water will rise. Convection brings large quantities of heat within the reach of wells and is responsible for the most economically important class of geothermal resources, the convective hydrothermal resources.

Physical Processes in Hydrothermal Systems

Convective hydrothermal resources are divided into water-dominated and vapor-dominated resources according to which phase controls the pressure. Models for water-dominated systems have been discussed by Mahon et al. (1980), Henley and Ellis (1983), and Norton (1984) among others. Rapid convection may produce nearly uniform temperatures over large volumes of the reservoir. In some places, boiling may occur and a two-phase region of mixed steam and water may exist, but the pressure is controlled by the water phase. Recharge takes place at the margins. Escape of hot fluids may be minimized by a near-surface sealed zone or cap formed by precipitation of minerals in fractures and pore spaces.

In vapor-dominated systems (White et al., 1971), convection of deep saline water brings heat upward to a level where boiling takes place. Boiling removes the heat of vaporization, thereby cooling the rock and allowing more heat to rise. Steam moves upward through fractures and is possibly superheated by the surrounding rock. At the top and sides of the system, heat is lost from the vapor and condensation results, with the condensed water moving downward to be vaporized again. Within the vapor-filled part of the reservoir, temperature is nearly uniform due to rapid steam flux, while pressure increases slowly with depth. If an open fracture reaches the surface, steam may vent or may heat the shallow groundwater to boiling. Because the surrounding rocks typically contain groundwater under hydrostatic pressure, a large horizontal pressure differential exists between the steam in the reservoir and the water in adjacent rocks. The formation of a vapor-dominated system appears to require restricted permeability at the edges and above the system as well as venting of steam at a rate in excess of water recharge (White et al., 1971).

Vapor-dominated systems may be formed from pre-existing water-dominated systems through special geological conditions. In fact, a hydrothermal system that is basically water dominated can have one or more natural zones which are vapor dominated, and vapor-dominated zones can result from production of fluids from a well if local water recharge is insufficient to keep pace with production.

Geochemistry of Geothermal Fluids

The characteristics of the major fluid are summarized in Table I and their occurrences are illustrated schematically in Figure 4. Except for hydrothermal systems recharged by seawater, the convecting fluids consist dominantly of meteoric water (Craig, 1963). As the fluids circulate through the rocks and react with them, the fluid composition changes from one dominated by Ca^{+2}, Mg^{+2}, and HCO_3^- to a sodium-chloride fluid depleted in Ca^{+2} and Mg^{+2}. Salinities typically range between 3,000 and 15,000 ppm, although values as high as 250,000 ppm are encountered in the Salton Sea geothermal field (Helgeson, 1968; Ellis and Mahon, 1977). Such fluids constitute the primary exploitable resource. Depending on the local hydrologic gradients and the extent of mixing with local groundwaters, the sodium-chloride fluids may discharge above upwelling zones as warm to boiling chloride springs that deposit sinter (amorphous silica), or as dilute chloride springs up to several kilometers from the main reservoir. Dilution of the deep sodium-chloride reservoir by overlying, cooler groundwaters may be inhibited by a subsurface silica-sealed zone developed where thermal gradients are steep (Mahon et al., 1980).

Sporadic boiling of the sodium-chloride fluids within the upflow zones is common (Mahon et al., 1980; Henley and Ellis, 1983). In areas of high relief or where the groundwater table is deep, the steam and gases, principally CO_2 and H_2S, may vent at the surface as fumaroles. Alternatively, boiling may lead to secondary, steam-heated reservoirs characterized by low chloride content. Condensation of steam and gases into oxygenated groundwater can result in acid-sulphate fluids through the oxidation of H_2S to SO_4^{-2}. Because the pH of acid-sulphate fluids is commonly less than 3 (Henley et al., 1984), they are frequently associated with advanced argillic alteration.

In contrast to the formation of acid-sulphate fluids, steam heating and dissolution of gases into deep, oxygen-depleted groundwaters produces a near-neutral sodium bicarbonate fluid. Such fluids are generally found directly overlying or immediately peripheral to the hydrothermal system. Sodium-bicarbonate springs are frequently associated with the formation of travertine (calcium carbonate) deposits.

Mixing of different fluids and dilution with cold groundwater is common in the shallow portions of hydrothermal systems characterized by relatively high permeabilities. Although mixing may significantly complicate the initial interpretation of the fluid geochemistry, use of cation ratios and the application of mixing models (see, for example, Fournier, 1981) can provide accurate information on the hydrothermal system. In contrast, little mixing may occur in low-permeability systems. Adams and Moore (1987) documented several compositionally distinct and unrelated sodium-chloride fluids in low-permeability rocks at Meager Mountain, B.C.

HIGH-TEMPERATURE HYDROTHERMAL ENVIRONMENTS IN THE U.S.

Hydrothermal systems can be grouped geologically in a number of ways, including classification by plate tectonic environment or by plutonic-volcanic associations. Table II indicates that the magmatic processes of volcano, caldera and rhyolite-dome formation take place in several plate-tectonic settings. In this table, we have attempted to characterize the predominant form of volcanism. Calderas are developed in the convergent environment at Newberry and Medicine Lake, in the divergent environment at Valles caldera, in the (back-arc?) extensional, Basin and Range environment at Long Valley and in the plume environment at Yellowstone. Dome fields are known from the divergent environment at the Salton Sea and from the Basin and Range at Coso, Long Valley and Roosevelt. In fact, we can probably generalize that calderas, dome fields and volcanos can all be found in any of the tectonic environments. We also note that some fields are not obviously associated with Quaternary volcanic rocks, as exemplified by those in the central Basin and Range at Desert Peak, Dixie Valley, and Beowawe.

We have chosen to discuss high-temperature hydrothermal systems in the U.S. under five headings: (1) systems associated with the andesitic volcanos of convergent plate boundaries such as those in the Cascades; (2) systems associated with rifting, extensional environments such as those in the Salton trough and the Rio Grande rift; (3) Basin and Range systems, which occur in an extensional environment characterized by normal, listric, thrust and gravity-glide faulting of possible back-arc origin, (4) systems associated with transform environments, perhaps illustrated by The Geysers, and (5) systems associated with mantle plumes such as Yellowstone. Even these categories are not mutually exclusive -- transform faulting is a prominent feature of the Salton trough, for example. Some occurrences, such as The Geysers, are not easily classified and the cause of Basin and Range structure is a topic of conjecture. However, this classification provides organization and a context for thought. In this section, we discuss geothermal occurrences in the Cascades, the Salton trough, the Valles Caldera and the Basin and Range, with implications for regional exploration.

Hydrothermal Systems of the Cascades -- Newberry

There are at least 12 volcanic systems in the Cascades stretching from Lassen volcano in California to Mount Meager in British Columbia. Despite this impressive display of active volcanism, however, high-temperature hydrothermal convection systems are known from only Lassen, Newberry and Mount Meager, although there are possible discoveries at Medicine Lake and Crater Lake as well. Surface hydrothermal manifestations in the Cascades are believed to be suppressed by downward flow of cold meteoric water through the porous volcanic rocks. The Cascades have not yet received the exploration attention they deserve, principally due to present low energy costs and a surplus of electrical generating capacity in the Northwest. They represent a fertile frontier area for future discoveries on the continental U. S. The Newberry volcanic system is an example of the Cascades environment.

Newberry is one of the largest Quaternary volcanos in the continental United States. It is 1,100 m high and covers 1,200 km^2. It contains a summit caldera 10 to 12 km across whose collapse was apparently caused by tephra eruptions rather than the draining of a subjacent magma chamber (MacLeod and Sammel, 1982).

A number of authors have discussed the depth and geometry of possible magma chambers beneath Newberry. Priest (1983) concluded that no convincing estimate can be made of the depth to magma based on the geometry of the caldera. Griscom and Roberts (1983) noted that Newberry is marked by a circular, positive gravity anomaly of about 18 mgals which is 19 km in diameter. The anomaly implies a causative body more than 3.9 km deep and larger in diameter than the summit caldera. The aeromagnetic data for the volcano do not indicate the presence of a large magma chamber, but the authors suggest that any pluton may still be above its Curie temperature.

Magmatic History

Geologic mapping has developed a picture of the complex magmatic history of Newberry volcano (MacLeod et al., 1981). The oldest voluminous ash-flow tuff eruptions have been dated at approximately 0.51 Ma (MacLeod and Sammel, 1982). Many of the older rhyodacite domes and flows such as the Paulina Peak dome also have similar K-Ar ages. Rhyolites of this age are good potential heat sources for high-temperature geothermal systems (Smith and

Shaw, 1975) even without the subsequent rhyolitic volcanism demonstrated at Newberry.

Our principal concern here is the latest magmatic episode, which is the heat source for the present hydrothermal system. MacLeod and Sammel (1982) point out that Newberry rhyolites that post-date the Mazama ash, erupted at Crater Lake 6856 years ago (Bacon, 1983), are chemically distinct from older rhyolites, suggesting that they represent a separate event. In addition, MacLeod et al. (1981) have mapped lower Holocene and Pleistocene basaltic-andesite flows within the caldera, demonstrating that a liquid rhyolitic chamber was not present directly beneath the caldera at that time. Figure 5 is a geologic map which summarizes the volcanic products of the youngest magmatic episode. MacLeod et al. (1981) show that this event began 6700 years bp with the eruption of several rhyolitic tephra units. Following this, basaltic cinder cones and fissure vents were formed about 6100 years bp, including the East Lake fissure which the authors describe as the only post-Mazama basaltic vent in the caldera. However, the geologic map seems to indicate that the vent may be located in the caldera wall rather than within the caldera, an important distinction in determining the areal extent of the granitic magma chamber beneath Newberry's summit. Between 6160 and 5800 years bp, numerous basaltic andesite flows occurred, principally in the NW rift zone. The Central Pumice Cone Obsidian formed at approximately 4500 years bp. The East Lake Obsidian Flow erupted at about 3500 years bp based on hydration rind dating.

At about 1600 years bp, a rhyolitic pumice fall occurred. This was followed by the Ash-Flow Tuff of Paulina Lake and the Big Obsidian Flow (BOF), with an ages of about 1300 years. The pumice fall, Ash-Flow Tuff of Paulina Lake, and the BOF are all chemically indistinguishable (MacLeod et al., 1981). Laidley and McKay (1971) have chemically analyzed the BOF, East Lake, Interlake, and Pumice Cone Obsidian Flows and find that they are chemically indistinguishable. From this they conclude that the flows originated from a common magma source.

Basaltic-andesite flows make up a major portion of the volcanism at Newberry (MacLeod et al., 1981). The northwestern rift has exerted a predominant structural control on the eruption of the mafic rocks whereas the rhyolitic events have been concentrated in the area of the caldera. It is also clear that the volume of the mafic flows is greater at lower elevations.

Structural Controls

Regional structural controls on the magmatism at Newberry are principally exercised by two features. The Brothers fault zone is a northwest-trending feature which contains numerous silicic dome complexes (Walker and Nolf, 1981). The silicic volcanism within this zone shows a progressive decrease in age toward the northwest with Newberry volcano constituting the youngest silicic activity. The other feature is the Tumalo-Walker Rim fault zone (Priest, 1983). This zone changes strike from northeast to northwest at the location of Newberry volcano (Figure 5).

Hydrothermal System

A high-temperature geothermal system occurs in the caldera at Newberry as demonstrated by core hole Newberry 2, drilled by the U.S. Geological Survey (Sammel, 1981; MacLeod and Sammel, 1982). This hole reached a bottom-hole temperature of 265 °C at a depth of 932 m. This and other geothermal test holes are shown on Figure 5. These include Newberry 1, drilled by the U.S.G.S. on the northeast flank of the volcano and N-1 and N-3, holes between 1,220 and 1,370 m deep cored by GEO-Operator. Other exploration holes also exist. To date, only the caldera proper is known to contain a hydrothermal convection system, although exploration continues on the western flank outside the caldera.

Potential reservoir rocks outside the caldera are unknown. Regionally metamorphosed rocks of the Western Cascades, which are believed to underlie the volcano at some unknown depth, may be too altered to support a fracture network unless areas of embrittlement are developed. The Cascades present a poorly understood geothermal environment, believed to have substantial geothermal potential, but for which effective surface exploration techniques are yet to be developed.

Hydrothermal Systems in Rift Zones

In the United States, there are two rift environments, the Salton trough and the Rio Grande rift, which have very different geologic characteristics. We describe the hydrothermal resources of the Salton trough first. We then discuss the Valles Caldera, both as an example of the caldera environment and as a system possibly related to the Rio Grande rift, since it occurs on its western margin.

The Salton Trough

The Salton trough is an active rift zone 320 km long by 120 km wide, representing the landward extension of the East Pacific Rise through the Gulf of California. More than a dozen geothermal systems have been found within this region of high heat flow, active seismicity, and young volcanism (Figure 6 from Elders et al. 1972). These features, and the physical and chemical characteristics of the geothermal systems, are intimately related to the tectonic setting of the trough. Significantly, only two of the geothermal systems within the trough, Cerro Prieto and the Salton Sea, display surface manifestations. Thus, discovery has relied on detection of geophysical (gravity) anomalies and thermal-gradient drilling (Elders and Cohen, 1983). Four systems are currently under development in the Salton trough -- Cerro Prieto, in Mexico, and Salton Sea, Heber, and East Mesa in the U.S.

<u>Geologic Setting.</u> The Salton trough is a sediment-filled rift zone located between the transform boundary of the San Andreas fault system and the divergent boundary of the East Pacific Rise (Elders et al., 1972). Seismic refraction data suggest that the sediments and their metamorphic equivalents may be as much as 10 km thick in the center of the trough near the U.S. - Mexican

boarder (Fuis et al., 1984). At greater depths, between 10 and 16 km, the rocks are believed to consist of mafic intrusives. Elders et al. (1972) have suggested that within the trough, right-lateral strike-slip faulting has resulted in the formation of pull-apart zones in the tensional gaps between en echelon faults which have been the focus of intrusions. The highest-temperature geothermal systems, at the Salton Sea, Brawley, and Cerro Prieto, are thought to be associated with the most active pull-apart zones.

Sedimentary rocks consist dominantly of deltaic and fluvial sandstones, siltstones and argillaceous deposits related to the Colorado River system (Muffler and Doe, 1968; Van de Kamp, 1973), believed to be no older than Plio-Pleistocene. Relatively little is known about the deeper sediments. Ingle (1982) suggested on the basis of micropaleontological studies that the Colorado River delta had begun to form by late Miocene or early Pliocene in the Gulf of California and at Cerro Prieto by mid to late Pliocene. Older sedimentary rocks exposed on the margins of the trough consist of Miocene marine and fan deposits, conglomerates, arenites and mudstone (Dibblee, 1954; Sharp, 1982).

Growth of the Colorado River delta isolated the basin now partially occupied by the Salton Sea (the Imperial Valley) from the Gulf of California and the Mexicali Valley to the south (Figure 6). Thick lacustrine and evaporite deposits indicate that freshwater lakes existed periodically in the northern part of the trough during growth of the delta. These deposits overlie the deltaic sequence at the Salton Sea and are locally interbedded with it (Helgeson, 1968; Randall, 1974; McKibben et al., 1986). The evaporites apparently represent the desiccation of the freshwater lakes (Helgeson, 1968).

Volcanic domes and mafic dikes are associated with the geothermal systems at the Salton Sea and Cerro Prieto (Muffler and White, 1969; Robinson et al., 1976; Lippmann and Manon, 1987). At the Salton Sea, five rhyolite domes intrude the Quaternary alluvium. Obsidian Butte, the western-most dome, has yielded a potassium-argon age of 16,000 years bp (Muffler and White, 1969). The Cerro Prieto volcano consists of two overlapping rhyodacite domes located northwest of the well field (de Boer, 1979; Elders, 1979). Two samples from these domes have yielded potassium-argon ages averaging 112,000 ± 71,000 years bp (Moore and Reed, unpublished data). The association of magnetic highs (Griscom and Muffler, 1971; Goldstein et al., 1984) with mafic dikes suggests that these geothermal systems are underlain and heated by large mafic intrusives.

Mafic dikes have also been encountered in holes at Heber (Browne, 1977), East Brawley (Keskinen and Sternfeld, 1982), and Mesa de Andrade (Elders and Cohen, 1983). However, the absence of significant magnetic anomalies in these systems suggests that heat is supplied by deep circulation of the geothermal fluids (Elders and Cohen, 1983).

General Features of the Geothermal Systems. The Salton trough is perhaps best known for the occurrence of the hot (365 °C), hypersaline brines (up to 250,000 ppm total dissolved solids) present in the Salton Sea geothermal system (Helgeson, 1968; Muffler and White, 1969). Similar, but slightly more dilute brines (200,000 ppm total dissolved solids) with temperatures near 300 °C, have also been discovered at Brawley and East Brawley (Elders and Cohen, 1983). No simple relationship between salinity and temperature exists within the trough, although the salinities of both the Salton Sea and Cerro Prieto fluids generally increase with depth and temperature (Helgeson, 1968; Lippmann and Manon, 1987). At Cerro Prieto, where reservoir temperatures are also near 360 °C, fluid salinities range from 15,000 to 20,000 ppm. These lower salinities are also typical of the moderate-temperature systems in the trough. For example, the Heber geothermal system is characterized by temperatures up to 199 °C and salinities of 15,000 ppm (James et al., 1987; Adams et al., 1988).

With the exception of the Cerro Prieto fluids, isotopic studies indicate that the geothermal fluids in the Salton trough represent highly evolved Colorado River water (Coplan et al., 1975). In contrast, the reservoir fluids at Cerro Prieto may be mixtures of Colorado River water and a brine of marine origin (Truesdell et al., 1981)

White (1968) suggested that the high salinities of the Salton Sea geothermal brines are produced by the dissolution of evaporite deposits. The existence of highly saline recharge fluids is supported by the fluid-inclusion data of Moore and Adams (1988), which indicate that moderate temperature (100 to 250 °C) brines with salinities in excess of 150,000 ppm have circulated through evaporite deposits contained within the upper 400 m of the system.

As geothermal fluids circulate through the sediments, matrix porosities and permeabilities are reduced (Elders, 1979) through alteration and mineral deposition. An important consequence is the development of impermeable caps that restrict near-surface hydrothermal circulation. Low cap-rock permeability results in the absence of hot springs and fumaroles over the hydrothermal systems. Cap development in the southern part of the Salton Sea field has been detailed by Moore and Adams (1988). The cap consists of an upper layer of low-permeability evaporite and lacustrine deposits and a lower layer of initially permeable deltaic sandstones. The sandstones were incorporated into the cap as downward-percolating waters were heated and deposited anhydrite and calcite in pore spaces. High thermal gradients indicative of heat transfer by conduction characterize the Salton Sea cap (Younker et al., 1982). In contrast, temperatures are nearly constant with depth within the reservoir, where heat is transferred by convection.

Hydrothermal alteration in the deeper parts of the hydrothermal systems has resulted in significant modifications to the physical and mineralogical properties of the rocks. In the highest-temperature systems, at Cerro Prieto and the Salton Sea, mineral assemblages typical of greenschist facies metamorphic rocks have been produced (Muffler and White, 1969; McDowell and Elders, 1979; Bird et. al., 1984). In contrast, secondary mineral assemblages in the moderate-temperature system at Heber are dominated by clay minerals (Adams et al., 1988). Increase in the density of the sedimentary rocks resulting from hydrothermal alteration causes positive gravity anomalies of 2 to 22 mgals (Elders et al., 1972). Many of the geothermal systems in the Salton trough were discovered by measuring thermal gradients in wells drilled into these positive gravity anomalies.

Despite locally intense hydrothermal alteration, secondary permeabilities related to fracturing and mineral dissolution can be high. In the Salton Sea geothermal field, horizontal permeabilities in the reservoir sandstones range from 100 to 500 md whereas vertical permeabilities across the shales are two to three orders of magnitude lower (Morse and Thorsen, 1978). The importance of secondary permeability and of low-permeability rocks in controlling fluid movement at Cerro Prieto has been demonstrated by Halfman et al. (1984), whose model is shown in Figure 7. In this model, the fluids enter the system from the east along faults at depths below 3,000 m and then move laterally through sandstone beds. Faults and sandstone gaps in the overlying shales allow the fluids to move vertically into progressively shallower horizons. Even though the effects of hydrothermal alteration on the sediments at Heber are much less intense, the geophysical and geologic data indicate that fluid circulation below 1600 m is controlled dominantly by faults (James et al., 1987). Here, faults connect the main upwelling center, located in the southern part of the field, with shallower production zones located to the north.

Valles Caldera

The Valles caldera is located within the Jemez volcanic field in New Mexico. This field has been continuously active for at least 13 Ma, with peaks of volcanic activity occurring in the intervals of 10 to 7 Ma and 3 Ma to the present (Gardner et al., 1986). The silicic volcanism that represents the heat source for the present geothermal activity was initiated at about 3 Ma with the eruption of a series of relatively small-volume ash-flow tuffs from vent areas that are thought to be located within the present Valles caldera but have been buried by subsequent eruptions (Nielson and Hulen, 1984). These perhaps represented the initial eruptions from a low-volume magma chamber that has continued to evolve up to the present time.

At 1.45 Ma, the Toledo caldera formed with the eruption of approximately 300 km^3 of rhyolitic ash-flow tuff that formed the Otowi Member of the Bandelier Tuff. An eruption of similar magnitude formed the Valles caldera at 1.12 Ma (Heiken et al., 1986). Soon after caldera collapse, renewed pressure from the magma chamber initiated the growth of the Redondo resurgent dome in the central portion of the caldera (Smith and Bailey, 1968), which eventually reached about 900 m of structural relief. Calculations by Nielson and Hulen (1984) suggest that the top of the magma body that caused the resurgent doming is located at a depth of approximately 5 km. During structural uplift, additional rhyolite flows, domes, and pyroclastic rocks were erupted from areas within the dome and from the ring-fracture system. Eruptions from the ring-fracture system have formed a series of rhyolite domes that range in age from 1.0 to 0.1 Ma (Doell et al., 1968).

The Valles caldera is located at the intersection of two regional structural features, the Rio Grande rift and the Jemez lineament (Figure 8). The Rio Grande rift is a major feature stretching from northern Mexico to southern Colorado. It is characterized by active extensional faulting, late Cenozoic basaltic volcanism, and high heat flow. The Jemez lineament is a northeast-trending regional structure defined by an alignment of Cenozoic volcanic centers and major faults (Aldrich, 1986). Movement began on this feature in the Oligocene and included at least some strike-slip motion. Since Miocene, movement has been normal. The Jemez fault is the most prominent feature of the Jemez lineament in the vicinity of the Valles caldera. The trend of this fault is coincident with the apical graben of the resurgent dome (Redondo Creek graben) that hosts part of the high-temperature geothermal system in the caldera. The coincidence of the graben faults with the trend of this regional structure demonstrates reactivation of this crustal flaw by magmatic-induced structural processes. Outside the caldera, the Jemez fault has hosted circulating geothermal fluids and has served as a control on formation of hydrothermal breccias (Hulen and Nielson, 1988). The fault presently channels discharge from the high-temperature hydrothermal system in the caldera (Goff and Shevenell, 1987).

During the formation of the Valles and Toledo calderas, collapse of the roof zone occurred along arcuate faults of the ring-fracture zone. This zone of faulting guided the emplacement of later rhyolite domes in both calderas (Self et al., 1986; Heiken et al., 1986). The ring-fracture zone also serves as a structural control for part of the high-temperature hydrothermal system, with Sulphur Springs being an area of extremely high heat flow (Swanberg, 1983) and intense surficial alteration (Goff et al., 1985).

Hydrothermal Systems of the Basin and Range

The Basin and Range province in this paper is taken to be a back-arc basin having extensional tectonics. The Basin and Range is the largest contiguous hydrothermal province in the United States and includes the largest number of distinct hydrothermal systems. Four high-temperature systems which occur within, but along the margins of this province, are associated with Quaternary rhyolitic volcanism. These systems are Roosevelt Hot Springs, Utah, Steamboat Springs, Nevada, and Coso Hot Springs and Long Valley, California. Common elements of hydrothermal resources in the central portions of the Basin and Range include a complex structural setting, deep circulation of meteoric waters in a region of relatively thin crust and high heat flow, reservoir permeability dependence upon fracture porosity and permeability, and extensional tectonics in areas of Tertiary volcanism.

Geologic Setting

Basement rocks throughout the Basin and Range are quite diverse. Precambrian metamorphic and intrusive rocks, common in the southern (Arizona, New Mexico) and eastern (Utah, eastern Nevada) portions of the province, are replaced by Mesozoic marine and continental arc batholiths in the central and western Great Basin.

Speed (1983) summarized data indicating that the edge of Precambrian sialic continental North America occurred in central Nevada, as defined by the western limit of lower Paleozoic continental shelf rocks, and by autochthonous and paraauthochtonous upper Precambrian and lower Paleozoic shelf strata. East of this continental margin, sedimentation throughout the Paleozoic and Mesozoic resulted in great accumulations of sandstones, shales and limestones. Silicic and intermediate volcanic rocks and plutonic rocks of the island-arc environment occur to the west (Speed, 1983).

Extensive deformation, accompanied by overthrusting, occurred in the western and central Great Basin during the Jurassic and Cretaceous while sedimentation continued to the south and east. In the early Cenozoic, the northern Basin and Range was largely an upland in which broad sedimentary basins locally formed, perhaps as a result of extensional faulting. Volcanic activity began in the northern part of the province approximately 43 Ma and spread southward and eastward to southern Nevada, Arizona, and Utah until about 6 Ma (Stewart, 1983). Volcanic and igneous activity was widespread throughout the Basin and Range during the Tertiary, and many of the major tectonic and structural features were formed during this period. Major east-west intrusive and volcanic complexes, apparent in regional aeromagnetic and gravity data and on regional geologic maps, were formed in this period. Stewart (1983) notes that local and regional extension, and local strike-slip faulting, dominated the tectonics of the northern Basin and Range during the middle and late Tertiary. Numerous low-angle extensional faults have recently been identified throughout the Basin and Range (Proffett, 1977; Gans, 1982; Wernicke, 1981; Allmendinger et al, 1983; Zoback, 1983). Stewart (1983), in summarizing these studies, quotes estimates of 50 percent to greater than 200 percent extension in Nevada and Utah, and depths of detachment of 6 to 15 km. Most of the low-angle faulting in the northern Basin and Range province probably occurred during the middle Miocene, approximately 20 to 10 million years ago (Stewart, 1983).

Zoback et al. (1981) and Zoback and Anderson (1983) conclude that basin-range faulting represents a late-stage event in a long history of extension. In the northern Basin and Range, the crest-to-crest spacing of the mountain blocks is generally 25 to 35 km, and mountain blocks are separated by alluvial valleys 10 to 20 km across. Major normal fault zones, often 50-80 km in length, bound the generally north-northwest to north-northeast tending range blocks. Zoback and Anderson (1983) studied seismic reflection data and concluded that faulting may have occurred as steeply-dipping planar, normal faults, as tilted ramps or listric faults and as assemblages of both listric and planar normal faults that may sole into a gently dipping detachment surface.

Active Seismicity

Many Basin and Range faults have remained active throughout the Quaternary (Slemmons, 1967). Historical seismicity has been noted throughout the entire province, but most particularly within the northern Basin and Range, and more frequently along the Intermountain Seismic Belt on the east, and the north-south trending Nevada seismic zone on the west (Sbar et al, 1972; Ryall et al, 1966; Smith, 1978). Zoback and Zoback (1980) discuss the Basin and Range and Rio Grande rift as a single region of active crustal spreading with common least principal stress orientations, WNW-ESE, and normal faulting with a consistent WNW direction of opening. They also note exceptions along the Wasatch fault in Utah (E-W crustal extension) and in southern Nevada (NW-SE least principal stress direction). Active seismicity has been documented locally for many hydrothermal resource areas, including major events such as an earthquake of magnitude 6.8 at Dixie Valley, Nevada and ongoing microearthquake activity at Cove Fort, Roosevelt Hot Springs, Beowawe, and Steamboat.

Crustal Thickness

Geophysical data have shown that the Basin and Range province is an area of thin crust (Smith, 1978; Keller et al, 1979).

Seismic refraction studies, and seismic P-wave velocity models by Eaton (1963), Hill and Pakiser (1966), Stauber and Boore (1978), and Priestly et al., (1982) show a variation in crustal thickness of 24 to 34 km, with the thinnest crust occurring near Fallon, Nevada. Stauber (1983) derived a new crustal thickness model across the Battle Mountain heat-flow high and adjacent areas which shows a crust 34 km thick on the southeast which thins to 23.4 km near the Copper Canyon Mine, and thickens again to 30.5 km northwest of the Battle Mountain heat-flow high. He notes that the long-wavelength Bouguer gravity field correlates well with topography on the crust-mantle boundary in northeastern Nevada. High gravity-field values occur where the crust-mantle boundary is shallow.

Heat Flow

Consistent with, and a result of, extensional tectonism and a thin crust, much of the Basin and Range province is a region of high heat flow (Lachenbruch, 1978). Sass et al. (1971) defined three heat-flow regimes within the Basin and Range: heat flow of approximately 85 mW/m^2, typical of the province average; a region of higher heat flow (>100 mW/m^2) designated the Battle Mountain Heat Flow High (BMHFH); and an area below the province average with heat-flow values less than 60 mW/m^2, named the Eureka Heat Flow Low (EHFL). These values compare with 30-50 mW/m^2 throughout much of the Sierra Nevada and Colorado Plateau provinces (Blackwell, 1988). Sass et al. (1981) presented a more detailed heat-flow map compiled from 93 measurements in the northern Basin and Range. Blackwell (1983) examined these data in great detail and provided a thoughtful interpretation of the heat flow for the Basin and Range. He noted that even with a relatively high density of heat-flow data, there are many uncertainties and unknowns in the details of heat transfer within the province due to the complex geology and hydrologic setting. Blackwell presented a reasonable heat-flow model which includes a background heat flow resulting from the superposition of a late Cenozoic thermal (intrusive and extrusive) event followed by regional extension, active to within the last 1 million years. Blackwell (1983) also noted that the high heat flow of many geothermal systems is associated with range-bounding faults. Locally higher heat flow, exceeding 200 mW/m^2, has been observed for large areas surrounding major hydrothermal systems. Blackwell (1983) notes that because of the major uncertainties in evaluating Basin and Range heat flow, it may often be more useful to evaluate the thermal data in terms of thermal gradient. The gradients determined from both granite and sedimentary lithologies overlap, and range from 25 to 40 °C/km.

Hydrothermal Systems

As a result of site-specific data made available through Department of Energy cost-sharing programs (Fiore, 1980), a substantial literature has developed for Basin and Range hydrothermal systems. Benoit and Butler (1983) reviewed the geology and reservoir characteristics of nine high-temperature

(>200 °C) reservoirs in the province, three of which are associated with Quaternary volcanism. Edmiston and Benoit (1985) extended this review with a discussion of six additional systems with reported temperatures over 150 °C. Yeamans (1983) described the hydrologic setting and probable convective flow systems for several of these same regimes. Mariner et al. (1983) reviewed the geochemistry of active geothermal systems in the northern Basin and Range. Waibel (1987), Ross and Moore (1985), Smith (1983), Struhsacker (1980), and others have presented detailed studies which help to establish the common characteristics useful in defining a Basin and Range hydrothermal resource model. Examples of such models will be presented by use of the Roosevelt Hot Springs and Cove Fort hydrothermal systems.

The Roosevelt Hot Springs geothermal system is located within the Mineral Mountains intrusive complex in central Utah. The complex magmatic and structural history of this area has been described by Nielson et al. (1978 and 1986). The heat source is manifested at the surface by rhyolitic volcanism ranging in age from 0.8 to 0.5 Ma (Lipman et al., 1978). Eruptions started with obsidian-rich, non-porphyritic flows and continued with nonwelded ash-flow tuffs and surge deposits. During the final stages of activity at 0.5 Ma, twelve rhyolite domes were emplaced.

The Mineral Mountains lie in a structural transition zone between the Basin and Range tectonic province to the west and the Colorado Plateaus to the east. Present and recent activity consist of a family of northerly and north-northeasterly faults represented by the Opal Mound fault as well as east-west trending normal faults represented by the Negro Mag fault (Figure 9). Presently, the region is undergoing east-west extension; however, within the geothermal system, east-west faults are seismically active (Zandt et al., 1982; Ward, 1983). Hydrothermal circulation is largely confined to the major faults in the area, principally the Opal Mound and Negro Mag faults.

A schematic model presented for the Cove Fort geothermal system (Ross and Moore, 1985) illustrates some characteristics of many Basin and Range systems (Figure 10). Precipitation in basin-bordering ranges composed of Tertiary volcanic and older rocks percolates to great depths through dipping horizons and fractures. Meteoric waters recharge deep alluvial aquifers within the basin, and fractures and solution-permeability zones within volcanic, sedimentary, igneous or metamorphic rocks through deep fracture zones. Fractures, sometime sealed with alteration products, may be reopened by ongoing tectonism. Percolation to depths of 3 to 5 km in areas of high heat flow, with thermal gradients of the order of 40 °C/km, provides the temperature for convection-driven upwelling through permeable channels along basin bordering faults, and at intersections with older regional structures such as east-west and northwest-trending fractures important at Cove Fort.

Implications for Regional Exploration

The best surface indication of underlying geothermal resources is active hydrothermal processes such as hot springs, fumaroles or geysers. Recent fossil manifestations may also be diagnostic and may take many forms, including sinter terraces, extinct hot springs or fumaroles and altered alluvium or bedrock. The occurrence of young volcanic rocks (<1 million years old) provides strong evidence of a heat source. This evidence is strengthened considerably by silicic volcanism because of the implication of an underlying magma chamber. Andesitic volcanos, dome fields and calderas are usually large features, with enough room for several hydrothermal systems of commercial interest. Thus, their recognition serves to select the right general area, but does not limit the area enough to proceed directly to drilling.

Active tectonism and high heat flow are also characteristic of geothermal provinces. Local lithology and structure control fluid circulation patterns and the extent of hydrothermal alteration. An important consequence of alteration is the formation of low-permeability caps that may seal hydrothermal systems and preclude surface expression. At depth, hydrothermal alteration may result in an increase in the density of sedimentary rocks, which is reflected in positive gravity anomalies at the surface. The discovery of resources in the Salton trough has relied on the recognition of these gravity anomalies and on thermal gradient measurements indicative of anomalous subsurface temperature.

Table III summarizes important characteristics reported for the better known hydrothermal systems in the Basin and Range. Most Basin and Range systems occur along or near a range-front fault, active in the Tertiary and perhaps throughout the Quaternary. Soda Lake is an exception where fluids are thought to rise along faults bordering a horst block buried beneath alluvium, as determined from reflection seismic data (Hill et al., 1979). At Beowawe, Dixie Valley, San Emidio, Cove Fort and other areas, gravity, reflection seismic, and/or magnetic data support an interpretation of substantial Quaternary alluvial fill or Tertiary volcanic basin fill, perhaps exceeding one kilometer in thickness, in the valley adjacent to the range-front fault. Vertical displacement along major fault zones may amount to two km or more. All systems occur in an area of high regional heat flow (>85 mW/m^2), when the local effects of shallow thermal aquifers, such as at Beowawe, Soda Lake, and Desert Peak, are discounted. Tertiary volcanic rocks are present at or near most systems, and may act as reservoir rocks when extensively fractured. Quaternary basaltic volcanism occurs at Cove Fort and at Soda Lake, but probably is not the main heat source for the systems.

Lower temperature reservoirs may form in porous and permeable horizons within the alluvium or Tertiary volcanic rocks, and mixing with meteoric waters may give rise to secondary, lower temperature reservoirs in fractured crystalline or sedimentary rocks within the range block. The continuity of major fault and fracture zones resulting from Basin and Range extension, provides fluid pathways for lateral migration along north-trending faults for tens of kilometers and may give rise to hot springs, fumaroles, altered alluvium, or sinter deposits, and to widespread shallow reservoirs and heat flow anomalies some distance lateral to the primary high temperature conduits. This appears to be the case at Dixie Valley, Soda Lake, Beowawe, Cove Fort, and many other resource areas.

Regional exploration strategies should be based on the above observations, but should recognize the many individual differences among hydrothermal environments and the systems in those environments.

REGIONAL EXPLORATION TECHNIQUES

Geothermal exploration is an interdisciplinary endeavor that includes geology, geochemistry, geophysics and drilling. In some sense, geology and geochemistry constitute the primary disciplines. Geophysical methods are applied mainly to help map geological or geochemical characteristics of geothermal systems, and geophysical data must always be interpreted in terms of their geological or geochemical significance. In this section, we summarize the principal techniques used for regional exploration (Table IV). Further details and excellent summaries of most techniques are provided by Goldstein (1988). Cost-effective application of these techniques requires proper training and experience.

Geological Techniques

The geological methods constitute basic documentation procedures required for any qualitative or quantitative approach to exploration and form the framework in which other types of exploration data are interpreted.

Geological Mapping

Geological mapping is the most fundamental, and in our experience, the most under-utilized tool available to the geothermal explorationist. It is a procedure in which the geologist systematically records on maps and sections geologic observations made in the field. Without systematic mapping, these observations go unrecorded, and as a result critical information is lost or overlooked. The geologic map provides the basis for the selection of drilling targets, and the planning and interpretation of subsequent geochemical and geophysical surveys.

During geologic reconnaissance, emphasis should be placed on mapping and describing rock types, faults, hot spring deposits, and hydrothermal alteration assemblages within the region. The age relationships among these features should be determined. Volcanic rocks should be especially studied and dated by radioactive means or through field relationships. In geologically complex terrains, such as those that characterize many young volcanic systems, surface exposures may be poor and lithologic correlations difficult to make. In these areas, chemical analyses of the rocks may greatly assist in regional correlations. From such correlations, the relative ages and offsets of faults can be deduced. These data can be supplemented with the various dating techniques discussed below to determine time-temperature relationships and the significance of the observed alteration assemblages.

Careful attention should be given to the characteristics and behavior of the rocks within the region. The density and orientation of joints and fractures, the distribution of rocks with high initial permeabilities (e.g. flow breccias, lithologic contacts), and the relative susceptibility of the rocks to hydrothermal alteration should be described. Such information is generally not recorded during regional mapping for purposes other than geothermal exploration. For example, the Valles caldera was covered by a regional map (Smith et al., 1970) that constitutes one of the classic works in volcanology. However, more detailed mapping and a greater understanding of the structure was required for exploration in the area (Behrman and Knapp, 1980).

During geologic reconnaissance, the opportunity to study samples from available drill holes arises. Special effort should be made to study data and samples from any geothermal occurrence in the area. Core logging is straightforward since core provides an excellent sample for documentation of hydrothermal alteration as well as lithologic and structural relationships. The logging and interpretation of drill cuttings is more challenging (Hulen and Sibbett, 1982) but is well worth the effort.

Imagery Interpretation

Stereo air photographs and satellite imagery both provide useful data sets for regional exploration. Enlightened interpretation has proven valuable for mapping regional lineaments that may control the locations of intrusions, mapping local fracture patterns that may control hydrothermal convection systems, detecting hydrothermally altered rocks, detecting areas of altered or unhealthy vegetation due to alteration or toxic trace elements in the soil and providing basic geologic data. Air photos are available for many areas of the world from governmental agencies. Skylab orbiting photographs and U-2 high-altitude photographs having high resolution are also available for limited areas of the earth. In areas where photos are not available, it may be worth the small cost per unit area to have them flown to specification.

The most comprehensive collection of imagery from orbit is that obtained from the Landsat program. Landsats 1, 2 and 3 flew the 4-band multispectral scanner (MSS) sensor system, whereas Landsats 4 and 5 flew the thematic mapper (TM), a 7-band imaging system designed to be especially useful for geologic application. The TM images have a spatial resolution (pixel size) of 30 m compared to 79 m for the MSS. Landsat images are available for most of the world from the Earth Observation Satellite Company (EOSAT) in Sioux Falls, South Dakota, who provide distribution under an agreement with the U. S. Government.

Because satellite imagery is in digital form, it is easily processed by computer. Digital processing systems include software written by NASA (the ELAS system) and the PC-based ERDAS system. Algorithms are available for contrast stretching, spatial-frequency filtering, principal component analysis, band ratioing and pattern recognition (Sabins, 1987; Drury, 1987). Drury (1987) and Watson and Regan (1983) give illuminating examples of these techniques applied to mineral and petroleum exploration, most of which is directly applicable to geothermal exploration. However, the use of satellite imagery in geothermal work seems to have been limited to date.

Structural Analysis

Essentially all high-temperature reservoirs are structurally controlled. In tectonically active terrains, such as the extensional provinces of the western U.S., structures important to the geothermal system may be of several different ages and origins. In the Basin and Range province, fluid movement is influenced by thrust faults of Mesozoic age and normal faults developed since Tertiary time. Studies of fault zones intersected in high-temperature reservoirs demonstrates complex histories of sealing and rupturing.

Examples of the importance of different structural trends within individual systems have recently been documented for both Roosevelt Hot Springs (Nielson et al., 1978; Sibbett and Nielson, 1980) and Cove Fort-Sulphurdale (Moore and Samberg, 1979; Ross and Moore, 1985). At Cove Fort-Sulphurdale, the major upflow zones are controlled by steeply dipping normal faults. Thrust faults containing cold water have been recognized at depths of 1,000 m within the system. In addition, geologic mapping has identified low-angle normal faults that bound large-scale gravitational glide blocks exceeding 600 m in thickness and 6 km in length. The glide blocks act as a tectonic cap to the system at Cove Fort. This cap has reduced near surface heat flow over the central part of the reservoir and covered many of the normal faults that control fluid movement.

Similar low-angle faults have been mapped at Roosevelt Hot Springs. In contrast to Cove Fort-Sulphurdale, where the glide blocks are composed of ash-flow tuffs, the glide blocks at Roosevelt are composed of granites and metamorphic rocks. These rocks have behaved in a brittle fashion. As a result of fracturing within the glide block, near-surface permeabilities are higher than those below its sole.

In-situ stress orientations have recently been found to vary significantly not only among wells in a geothermal field, but also within individual wells (Allison and Nielson, 1987). Analysis of borehole breakouts derived from oriented caliper logs (e.g. dipmeter and fracture identification logs) provides semi-continuous stress data relatively cheaply and quickly. Breakouts occur along the minimum principal horizontal stress direction as a result of stress concentration on the borehole wall, 90 deg to the applied (i.e., tectonic) stress direction. Tensile and shear failure of the rock at the stress concentration points results in borehole elongation, or breakout.

In situ stresses throughout North America as determined from breakout and other data are mostly of regional extent. However, in geothermal wells, the stress system changes orientation abruptly at almost every fault. Multiply-oriented stress fields have profound implications for siting and completing geothermal exploration, production and injection wells. The stress system determines the direction of hydrofracs as well as which fractures are likely to be kept open or forced shut within a block. Analysis of borehole breakouts can and should be done with the first exploration wells in an area, with the results being analyzed in terms of regional and local geologic structure. Such analyses can help to indicate the appropriate direction to deviate a well to crosscut open fractures, for example.

Age Dating

As discussed previously, the Smith-Shaw model for volcanic-related geothermal systems is predicated on age and duration of the associated magmatic system. The technique most often employed to determine age of volcanic rocks is the potassium-argon method, the details of which are described in numerous texts on the subject (see Durrance, 1986, for a review of radioactive techniques in geology).

Recent work has concentrated on determination of the timing of thermal events rather than simply the age of volcanic rock units. Evans and Nielson (1982) used fission-track ages along with information on the closure temperature of mineral-potassium argon systems to determine the uplift history of the Mineral Mountains intrusive complex. This enabled them to determine the genesis of some of the faults that control the Roosevelt Hot Springs geothermal system. Harrison et al. (1986) applied $^{40}Ar/^{39}Ar$ systematics to determine the age of heating of the Fenton Hill hot dry rock site adjacent to the Valles caldera. They determined that this area was heated after the caldera-forming events, probably in the past few tens of thousands of years.

Geochemical Techniques

Geochemistry of fluids, rocks, and soils plays a key role in regional geothermal exploration. Although some geothermal systems display little or no surface manifestation, many are marked by hot springs and surficial alteration. Drilling to depths in excess of 3 km in many geothermal systems has provided data on the chemical composition, evolution, and thermal structure of the deep reservoirs and their relationships to overlying near-surface features. Of particular significance has been the recognition of compositionally different fluids within different parts of individual systems and the development of quantitative geothermometers based on their chemical and isotopic compositions. Concurrent investigations of altered rocks and soils have clarified the relationships among the hydrothermal alteration and the physical and chemical processes occurring at depth. These relationships provide the basis for regional geochemical exploration.

Fluid Geochemistry

The sampling and analysis of hot spring and well fluids is a primary tool in the regional exploration for geothermal systems. Fluid composition can be related to the geometry, size and temperature of a geothermal resource. Despite the diversity of geologic environments hosting geothermal systems, most systems display similar fluid types that reflect common physical processes.

When the existing springs and wells have been sampled and analyzed, a variety of interpretation techniques can be applied. As a first step, it is frequently possible to obtain an indication of the regional hydrologic patterns and the locations of the upwelling centers through the distribution of conservative, or nonreactive, elements. The most conservative elements in geothermal fluids are boron and chloride, although other elements such as fluoride and the trace metals can also be used. Because groundwaters seldom contain appreciable amounts of trace elements, variations in their concentrations can be used to detect dilution of geothermal fluids (Mahon, 1970). Figure 11 shows the distributions of boron and chloride from a large region around Roosevelt Hot Springs in southern Utah where there has been significant exploration. Most of the samples were obtained from cold, shallow water wells. The high boron and chloride concentrations in the northern part of the area result from thermal fluids that move laterally away in the alluvial valley fill, the outflow from the Roosevelt system. A second upwelling center is seen in the southwestern corner of the map, where high boron values are related to the Thermo geothermal area, a lower-temperature occurrence.

Analysis of fluids can provide an estimate of the reservoir temperature and the amount of dilution affecting the fluids. Chemical geothermometers are empirically-derived or theoretical chemical relationships based on the change in chemical composition of the fluids with temperature. An extensive review of the use of geothermometers can be found in Fournier (1981) and in Henley et al. (1984). The equations for these geothermometers are listed in these publications and in Goldstein (1988).

The most reliable geothermometers for regional exploration are the Na-K-Ca(-Mg) (Fournier and Truesdell, 1973; Fournier and Potter, 1979) and the silica geothermometers (Fournier and Rowe, 1966; Fournier and Potter, 1979). The Na-K-Ca(-Mg) geothermometer is based on ion ratios in the fluid, and is less affected by dilution than is the silica geothermometer, which is based on the concentration of silica. Because of these characteristics, the Na-K-Ca(-Mg) geothermometer can be used to predict reservoir temperatures from analysis of surface fluids, while the silica geothermometer can be used to estimate the amount of mixing of the geothermal fluid with meteoric water (Fournier and Truesdell, 1974; Fournier, 1977). As a prospect is identified by regional exploration and additional data become available through drilling, the effects of mixing, boiling and conductive cooling can be further evaluated by consideration of chloride-enthalpy relationships (Fournier, 1979).

Soil Geochemical Surveys

Soil surveys represent a relatively low-cost method of detecting geothermal systems in a regional exploration program, of obtaining more detailed information on the areal extent of a geothermal system and of locating areas of high near-surface permeability where one might want to begin exploration drilling. Among the volatile constituents commonly transported by geothermal fluids, several have proven to be particularly useful. These constituents include mercury, radon, helium, carbon monoxide and carbon dioxide. Soils can be analyzed directly for species that adsorb, such as mercury. Alternatively, volatile constituents can be collected in gas traps buried in the soil.

Mercury soil surveys have proven to be effective in a variety of geologic environments (e.g. Matlick and Buseck, 1976; Capuano and Moore, 1980; Klusman and Landress 1978, 1979; Christensen et al., 1983; Varekamp and Buseck, 1984). The relationships among soil geochemistry, heat flow, faulting, and subsurface movement of geothermal fluids in shallow aquifers is illustrated by studies of the Roosevelt Hot Springs geothermal system (Christensen et al., 1983). Figure 12 shows the distribution of mercury in an area over the thermal system. Log-normal cumulative frequency plots show that at least two populations of mercury occur in the Roosevelt soil data. These plots define a background value of 29 ppb and a threshold value for anomalies of 58 ppb mercury. Anomalous concentrations of mercury in soils occur in a series of closely spaced northeast- and northwest-trending zones that parallel major fault directions. Zones of high permeability, characterized by extreme enrichment of mercury, occur at the intersections of structural trends and frequently coincide with hot-spring deposits. Extension of the anomalous mercury concentrations to the northwest occurs beyond the known limits of the reservoir and appears to reflect the release of mercury vapor from thermal fluids in the outflow plume within shallow alluvial aquifers. This flow is documented as well by an extension of the shallow thermal anomaly and by a plume evident in regional groundwater chemistry (refer to Figure 11).

Geophysical Exploration Techniques

Wright et al. (1985) provided a review of the application of geophysical methods to geothermal exploration. A further review is presented in Goldstein (1988). Because of these previous publications, we will only briefly discuss the most important regional geophysical methods in this section.

The Thermal Methods

A variety of thermal methods respond directly to high rock or fluid temperature, the most direct indication of a geothermal resource. Among these methods are measurements of thermal gradient and heat flow, shallow-temperature surveys, snow-melt photography and thermal-infrared imagery. The first two of these methods have proved to be the most useful by far. Shallow-temperature surveys (LeSchack and Lewis, 1983) have the advantage that drilling costs are much less than in conventional thermal gradient and heat-flow surveys. However, corrections must be carefully applied to make the data useful. An additional limitation is that shallow surveys are highly susceptible to disturbance due to ground water flow. Masking of thermal anomalies with resultant downgrading in priority of an area is a possibility that must be avoided. Hydrologic disturbance of conventional thermal gradient surveys in also a threat that must be evaluated in interpreting results. Smith and Chapman (1983) have begun to deal quantitatively with this important problem, and further work on it is warranted.

The Electrical Methods

Perhaps the most important physical property change due to the presence of a hydrothermal system, other than elevated temperature and heat flow, is the change in electrical resistivity of the rock-fluid volume (Moskowitz and Norton, 1977). Higher temperature increases ionic mobility up to about 300 °C, and hence increases conductivity. Ionic conduction in rocks also increases with increasing porosity, increasing salinity, and increasing amounts of certain minerals such as clays and zeolites. Most hydrothermal systems have an associated zone of anomalously low resistivity (high conductivity) due to one or more of these factors. Goldstein's (1988) discussion of the electrical methods is quite complete, and should be consulted for more information.

The Gravity Method

Density contrasts among rock units permit use of the gravity method to map intrusions, faulting, deep valley fill, and geologic structure in general. Gravity surveys are used in the Basin and Range and similar settings as a relatively inexpensive means of obtaining structure and thickness of alluvium. Gravity has proven useful in the location of positive anomalies associated with densification of sediments due to metamorphism and silica deposition in the Imperial Valley of California (Muffler and White, 1969; Biehler, 1971; Elders et al., 1978). In other areas, gravity highs may be due to rhyolite domes and hydrothermal alteration (MacDonald and Muffler, 1972). A common association of negative gravity anomalies with granitic intrusion into crystalline rock is well known to mining geophysicists (Wright, 1981). Isherwood (1976) concluded that a large gravity low over the Mt. Hannah area at The Geysers field in California is most likely due to a hot, silicic magma under this area. This interpretation has been supported by teleseismic studies (Iyer et al., 1979).

The Magnetic Method

Magnetic surveys, either airborne or ground, have been conducted in many geothermal exploration programs. Their use is in structural or lithologic mapping or in locating decreased rock magnetization caused by hydrothermal alteration. Magnetic anomalies in New Zealand geothermal fields have been interpreted as being due to conversion of magnetite to pyrite (Studt, 1964). A magnetic low occurs over a part of the hot spring area at Long Valley, and is interpreted by Kane et al. (1976) as due to magnetite destruction. Such an effect would, of course, remain in extinct hydrothermal systems. The locations of faults, fracture zones, intrusives, silicic domes and major alteration areas are apparent on data we have examined from Coso Hot Springs, California, from Baltazor, Tuscarora, McCoy, and Beowawe in Nevada, and from Cove Fort-Sulphurdale and Roosevelt Hot Springs, in Utah. Magnetics are routinely used in Iceland to delineate dikes, some of which are bordered by zones of high permeability (Palmason, 1976; Flovenz and Georgeson, 1982).

Magnetic data may also yield information of value in defining major regional trends. The Monroe Hot Springs, Chief Joseph, Cove Fort-Sulphurdale, and Roosevelt Hot Springs KGRAs are all located in the Pioche-Beaver-Tushar mineral belt, as discussed in the next section. Bacon (1981) interpreted major structural trends and fault zones from aeromagnetic data in the Cascades. Magnetic data can also be used to estimate the depth to the Curie isotherm (Bhattacharyya and Leu, 1975; Shuey et al., 1977; Okubo et al., 1985 and many others).

The Seismic Methods

Seismic noise surveys and microearthquake studies have both been used in regional exploration. There is limited evidence that seismic waves propagate from some hydrothermal occurrences. Liaw and Suyenaga (1982) detected high-velocity body waves in data recorded at Beowawe, Nevada, but they did not detect such waves at Roosevelt Hot Springs, Utah. Liaw and McEvilly (1979) failed to find body waves at Leach Hot Springs, Nevada, but they did find energy propagating as surface waves from the vicinity of the thermal manifestations. Their paper presents the foundations for proper survey design and data analysis. There is still considerable question about the utility and reliability of seismic noise surveys in helping to prioritize exploration areas.

The frequency of earthquake occurrence is observed world wide to vary with magnitude. Small-magnitude earthquakes occur much more frequently than do large-magnitude earthquakes. At the microearthquake level, Richter magnitude -1 to 2, many areas are found to be quite active. Accurate location of these microearthquakes can provide data helpful in locating active faults that may channel hot water toward the surface (Hamilton and Muffler, 1972; Hunt and Lattan, 1982). Microseismic activity is often found to be episodic. Zandt et al. (1982) document a two-year survey at Roosevelt Hot Springs during which only a few events were recorded until the last two months of the survey, when more than 1,000 microearthquakes occurred. These microearthquakes appeared to occur on the down-dip extension of one of the permeable faults in the area, the Negro Mag fault.

REGIONAL EXPLORATION STRATEGIES

Regional, or reconnaissance, exploration is used to identify and prioritize prospect areas for further exploration. Many of the same techniques used for regional exploration are also used to map and define a system after discovery and to help assess the feasibility of economic development. The primary reason for organized exploration is to decrease the risk of failure and the cost of discovery and evaluation of resources. Essentially all activities are ultimately focussed toward the siting of successful drill holes. Because geothermal drilling is so costly, refinement of the exploration process and systematic application of proven techniques have great potential for lowering development costs by avoiding wasted drill holes. We advocate the formulation of exploration strategies based on conceptual resource models which apply to the area under consideration and the constant updating of these strategies and models as exploration proceeds. Such an approach requires knowledge of the several modes of genesis of hydrothermal systems, of the physical and chemical processes within hydrothermal systems and of the responses of various exploration methods to a significant geothermal occurrence. Also required are enlightened data collection and interpretation.

In this section, we will consider how the hydrologic environment affects the expressions of hydrothermal systems, selection of regional exploration areas and exploration strategies.

Arid and Wet Environments

Processes of intrusion, development of permeability in rocks and convective circulation occur world wide from the same mechanisms. However, each particular geothermal resource is a unique blend of these fundamental processes and the environment in which they operate. We have considered the interaction of geological process with the geological environment in previous sections. Another important factor is interaction with the hydrologic environment.

We recognize two general classes of hydrologic environment important in geothermal exploration, the arid and the wet environments. The other general class, the marine environment, is not discussed in this article. Essentially all of the explored geothermal resources in the United States occur in arid regions. Only the Cascades range of California, Oregon, and Washington could be classified as a wet environment having geothermal resources, but few resources have been discovered there relative to the number expected on the basis of the active volcanism. Other arid regions with geothermal resources include the East African rift and the South American Andes mountains. Wet environments containing geothermal resources include Japan, the Philippines, Indonesia, New Zealand, the Mexican neovolcanic belt and Central America.

The primary effect of the hydrologic environment in terms of regional exploration is on the surface manifestations of underlying hydrothermal systems. In wet regions, there are more natural springs of all types, including thermal springs associated with hydrothermal systems. Dilution of upwelling thermal waters by near-surface ground water is common. Boiling at depth is likely to produce a blanket of shallow, steam-heated ground water over a deeper system and may constitute an exploitable resource by itself. Fumaroles are common, resulting from boiling of either the deep sodium-chloride fluid or of the steam-heated ground water. Springs having a component of thermal water, recognized by temperature or chemistry, may occur kilometers from the center of convective upwelling.

In arid areas, surface manifestations of underlying hydrothermal systems are fewer and sometimes different in character from those in wet areas. It is common for hydrothermal systems in arid environments to have an outflow plume that moves in aquifers at or below the depth of the water table and yields little no surface indication. Thermal springs may be less dilute, resulting in sinter or travertine aprons of considerable size. Surface indications are usually more closely confined about the center of upwelling, and one may often drill the immediate areas of springs with success.

The hydrologic regime is modified by the surface and near-surface rock types. In areas covered by recent volcanic rocks, vertical permeability is usually great. In two core holes on the flanks of Newberry volcano in the Cascade range, for example, cold water flows downward about 1 km before vertical permeability diminishes and the water flows away laterally (Swanberg and Combs, 1986; Lemieux et al., 1988). Downward and outward flow of water in high-relief, wet environments pushes thermal springs kilometers to the side, as documented in the Philippines (Whittome and Smith, 1979; Lawless et al., 1983) and lowers the near-surface thermal gradient. Heat-flow data in the upper, disturbed portions of holes, may not reflect an underlying resource. By contrast, in areas where crystalline rocks outcrop, there is little vertical permeability and little downward percolation of meteoric water except in fracture zones, contacts and other structures. Holes drilled for measurement of temperature gradient and heat flow may need to be only a few hundred feet deep to encounter undisturbed temperatures. Surface manifestations are more likely to be confined to upflow on fracture and fault zones and to lie closer to the center of the system than in wet, volcanic environments.

The hydrologic environment may also control the total amount of recharge available to a hydrothermal system. This has been recognized and taken into account in exploration in the Basin and Range. Small topographic basins with low bordering ranges may capture little precipitation, and the regional exploration priority of such basins may be lowered at an early stage on this account.

One must also recognize that the hydrologic environment may have changed in the geologically recent past. For example, the climate in the Basin and Range province of the western United States has become progressively drier over the past 20,000 years. Many more springs occurred in the recent past than occur today. The search for deposits from past thermal springs forms an important part of regional exploration. One problem encountered is that large lakes (Bonneville, Lahonton) once filled some of the valleys, and tufa and caliche deposits associated with these ancient lakes are easily mistaken for hot spring deposits. One should also note that travertine and sinter are easily weathered in the surface environment. Consequently, even small, older spring deposits can be significant.

Regional Area Selection

The objective of regional exploration is to locate and prioritize prospect areas 20 to 100 km^2 in size for subregional and detailed exploration within large areas 10,000 to 1,000,000 km^2 in size. The choices to be made are the selection of the regional or reconnaissance area and of the techniques that will be used to explore the area and prioritize prospects.

Geothermal systems do not occur at random, but their locations are controlled by the geologic processes we have discussed above. On a global scale, geothermal occurrences are concentrated near and result from the plate tectonic processes of spreading at divergent plate boundaries and subduction at convergent plate boundaries. Also important in some areas are point sources of heat caused by mantle hot spots. All of these processes cause intrusion of magmas which carry their heat to shallow enough levels in the crust that significant permeability can be maintained under the ambient pressure and temperature conditions through active tectonism, and convection can be sustained.

High- and moderate-temperature hydrothermal resources occur predominantly in areas where tectonic and intrusive processes have been active during at least the last few million years. In such areas, regional heat flow is found to be high, but reliable measurements may be difficult to make depending on the near surface hydrology. These areas are usually also marked by thermal springs, spring deposits and fumaroles. Regional geologic maps and reports should be consulted to outline areas of recent volcanism, faulting and surface geothermal manifestations, and these areas chosen first for regional exploration.

Other information may also be useful in regional area selection. Because disseminated and replacement-type base-metal deposits are believed to have been formed by circulating hydrothermal fluids, known hydrothermal systems can be viewed as the present-day manifestation of ore-forming processes (White, 1981). It is well known that many mineral deposits occur in provinces, often within linear or arcuate belts, that represent crustal and mantle geologic controls on the processes of formation, and that within these provinces are mineral districts containing one to many significant deposits. A reasonable extension of the concept of mineral provinces to geothermal occurrences would indicate that geothermal exploration should take into account the totality of mineral occurrences of hydrothermal origin leading up to present-day hydrothermal systems to help define the regional area most prospective geothermally. Within such provinces or belts, one might go one step further by using the district concept to postulate that highly prospective areas exist in the near vicinity of known hydrothermal systems.

These ideas are illustrated by an example from Utah with which the authors are familiar. Figure 13 shows a metalogenic trend map of Utah, indicating the several mineral belts that have been recognized. It can be seen that all of the known high-temperature hydrothermal systems as well as several moderate- and low-temperature systems in Utah lie in the Pioche-Beaver-Tushar mineral belt. This belt is characterized by extensive intrusive and extrusive rocks to the south and thin volcanic rocks overlying thick Paleozoic through Tertiary sedimentary rocks with a few intrusions to the north. Figure 13 also shows the relationship of the known geothermal occurrences to the Intermountain Seismic Belt, a highly active zone of seismicity. Figure 14 shows a portion of the aeromagnetic map of Utah across the Pioche-Beaver-Tushar trend. The northern boundary is clearly delineated in the magnetic data. The abrupt change in the magnetic character is interpreted by Mabey et al. (1978) to be due to a large-scale crustal structure associated with the mineral belt. The relatively young rhyolite domes at Roosevelt Hot Springs and the andesitic lavas at Cove Fort attest to the fact that magmatic processes are still active in this belt today. The recognition of the Pioche-Beaver-Tushar trend helps to prioritize the total area in this portion of the Basin and Range for regional exploration.

Regional Exploration and Exploration Strategies

Ward et al. (1981) presented a strategy for exploration for high-temperature hydrothermal systems in the Basin and Range province. Their strategy extended from regional exploration through reservoir modeling and feasibility study, and is, thus more comprehensive than the one given here, which considers only regional exploration as previously defined.

An exploration strategy helps one systematically to apply a selected mix of exploration techniques tailored to the geological environment and the expected geothermal resource characteristics. A strategy promotes staged exploration and helps minimize risk of failure and cost. Because geothermal resources are so variable in detail, even within resources of the same general type, it is not possible to specify a certain mix or sequence of exploration techniques that will work or be the most cost-effective in all circumstances. There is no exploration strategy that can be blindly applied with the expectation of success. The appropriate strategy must be designed specifically for application to the area by the geoscientists performing the work and interpreting the data.

The Exploration Process

Figure 15 illustrates the general processes of exploration. The first step (denoted by (1) in the figure) is to assemble available data and determine (2) if there are missing data critical to further progress. If there are, these missing data are collected (3) and an integrated interpretation (4) is performed.

By integrated interpretation, we mean one that satisfies all data sets simultaneously. In order to perform this integrated interpretation, several items are required (5). One must have in mind conceptual geological, geochemical, geophysical and hydrological models of the various types of hydrothermal resources expected to occur in the area so that their signature can be recognized when it occurs in the available data. Models are formulated through knowledge of the processes of formation of geothermal resources and of the geologic and hydrologic environments in the exploration area. Study of case histories is a great aid in the formulation of conceptual models. Various data interpretation aids are also needed. Computer programs for interpretation of magnetic and electrical geophysical survey data and for treatment and interpretation of geochemical data may be cited as examples. Experience is an additional essential ingredient. The explorationists must have appropriate educational backgrounds in the earth sciences and experience in application of the techniques in order to be effective. Strange, but there is an unfortunate tendency in geothermal exploration, as well as in petroleum, mineral and groundwater exploration, for people with little or no earth science background to conduct exploration programs and make exploration decisions. The result is almost always wasted money and time. A well-qualified team of geologists, geochemists and geophysicists is a good investment in any exploration program.

After an integrated interpretation, there is a decision whether or not to continue (6). Are the geological conditions in the area permissive for geothermal occurrences? Is there evidence of thermal springs or of recent volcanic rocks? If so, the next step is to plan (7) and execute (9) a rational exploration program. For planning, several essential ingredients are needed (8). The same types of quantitative interpretation aids that were used in (5) can be used to predict, from the updated conceptual models, the responses to be expected from the various exploration techniques. Those techniques whose predicted responses are the least ambiguous, balanced against the costs of application, are chosen and the surveys are designed based on the predicted areal size and magnitude of anomalies.

Once the surveys are performed, integrated interpretation is again needed (10). The same set of ingredients (11) as discussed above in (5) is needed for this interpretation. A decision is then made whether to continue or not. Continuation may involve either the collection and interpretation of more regional data (return to (7)) or the *movement* to subregional exploration. This general process can be extended to detailed exploration and applied equally well to mineral or petroleum resources.

We note that the conceptual models used in exploration guide thinking to an extent perhaps not consciously realized. Having an incorrect model in mind can lead to failure to recognize the response of a geothermal resource in the data. In the initial stages of exploration, conceptual resource models lack the detail that it will be possible to add as more information is developed. However, the models are vital at each stage in designing the next stage of exploration because they can be used to predict the various surface geothermal manifestations one might expect, the geochemical signatures of a hydrothermal system in the particular environment and the types of geophysical surveys that may succeed in detecting a resource. Continued updating of the models is an important factor in successful exploration.

Regional Exploration Strategy

Figure 16 is a suggested regional exploration strategy that should be applicable to many environments but is given here mainly as an illustration. It begins with the collection and integrated interpretation of available geological, geochemical and geophysical data sets as the first step. The result of this step is the identification of missing data that are needed to make decisions about the geothermal resource potential of the area. Typically one will need to collect chemical data from springs and existing wells, do reconnaissance-scale geologic mapping in part of the region under consideration and log temperatures in available wells. Geochemistry and temperature logging are used to search for thermal signatures in the ground water, while the geologic mapping and age dating are used to identify evidence for recent volcanism and intrusion. The new data are added to the data base, and an integrated interpretation identifies portions of the region for a second round of data collection while eliminating the low-priority areas. At this stage, geologic mapping and age dating locate specific areas of recent volcanism and structural settings conducive to development of permeability in the subsurface. Soil and/or gas geochemistry may be used to detect emanations from underlying geothermal systems. A limited amount of drilling may be done for the purpose of detecting anomalously high temperature gradients in the subsurface, while magnetic and/or gravity data may be collected to help map structure or detect possible effects of hydrothermal alteration. Once again, integrated interpretation is performed on the total data base. The product of this step is a prioritized list of prospects for consideration in subregional or detailed exploration programs, as discussed by Goldstein (1988).

ACKNOWLEDGEMENTS

Much of the material for this paper was generated during our work on geothermal research and technology development sponsored by the U. S. Department of Energy under contracts DE-AC03-84SF12196 and DE-AC07-85ID12489. We gratefully acknowledge DOE's support for this work. We also acknowledge the many discussions with colleagues that have influenced our thinking about regional geothermal exploration. We especially thank James C. Bresee for his invitation to write this article and for his review of it. The illustrations were prepared by Patrick Daubner and the manuscript was typed by Kathryn Ruth.

REFERENCES

Adams, M. C. and J. M. Moore, 1987, Hydrothermal alteration and fluid geochemistry of the Meager Mountain geothermal system, *American Journal of Science, 267,* 720-755 pp.

Adams, M. C., M. M. Lemieux, J. N. Moore and S. D. Johnson, 1988, Fluid chemistry and hydrology of the Heber geothermal system, California, *Geothermal Resources Council Transactions, 12,* in press.

Aldrich, M. J., 1986, Tectonics of the Jemez lineament in the Jemez Mountains and Rio Grande Rift, *Journal of Geophysical Research, 91,* 1753-1763 pp.

Allison, M. L. and D. L. Nielson, 1987, Survey of borehole breakouts from active geothermal systems, *EOS, Transactions American Geophysical Union, 68,* 1460-1461 pp.

Allmendinger, R. W., J. W. Sharp, D. Von Tish, L. Serpa, L. Brown, J. E. Oliver and S. Kaufman, 1983, COCORP seismic reflection data from the eastern Basin and Range, west-central Utah: Abstracts with Programs, *Geological Society of America, 15, No. 5,* 287 p.

Amyx, J. W., F. M. Bass, Jr. and R. L. Whiting, 1960, *Petroleum Reservoir Engineering,* McGraw-Hill, New York.

Bacon, C. R., 1981, Geology and geophysics of the Cascade Range, Presented at 41st Annual International Meeting and Exposition, *Society of Exploration Geophysicists,* Los Angeles.

Bacon, C. R., 1983, Eruption history of Mount Mazama and Crater Lake caldera, Cascade Range, U.S.A., *Journal of Volcanology and Geothermal Research, 18,* 57-115 pp.

Bacon, C. R., W. A. Duffield and K. Nakamura, 1980, Distribution of Quaternary rhyolite domes of the Coso Range, California: Implications for extent of the geothermal anomaly, *Journal of Geophysical Research, 85,* 2425-2433 pp.

Behrman, P. G. and R. B. Knapp, 1980, Structure of the Redondo Creek area, Baca project: Implications concerning the nature of permeability and control on production, Union Oil Co. Internal Report: University of Utah Research Institute Open File Report, 43 p.

Benoit, W. R. and R. W. Butler, 1983, A review of high-temperature geothermal developments in the northern Basin and Range Province: *Geothermal Resources Council Special Report No. 13:* in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 57-80 pp.

Bhattacharyya, B. K. and L. K. Leu, 1975, Analysis of magnetic anomalies over Yellowstone National Park: mapping of Curie-point isothermal surface for geothermal reconnaissance, *Journal of Geophysical Research, 80,* 4461-4465 pp.

Biehler, S., 1971, Gravity studies in the Imperial Valley: in Cooperative geological-geophysical-geochemical investigations of geothermal resources in the Imperial Valley area of California: University of California, Riverside, Education Resource Services, 29-41 pp.

Bird, D. K., P. Schiffman, W. A. Elders, A. E. Williams and S. D. McDowell, 1984, Calc-silicate mineralization in active geothermal systems, *Economic Geology, 79,* 671-695 pp.

Bird, J. M. and Bryan Isacks, (eds.), 1980, Plate Tectonics: in Selected Papers for the Journal of Geophysical Research, *American Geophysical Union, Washington,* D. C.

Blackwell, D. D., 1983, Heat flow in the northern Basin and Range Province: *Geothermal Resources Council Special Report No. 13:* in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 81-92 pp.

Blackwell, D. D., 1988, U. S. geothermal database and Oregon Cascade thermal studies: Southern Methodist University Report to U.S.D.O.E., Contract No. DE-FG07-86ID12623, 17 p.

Bott, M. H. P., 1982, *The interior of the earth: Its structure, constitution and evolution,* Edward Arnold, Publishers, Ltd., London 403 p.

Brace, W. F., 1980, Permeability of crystalline and argillaceous rocks, *International Journal of Rock Mechanics, Mineral Science & Geomechanics Abstracts, 17,* 241-251 pp.

Browne, P. R. L., 1977, Occurrence and hydrothermal alteration of diabase, Heber geothermal field, Imperial Valley, California: Institute of Geophysics and Planetary Physics, University of California, Riverside, UCR/IGPP-77/9, 61.

Burke, K. C. and J. T. Wilson, 1976, Hot spots on the Earth's surface: in *Continents Adrift and Continents Aground, Scientific American,* 58-66 pp.

Capuano, R. M. and J. N. Moore, 1980, Hg and As soil geochemistry as a technique for mapping permeable structures over a hot-water geothermal system, Abstracts with Programs, *Rocky Mountain Section of the Geological Society of America, 12,* 269 p.

Christensen, O. D., R. M. Capuano and J. N. Moore, 1983, Trace element distribution in an active hydrothermal system, Roosevelt Hot Springs thermal area, Utah, *Journal of Volcanology and Geothermal Research, 16,* 99-129 pp.

Christiansen, R. L. and E. H. McKee, 1978, Late Cenozoic volcanic and tectonic evolution of the Great Basin and Columbia Intermontaine regions: in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, (Smith, R. B. and G. P. Eaton, eds.), *Geological Society of America Memoir 152,* 283-311 pp.

Coplan, T. B., P. Kolesar, R. E. Taylor, C. Kendall and C. Mooser, 1975, Investigations of the Dunes geothermal anomaly, Imperial Valley, California, Part IV, Geochemical studies of water, calcite, and silicates: Institute of Geophysics and Planetary Physics, University of California, Riverside, UCR/IGPP-75/20, 43 p.

Costain, J. K., L. Glover, III and A. K. Sinha, 1980, Low temperature geothermal resources in the eastern United States, *EOS, Transactions, American Geophysical Union, 61,* 1-13 pp.

Craig, H., 1963, The isotopic geochemistry of water and carbon in geothermal areas: in *Nuclear Geology of Geothermal Areas,* Spoleto: Pisa, Consiglio Nazionale Delle Recerche, Laboratorie de Geologia Nucleare, 17-53 pp.

Cummings, D., 1968, Mechanical analysis of the effect of the Timber Mountain caldera on basin and range faults, *Journal of Geophysical Research, 73,* 2787-2794 pp.

de Boer, J., 1979, Paleomagnetism of the Quaternary Cerro Prieto, Crater Elegante, and Salton Buttes volcanic domes in the northern part of the Gulf of California rhombochasm: in *Proceedings of the Second Symposium on the Cerro Prieto Geothermal Field,* Com. Fed. de Electr., Mexicali Mexico, 91-98 pp.

Dibblee, T. W., Jr., 1954, Geology of the Imperial Valley region, California: in Geology of Southern California, (R. H. Jahns, ed.), *California Division of Mines and Geology Bulletin, 170,* 21-28 pp.

Doell, R. R., G. B. Dalrymple, R. L. Smith and R. A. Bailey, 1968, Paleomagnetism, potassium-argon ages, and geology of rhyolite and associated rocks of the Valles caldera, New Mexico, *Geological Society of America Memoir 116*, 211-248 pp.

Drury, S. A., 1987, *Image Interpretation in Geology*, Alley & Unwin, London, 243 p.

Durrance, M. E., 1986, *Radioactiviy in Geology: Principles and Applications*, John Wiley & Sons, New York, 441 p.

Eaton, J. P., 1963, Crustal structure from San Francisco, California to Eureka, Nevada, from seismic-refraction measurements, *Journal of Geophysical Research, 68*, 5789-5806 pp.

Edmiston, R. C. and W. R. Benoit, 1985, Characteristics of Basin and Range geothermal systems with fluid temperatures of 150° to 200°C, *Geothermal Resources Council Bulletin, 14*, N. 4, 3-10 pp.

Elders, W. A., 1979, The geological background of the geothermal fields of the Salton trough: in *Geology and Geothermics of the Salton Trough*, (W. A. Elders, ed.), Campus Museum Contribution No. 5, University of California, Riverside, 1-19 pp.

Elders, W. A. and L. H. Cohen, 1983, The Salton Sea geothermal field, California, as a near-field natural analog of a radioactive waste repository in salt: Report to Columbus, Battelle Memorial Institute, Office of Nuclear Waste Isolation, BMI/ONWI-513, 146 p.

Elders, W. A., J. R. Hoagland, S. D. McDowell and R. J. M. Cobo, 1978, Hydrothermal mineral zones in the geothermal reservoir of Cerro Prieto: in *Proceedings of the First Symposium on the Cerro Prieto Geothermal Field, Baja, California, Mexico*, September, 1978: Lawrence Berkeley Laboratory, LBL-7098, 68-75 pp.

Elders, W. A., R. Rex, T. Meidav, P. T. Robinson and S. Biehler, 1972, Crustal spreading in southern California, *Science, 178*, 15-24 pp.

Ellis, A. J. and W. A. J. Mahon, 1977, *Chemistry and Geothermal Systems*, Academic Press, Inc, 392 p.

Evans, S. H. and D. L. Nielson, 1982, Thermal and tectonic history of the Mineral Mountains intrusive complex, *Geothermal Resources Council Transactions, 6*, 15-18 pp.

Fiore, J. N., 1980, Overview and status of the U. S. Department of Energy Industry Coupled Geothermal Reservoir Assessment Program, *Geothermal Resources Council Transactions, 4*, 201-204 pp.

Flovenz, O. G. and L. S. Georgeson, 1982, Prospecting for near vertical aquifers in low-temperature geothermal areas in Iceland, *Geothermal Resources Council Transactions, 6*, 19-22 pp.

Freeze, R. A. and J. A. Cherry, 1979, *Groundwater*, Prentice-Hall, Englewood Cliffs, N.J., 604 p.

Fournier, R. O., 1977, Chemical geothermometers and mixing models for geothermal systems: in Proceedings of the International Atomic Energy Agency Advisory Group on the Application of Nuclear techniques to Geothermal Studies, Pisa, 1975, *Geothermics, Special Issue 5*, 41-50 pp.

Fournier, R. O., 1979, Geochemical and hydrologic considerations and the use of enthalpy-chloride diagrams in the prediction of underground conditions in hot-spring systems, *Journal of Volcanology and Geothermal Research, 5*, 1-16 pp.

Fournier, R. O., 1981, Application of water geochemistry to geothermal exploration and reservoir engineering in geothermal systems: in *Principles and Case Histories*, (L. Ryback and L. P. J. Muffler, eds.), John Wiley and Sons, New York, 109-143 pp.

Fournier, R. O. and R. W. Potter, II, 1979, Magnesium correction to the Na-K-Ca chemical geothermometer, *Geochemica Cosmochemica Acta, 43*, 1543-1550 pp.

Fournier, R. O. and J. J. Rowe, 1966, Estimation of underground temperatures from the silica content of water from hot springs and wet-steam wells, *American Journal of Science, 264*, 685-697 pp.

Fournier, R. O. and A. H. Truesdell, 1973, An empirical Na-K-Ca geothermometer for natural waters, *Geochemica Cosmochemica Acta, 37*, 1255-1275 pp.

Fournier, R. O. and A. H. Truesdell, 1974, Geochemical indicators of subsurface temperature, 2. Estimation of temperature and fraction of hot water mixed with cold water, *U.S. Geological Survey Journal of Research, 2*, 263-270 pp.

Fuis, G. S., W. D. Mooney, J. H. Healy, G. A. McMechan and W. J. Lutter, 1984, A seismic refraction survey of the Imperial Valley region, California, *Journal of Geophysical Research, 89(B2)*, 1165-1189 pp.

Gans, P. B., 1982, Geometry of mid Tertiary extensional faulting, northern Egan Range, east-central Nevada, Abstracts with Programs, *Geological Society of America*, *14*, No. 4, 165 p.

Gardner, J. N., F. E. Goff, S. Garcia and R. C. Hagen, 1986, Stratigraphic relations and lithologic variations in the Jemez volcanic field, New Mexico, *Journal of Geophysical Research*, *91*, 1763-1778 pp.

Goff, F. E. and L. Shevenell, 1987, Travertine deposits of Soda Dam, New Mexico, and their implications for the age and evolution of the Valles caldera hydrothermal system, *Geological Society of America Bulletin*, *99*, 292-302 pp.

Goff, F. E., J. N. Gardner, R. Vidale and R. Charles, 1985, Geochemistry and isotopes of fluids from Sulphur Springs, Valles caldera, New Mexico, *Journal of Volcanology and Geothermal Research*, *23*, 2734-2797 pp.

Goldstein, N. E., 1988, Subregional and detailed exploration for geothermal-hydrothermal resources, *Geothermal Science and Technology*, *4*, in press.

Goldstein, N. E., M. J. Wilt and D. J. Corrigan, 1984, Analysis of the Nuevo Leon magnetic anomaly and its possible relation to the Cerro Prieto magmatic-hydrothermal system, *Geothermics*, *13*, 3-11 pp.

Griscom, A. and L. J. P. Muffler, 1971, Aeromagnetic map and interpretation of the Salton Sea geothermal area, California, *U.S. Geological Survey Geophysical Investigations*, Map GP 754.

Griscom, A. and C. W. Roberts, 1983, Gravity and magnetic interpretation of Newberry Volcano, Oregon: in *Survey of Potential Geothermal Sites at Newberry Volcano, Deschutes County, Oregon*, (Priest, G. R., B. F. Vogt and G. L. Black, eds.), Oregon Department of Geology and Mineral Industries, Open File Report 0-83-3, 68-81 pp.

Halfman, S. E., M. J. Lippmann, R. Zelwer and J. H. Howard, 1984, Geologic interpretation of geothermal fluid movement in Cerro Prieto field, Baja California, Mexico, *American Association of Petroleum Geologists Bulletin*, *68*, 18-30 pp.

Hamilton, R. M. and L. J. P. Muffler, 1972, Microearthquakes at The Geysers geothermal area, California, *Journal of Geophysical Research*, *77*, 2081-2086 pp.

Harris, P. G., W. Q. Kennedy and C. M. Scarfe, 1970, Volcanism vs. plutonism - The effect of chemical composition: in *Mechanism of Igneous Intrusion*, (Newall, G. and N. Rast, eds.), Gallery Press, Liverpool, 187-200 pp.

Harrison, T. M., P. Morgan and D. D. Blackwell, 1986, Constraints on heating at the Fenton Hill site, Valles caldera, New Mexico, *Journal of Geophysical Research, 91,* 1899-1908 pp.

Hearst, J. R. and P. H. Nelson, 1985, *Well Logging for Physical Properties*, McGraw-Hill, 571 p.

Heiken, G., F. E. Goff, J. Stix, S. Tamanyu, M. Shafiqullah, S. Barcia and R. Hagan, 1986, Intracaldera volcanic activity, Toledo caldera and embayment, Jemez Mountains, New Mexico, *Journal of Geophysical Research, 91,* 1799-1815 pp.

Helgeson, H. A., 1968, Geologic and thermodynamic characteristics of the Salton Sea geothermal system, *American Journal of Science, 266,* 129-166 pp.

Henley, R. W. and A. J. Ellis, 1983, Geothermal systems ancient and modern, a geochemical review, *Earth Science Review, 19,* 1-50 pp.

Henley, R. W., A. H. Truesdell and P. B. Barton, Jr., 1984, Fluid-mineral equilibria in hydrothermal systems, *Reviews in Economic Geology, 1,* 267 p.

Hildreth, W., 1981, Gradients in silicic magma chambers: Implications for lithospheric magmatism, *Journal of Geophysical Research, 86,* 10153-10192 pp.

Hill, D. G., E. B. Layman, C. M. Swift and S. H. Yungul, 1979, Soda Lake, Nevada thermal anomaly, *Geothermal Resources Council Transactions, 3,* 305-308 pp.

Hill, D. P. and L. C. Pakiser, 1966, Crustal structure between the Nevada Test Site and Boise, Idaho, from seismic-refraction measurements, *American Geophysical Union Monograph 10,* 291-319 pp.

Hilpert, L. S. and R. J. Roberts, 1964, Geology--Economic geology: in *U. S. Geological Survey, Mineral and Water Resources of Utah,* U. S. 88th Congress, 2nd Session, 28-38 pp.

Hulen, J. B. and D. L. Nielson, 1986, Hydrothermal alteration in the Baca geothermal system, Redondo dome, Valles caldera, New Mexico, *Journal of Geophysical Research, 91,* 1867-1886 pp.

Hulen, J. B. and D. L. Nielson, 1988, Hydrothermal brecciation in the Jemez fault zone, Valles caldera, New Mexico - Results from CSDP Corehole VC-1, *Journal of Geophysical Research, 93* 6077-6089 pp.

Hulen, J. B. and B. S. Sibbett, 1982, Sampling and interpretation of drill cuttings from geothermal wells, *Society of Professional Well Log Analysts, Reprint Volume*, Tulsa, IV3-IV54.

Hunt, T. M. and J. H. Lattan, 1982, A survey of seismic activity near Wairakei geothermal field, New Zealand, *Journal of Volcanism and Geothermal Research, 14*, 319-334 pp.

Ingle, J. C., Jr., 1982, Microfaunal evidence of age and depositional environments of the Cerro Prieto section, Plio-Pleistocene), Baja California, Mexico: Report LBL-13897, Lawrence Berkeley Laboratory, Berkeley, California.

Isherwood, W. F., 1976, Complete Bouguer gravity map of The Geysers Area, California, *U.S. Geological Survey Open File Report*, 76-357 pp.

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

James, E. D., T. Hoang and I. J. Epperson, 1987, Structure, permeability and production characteristics of the Heber, California geothermal field: in *Proceedings of the Twelfth Workshop on Geothermal Reservoir Engineering*, Stanford University, California, 267-271 pp.

Kane, M. F., D. R. Mabey and R. Brace, 1976, A gravity and magnetic investivation of the Long Valley caldera, Mono County, California, *Journal of Geophysical Research, 81*, 754-762 pp.

Keith, T. E. C., D. E. White and M. H. Beeson, 1978, Hydrothermal alteration and self-sealing in Y-7 and Y-8 drill holes in northern part of Upper Geyser Basin, Yellowstone National Park, Wyoming, *U.S. Geological Survey Prof. Paper 1054-A*, 26 p.

Keller, G. P., L. W. Braile and P. Morgan, 1979, Crustal structure, geophysical models, and contemporary tectonism of the Colorado Plateau, *Tectonophysics, 61*, 131-148 pp.

Keskinen, M. and J. Sternfeld, 1982, Hydrothermal alteration and tectonic setting of intrusive rocks from East Brawley, Imperial Valley: An application of petrology to geothermal reservoir analysis: in *Proceedings of the Eighth Workshop on Geothermal Reservoir Engineering*, Stanford University, California, 39-44 pp.

Klusman, R. W. and R. A. Landress, 1978, Secondary controls on mercury in soils of geothermal areas, *Journal of Geochemical Exploration, 9*, 75-91 pp.

Klusman, R. W. and R. A. Landress, 1979, Mercury in soils of the Long Valley, California geothermal system, *Journal of Volcanology and Geothermal Research, 5*, 49-65 pp.

Lachenbruch, A. H., 1978, Heat flow in the Basin and Range Province and thermal effects of tectonic extension, *Pure and Applied Geophysics, 117*, 34-50 pp.

Lachenbruch, A. H., J. H. Sass, R. J. Munroe and T. H. Moses, 1976, Geothermal setting and simple heat conduction models for the Long Valley caldera, *Journal of Geophysical Research, 81*, 769-784 pp.

Laidley, R. A. and D. S. McKay, 1971, Geochemical examination of obsidians from Newberry caldera, Oregon, contribution to *Mineralogy and Petrology, 40*, 336-342 pp.

Lawless, J. V., C. J. Bromely, T. M. Leach, A. C. Licup, Jr., D. M. Cope, C. M. Recio, 1983, Bacon-Manito geothermal field: A geoscientific exploration model: in *Proceedings of the 5th New Zealand Geothermal Workshop*, 97-102 pp.

Lemieux, M. M., P. M. Wright and J. N. Moore, 1988, Research coring in the Cascades -- a status report, *Geothermal Resources Council Transactions, 12*, in press.

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

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

Liaw, A. L. and W. Suyenaga, 1982, Detection of geothermal microtremors using seismic arrays: Presented at the 52nd Annual Internation Meeting and Exposition, *Society of Exploration Geophysicists*, Dallas.

Lippman, M. J. and A. Mañón, 1987, The Cerro Prieto geothermal field, *Geothermal Science and Technology, 1*, 1-38 pp.

Lipman, P. W., P. D. Rowley, H. H. Mehnert, S. H. Evans, W. P. Nash and F. H. Brown, 1978, Pleistocene rhyolite of the Mineral Range, Utah: Geothermal and archeological significance, *U.S. Geological Survey Journal of Research, 6*, 133-147 pp.

Mabey, D. R., I. Zietz, G. P. Eaton, M. D. Kleinkopf, 1978, Regional magnetic patterns in part of the Cordillera in the western United States: in Cenozoic tectonics and regional geophysics of the western Cordillera, (Smith, R. B. and G. P. Eaton, eds.), *Geological Society of America Memoir 152*, 313-340 pp.

MacDonald, W. J. P. and L. J. P. Muffler, 1972, Recent geophysical exploration of the Kawerau geothermal field, North Island, New Zealand, *New Zealand Journal of Geology and Geophysics, 15*, 303-317 pp.

MacLeod, N. S. and E. A. Sammel, 1982, Newberry volcano, Oregon: A Cascade Range geothermal prospect, Oregon Department of Geology and Mineral Industries, *Oregon Geology, 44*, 123-131 pp.

MacLeod, N. S., D. R. Sherrod and L. A. Chitwood, 1982, Geologic map of Newberry volcano, Deschutes, Klamath and Lake Counties, Oregon, *U. S. Geological Survey Open File Report 82-847*, 27 p.

MacLeod, N. S., D. R. Sherrod, L. A. Chitwood and E. H. McKee, 1981, Newberry volcano, Oregon: in Guides to some volcanic terranes in Washington, Idaho, Oregon and northern California, (Johnston, D. A. and J. Donnelly-Nolan, eds.), *U.S. Geological Survey Circular 838*, 85-103 pp.

Mahon, W. A. J., 1970, Chemistry in the exploration and exploitation of hydrothermal systems: in United Nations Symposium on the Development and Utilization of Geothermal Resources, Pisa, *Geothermics, Special Issue 2*, 1310-1322 pp.

Mahon, W. A. J., L. E. Klyen and M. Rhode, 1980, Neutral sodium/bicarboante/sulfate hot waters in geothermal systems, Chinetsu, *Journal of the Japan Geothermal Energy Association, 17*, 11-24 pp.

Mariner, R. H., T. S. Presser and W. C. Evans, 1983, Geochemistry of active geothermal systems in the northern Basin and Range Province: *Geothermal Resources Council, Special Report No. 13*, in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 95-119 pp.

Matlick, J. S. and P. R. Buseck, 1976, A new exploration method for geothermal sources using mercury: in *Proceedings of the Second United Nations Symposium on the Development and Use of Geothermal Resources*, 785-792 pp.

McDowell, S. D. and W. A. Elders, 1979, Geothermal metamorphism of sandstone in the Salton Sea geothermal system: in *Geology and Geothermics of the Salton Sea Trough*, (W. A. Elders, editor), Campus Museum Contribution No. 5, University of California, Riverside, 70-76 pp.

McKibben, M. A., A. E. Williams, J. P. Andes, Jr., C. S. Oakes, S. Okubo and W. A. Elders, 1986, Metamorphosed Plio-Pleistocene evaporites in the Salton Sea geothermal field, Salton Trough Rift, California, Abstracts with Programs, *Geological Society of America 18*, 690 p.

Moore, J. N and M. C. Adams, 1988, Evolution of the thermal cap in two wells from the Salton Sea geothermal system, California, *Geothermics*, in press.

Moore, J. N. and S. M. Samberg, 1979, Geology of the Cove Fort/Suphurdale KGRA: University of Utah Research Institute Earth Science Laboratory Report 18, Salt Lake City, 44 p.

Morse, J. G. and L. D. Thorsen, 1978, Reservoir engineering study of a portion of the Salton Sea geothermal field, *Geothermal Resources Council Transactions, 2*, 471-474 pp.

Moskowitz, B. and D. L. Norton, 1977, A preliminary analysis of intrinsic fluid and rock resistiviy in active hydrothermal systems, *Journal of Geophysical Research, 82*, 5787-5795 pp.

Muffler, L. J. P., editor, 1979, Assessment of Geothermal Resources of the United States-1978, *U. S. Geological Survey Circular 790*, 163 p.

Muffler, L. J. P. and B. R. Doe, 1968, Composition and mean age of detritus of the Colorado River delta in the Salton Trough, southeastern California, *Journal of Sedimentary Petrology, 38*, 384-399 pp.

Muffler, L. J. P. and D. E. White, 1969, Active metamorphism of upper Cenozoic sediments in the Salton Trough, southeastern California, *Geological Society of America Bulletin, 80*, 157-182 pp.

Muller, O. H. and D. D. Pollard, 1977, The state of stress near Spanish Peaks, Colorado determined from a dike pattern, *Pure Applied Geophysics, 115*, 69-86 pp.

Nakamura, K., 1977, Volcanoes as possible indicators of tectonic stress orientation-principal and proposal, *Journal of Volcanology and Geothermal Research, 2*, 1-16 pp.

Nielson, D. L. and J. B. Hulen, 1984, Internal geology and evolution of the Redondo Dome, Valles caldera, New Mexico, *Journal of Geophysical Research, 89*, 8695-8711 pp.

Nielson, D. L., S. H. Evans and B. S. Sibbett, 1986, Magmatic, structural and hydrothermal evolution of the Mineral Mountains intrusive complex, Utah, *Geological Society of America, 97*, 765-777 pp.

Nielson, D. L., B. S. Sibbett, D. B. McKinney, J. B. Hulen, J. N. Moore and S. M. Samberg, 1978, Geology of Roosevelt Hot Springs KGRA, Beaver County, Utah: University of Utah Research Institute Earth Science Laboratory, Report 12, Salt Lake City, 121 p.

Norton, D. L., 1979, Transport phenomena in hydrothermal systems: The redistribution of chemical components around cooling magmas, *Bulletin of Mineralogy, 102*, 471-486 pp.

Norton, D. L., 1984, Theory of hydrothermal systems, *Annals Reviews of Earth and Planetary Science, 12*, 155-177 pp.

Norton, D. L., 1988, Metasomatism and permeability, *American Journal of Science, 288*, 604-618 pp.

Norton, D. L. and R. Knapp, 1977, Transport phenomena in hydrothermal systems: Nature of porosity, *American Journal of Science, 277*, 913-936 pp.

Ode, H., 1957, Mechanical analysis of the dike pattern of the Spanish Peaks area, Colorado, *Geological Society of America Bulletin, 68*, 567-576 pp.

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

Pàlmâson, G., 1976, Geophysical methods in geothermal exploration: in *Proceedings of the 2nd United Nations Symposium on the Development and Use of Geothothermal Resources*, 1975, San Francisco, 2, 1175-1184 pp.

Priest, G. R., 1983, Geology of the Newberry volcano area, Deschutes County, Oregon: in *Survey of Potential Geothermal Sites at Newberry Volcano, Deschutes County, Oregon*, (Priest, G. R., B. F. Vogt and G. L. Black, eds.), Oregon Department of Geology and Mineral Industries Open File Report 0-83-3, 5-20 pp.

Priestley, K. F., A. S. Ryall and G. S. Fezie, 1982, Crust and upper mantle structure in the northwest Basin and Range Province, *Seismological Society of America Bulletin, 72*, 911-923 pp.

Proffett, J. M., Jr., 1977, Cenozoic geology of the Yerington district, Nevada, and implications for the nature and origin of Basin and Range faulting, *Geological Society of America Bulletin, 88, No. 2*, 247-266 pp.

Randall, W., 1974, An analysis of the subsurface structure and stratigraphy of the Salton Sea geothermal anomaly, Imperial County, California, Ph.D. thesis, University of California, Riverside.

Robinson, P. T., W. A. Elders and L. J. P. Muffler, 1976, Quaternary volcanism in the Salton Sea geothermal field, *Geological Society of America Bulletin, 87*, 347-360 pp.

Rona, P. A. and R. P. Lowell, (eds.), 1980, Seafloor Spreading Centers--Hydrothermal Systems, *Benchmark Papers in Geology, 56*, Dowden, Hutchinson and Ross, 424 p.

Ross, H. P. and J. N. Moore, 1985, Geophysical investigations of the Cove Fort-Sulphurdale geothermal system, Utah, *Geophysics, 50*, 1732-1745 pp.

Ross, H. P., D. L. Nielson and J. N. Moore, 1982, Roosevelt Hot Springs geothermal system, Utah -- case study, *American Association of Petroleum Geologists, 66*, 879-902 pp.

Ryall, A., D. B. Slemmons and L. D. Gedney, 1966, Seismicity, tectonism, and surface faulting in the western United States during historic time, *Seismological Society of America Bulletin, 56*, 1105-1136 pp.

Ryan, M. P., 1988, The mechanics and three-dimensional internal structure of active magmatic systems: Kilauea volcano, Hawaii, *Journal of Geophysical Research, 93*, 4213-4248 pp.

Sabins, Floyd F., Jr., 1987, *Remote Sensing - Principles and Interpretation*, 2nd edition, W. H. Freemany & Co., New York, 449 p.

Sammel, E. A., 1981, Results of test drilling at Newberry Volcano, Oregon, *Geothermal Resources Council Bulletin*, 3-8 pp.

Sass, J. H., D. D. Blackwell, D. S. Chapman, J. K. Costain, E. R. Decker, L. A. Lawyer and C. A. Swanberg, 1981: Heat flow from the crust of the United States, (Touloukian, Y. S., W. R. Judd and R. F. Roy, eds.): in *Physical Properties of Rocks and Minerals, McGraw-Hill/CINDAS Data Series on Material Properties, Vol. II-2*, McGraw-Hill, 503-548 pp.

Sass, J. H., A. H. Lachenbruch, R. J. Munroe, G. W. Greene and T. H. Moses, Jr., 1971, Heat flow in the western United States, *Journal of Geophysical Research, 76*, 6367-6413 pp.

Sbar, M. L., M. Barazangi, J. Dorman, C. H. Scholz and R. B. Smith, 1972, Tectonics of the Intermountain Seimic Belt, western United States: Microearthquake seismicity and composite fault plane solutions, *Geological Society of America Bulletin, 83*, 13-28 pp.

Self, S., F. E. Goff, J. N. Gardner, J. Wright and W. M. Kite, 1986, Explosive rhyolitic volcanism in the Jemez Mountains: Vent locations, caldera development and relation to regional structure, *Journal of Geophysical Research, 91*, 1779-1798 pp.

Sharp, R., 1982, Tectonic setting of the Imperial Valley region, *U.S. Geological Survey Prof. Paper 1254*, 5-14 pp.

Shuey, R. T., D. K. Schellinger and A. C. Tripp, 1977, Curie depth determination from aeromagnetic data, *Geophysical Journal of the Royal Astronomical Society, 50*, 75-101 pp.

Sibbett, B. S. and D. L. Nielson, 1980, Geology of the central Mineral Mountains, Beaver County, Utah: University of Utah Research Institute/ESL Report No. 33, Salt Lake City, 42 p.

Slemmons, D. B., 1967, Pliocene and Quaternary crustal movements of the Basin and Range province, USA, *Journal of Geoscience, 10*, 91-101 pp.

Smith, C., 1983, Thermal hydrology and heat flow of Beowawe geothermal area, Nevada, *Geophysics, 48*, 618-626 pp.

Smith, L. and D. S. Chapman, 1983, On the thermal effects of groundwater flow, 1, Regional scale systems, *Journal of Geophysical Research, 88*, 593-608 pp.

Smith, R. B., 1978, Seismicity, crustal structure, and intra-plate tectonics of the interior of the western Cordillera: in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, (Smith, R. B. and G. P. Eaton, eds.), *Geological Society of America Memoir 152*, Boulder, Colorado, 111-144 pp.

Smith R. B. and M. Sbar, 1974, Contemporary tectonics and seismicity of the Western States with emphasis on the Intermountain seismic belt, *Geological Society of America Bulletin, 85*, 1205-1218 pp.

Smith, R. L. and R. A. Bailey, 1968, Resurgent cauldrons: in Studies In Volcanology, (Coats, R. R., R. L. Hay and C. A. Anderson, eds.), *Geological Society of America Memoir 116*, 613-662 pp.

Smith, R. L., R. A. Bailey and C. S. Ross, 1970, Geologic map of the Jemez Mountains, New Mexico, *U.S. Geological Survey Miscellaneous Investigations*, Map I-571.

Smith, R. L. and H. R. Shaw, 1975, Igneous-related geothermal systems: in Assessment of Geothermal resources of the United States - 1975, (White, D. E. and D. L. Williams, eds.), *U.S. Geolological Survey Circular, 726*, 58-83 pp.

Sorey, M. L., R. E. Lewis and F. H. Olmsted, 1978, The hydrothermal system of Long Valley caldera, California, *U.S. Geological Survey Prof. Paper, 1044-A*, 60 p.

Speed, R. C., 1983, Pre-Cenozoic tectonic evolution of northeastern Nevada: *Geothermal Resources Council Special Report No. 13:* in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 11-24 pp.

Stauber, D.A. and D. M. Boore, 1978, Crustal thickness in northern Nevada from seismic refraction profiles, *Seismological Society of America Bulletin, 68*, 1049-1058 pp.

Stauber, D. A., 1983, Crustal structure in northern Nevada from seismic refraction data: *Geothermal Resources Council Special Report No. 13:* in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 319-332 pp.

Steven, T. A. and G. P. Eaton, 1975, Environment of ore deposition in the Creede mining district, San Juan Mountains, Colorado: in Geologic, Hydrologic, and Geophysical Setting, *Economic Geology, 70*, 1023-1037 pp.

Stewart, J. H., 1983, Cenozoic structure and tectonics of the Northern Basin and Range Province, California, Nevada, and Utah: *Geothermal Resources Council Special Report No. 13:* in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 25-40 pp.

Struhsacker, E. M., 1980, The geology of the Beowawe geothermal system, Eureka and Lander Counties, Nevada: University of Utah Research Institute/ESL Report No. 37, 78 p.

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

Swanberg, C. A., 1983, Geothermal resources of rifts: A comparison of the Rio Grande rift and the Salton trough, *Tectonophysics, 94,* 659-678 pp.

Swanberg, C. A. and J. Combs, 1986, Geothermal drilling in the Cascade range: Preliminary results from 1,387 m core hole, Newberry volcano, Oregon, *EOS, Transactions, American Geophysical Union, 67,* 578-580 pp.

Truesdell, A. H., J. M. Thompson, T. B. Coplan, N. L. Nehring and C. K. Janik, 1981, The origin of the Cerro Prieto geothermal brine, *Geothermics, 10,* 225-238 pp.

United States Geological Survey, 1972, Geologic map of Yellowstone National Park, *U.S. Geological Survey Miscellaneous Geological Investigations,* Map I-711.

Van de Kamp, P. C., 1973, Holocene continental sedimentation in the Salton Basin, California, *Geological Society of America Bulletin, 84,* 827-848 pp.

Varekamp, J. C. and P. R. Buseck, 1984, Changing mercury anomalies in Long Valley, California: Indication for magma movement or seismic activity, *Geology, 12,* 283-286 pp.

Villas, R. N. and D. L. Norton, 1977, Irreversible mass transfer between circulating hydrothermal fluids and the Mayflower stock, *Economic Geology, 72,* 1471-1504 pp.

Waibel, A. F., 1987, An overview of the geology and secondary mineralogy of the high temperature geothermal system in Dixie Valley, Nevada, *Geothermal Resources Council Bulletin, 16, N. 9,* 5-11 pp.

Walker, G. W. and B. Nolf, 1981, High Lava Plain, Brothers fault zone, Harney Basin, Oregon: in Guide to some volcanic terranes in Washington, Idaho, Oregon and northern California, (Johnston, D. A. and J. Donnelly Nolan, eds.), *U.S. Geological Survey Circular, 838,* 105-111 pp.

Ward, S. H., 1983, Geophysical studies of active geothermal systems in the northern Basin and Range: *Geothermal Resources Council, Special Report No. 13:* in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 121-157 pp.

Ward, S. H., H. P. Ross and D. L. Nielson, 1981, Exploration strategy for high-temperature hydrothermal systems in Basin and Range province, *American Association of Petroleum Geologists, 65,* 86-102 pp.

Watson, K. and Robert D. Regan, (eds.), 1983, *Remote Sensing,* Geophysics Reprint Series, 3, Society of Exploration Geophysicists, 581 p.

Wernicke, Brian, 1981, Low-angle normal faults in the Basin and Range province: Nappe tectonics in an extending orogen, *Nature, 291, No. 5817,* 645-648 pp.

White, D. E., 1965, Geothermal energy, *U. S. Geological Survey Circular,* 419 p.

White, D. E., 1968, Environments of generation of some base metal ore deposits, *Economic Geology, 63,* 301-335 pp.

White, D. E., 1969, Rapid heat-flow surveying of geothermal areas utilizing individual snowfalls as colorimeters, *Journal Geophysical Research, 74,* 5191-5201 pp.

White, D. E., 1981, Active geothermal systems and hydrothermal ore deposits, *Economic Geology, 75th Anniversary Volume,* 392-423 pp.

White, D. E. and D. L. Williams, (eds.), 1975, Assessment of geothermal resources of the United States - 1975, *Geological Survey Circular, 726,* 155 p.

White, D. E., L. J. P. Muffler and A. H. Truesdell, 1971, Vapor dominated hydrothermal systems compared with hot water systems, *Economic Geology, 66,* 75-97 pp.

Whittome, A. J. and E. W. Smith, 1979, A model of the Tongonan geothermal field: in *Proceedings of the New Zealand Geothermal Workshop,* 141-147 pp.

Williams, D. L. and R. P. Von Herzen, 1974, Heat loss from the earth: New estimate, *Geology, 2*, 327-328 pp.

Wright, P. M., 1981, Gravity and magnetic methods in mineral exploration, *Economic Geology, 75th Anniversary Volume*, 829-839 pp.

Wright, P. M., S. H. Ward, H. P. Ross and R. C. West, 1985, State-of-the-art geophysical exploration for geothermal resources, *Geophysics, 50*, 2666-2699 pp.

Wyllie, M. R. J. and M. B. Spangler, 1952, Applications of electrical resistivity measurement to problems of fluid flow in porous media, *American Association of Petroleum Geologists, 36*, 359-403 pp.

Yeamans, F. L., Jr., 1983, Basin and Range hydrology: An empirical approach: *Geothermal Resources Council Special Report No. 13:* in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 159-175 pp.

Younker, L. W., P. W. Kasameyer and J. D. Tewhey, 1982, Geological, geophysical, and thermal characteristics of the Salton Sea geothermal field, California, *Journal of Volcanology and Geothermal Research, 12*, 221-258 pp.

Zandt, G., L. S. McPherson, S. Schaff and S. Olsen, 1982, Seismic baseline and induction studies, Roosevelt Hot Springs, Utah and Raft River, Idaho: University of Utah Research Institute Report ESL-100, 58 p.

Zietz, I., R. Shuey and J. R. Kirby, Jr., 1976, Aeromagnetic map of Utah, *U. S. Geological Survey Map* GO-907.

Zoback, M. L., 1983, Shallow structure of the Sevier Desert detachment, Abstracts with Programs, *Geological Society of America 15, No. 5*, 287 p.

Zoback, M. L. and R. E. Anderson, 1983, Style of Basin-Range faulting as inferred from seismic reflection data in the Great Basin, Nevada and Utah: *Geothermal Resources Council Special Report No. 13:* in The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range Province, 363-381 pp.

Zoback, M. L. and Mark Zoback, 1980, State of stress in the conterminous United States, *Journal of Geophysical Research, 85, No. B11*, 6113-6156 pp.

Zoback, M. L., R.E. Anderson and G. A. Thompson, 1981, Cenozoic evolution of the state of stress and style of tectonism of the Basin and Range province of the western United States: in *Extensional Tectonics Associated with Convergent Plate Boundaries*, (Vine, F. J. and A. D. Smith, eds.), The Royal Society, London, 189-216 pp.

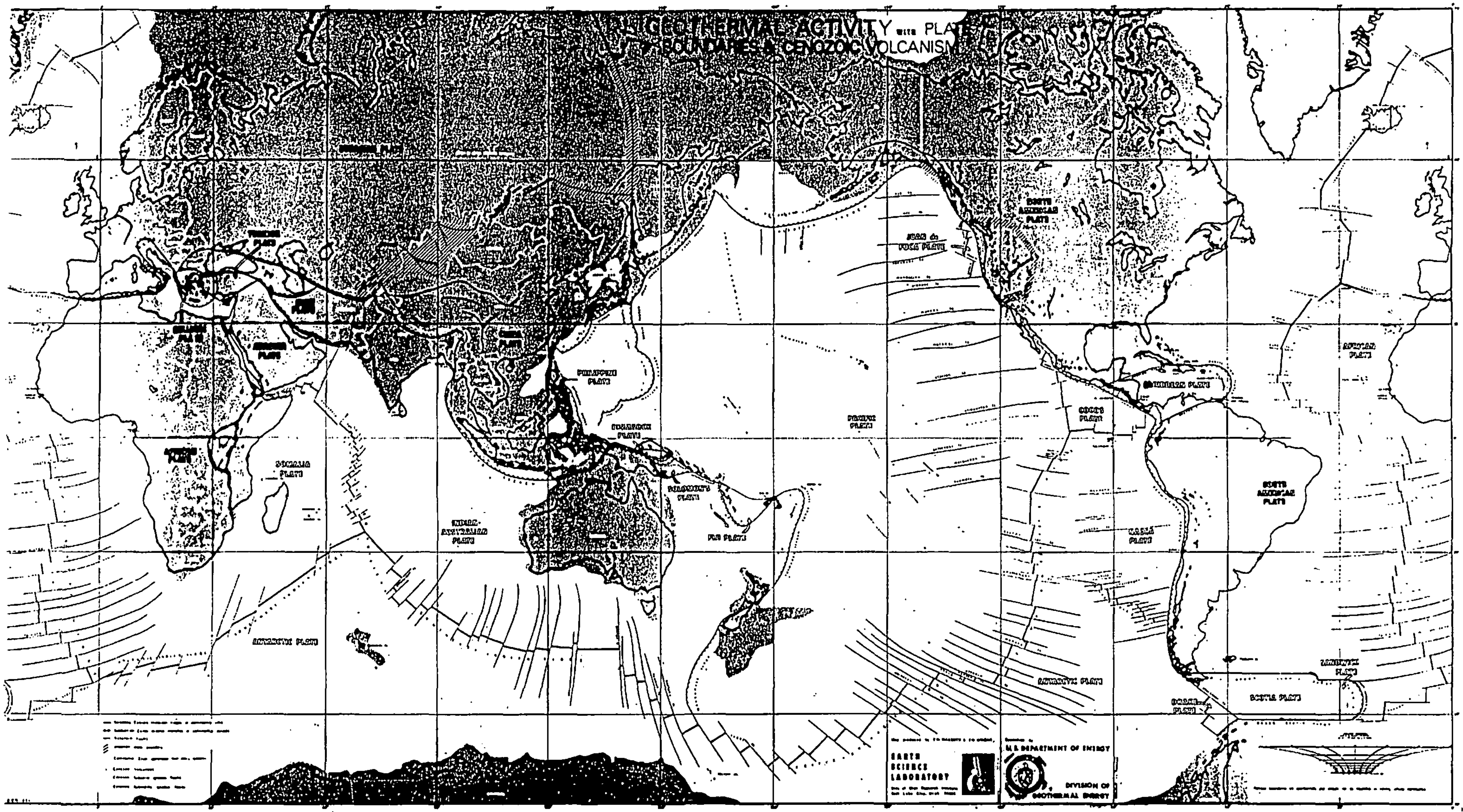


Figure 1. Geothermal occurrences and plate boundaries.

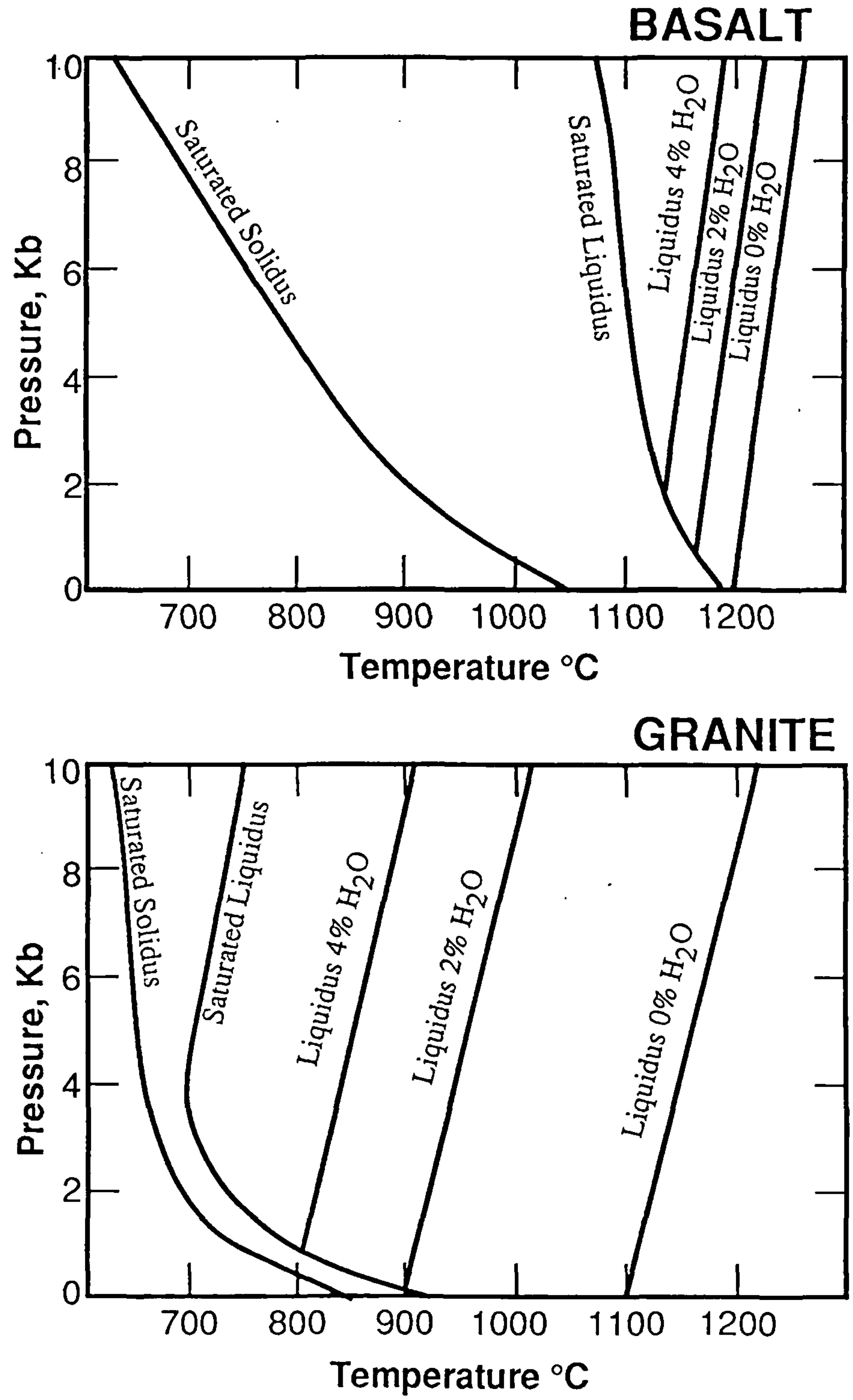


Figure 2 Phase relationships for rocks of basaltic composition (2a) and granitic compostion (2b). (from Harris, Kennedy and Scarfe, 1970).

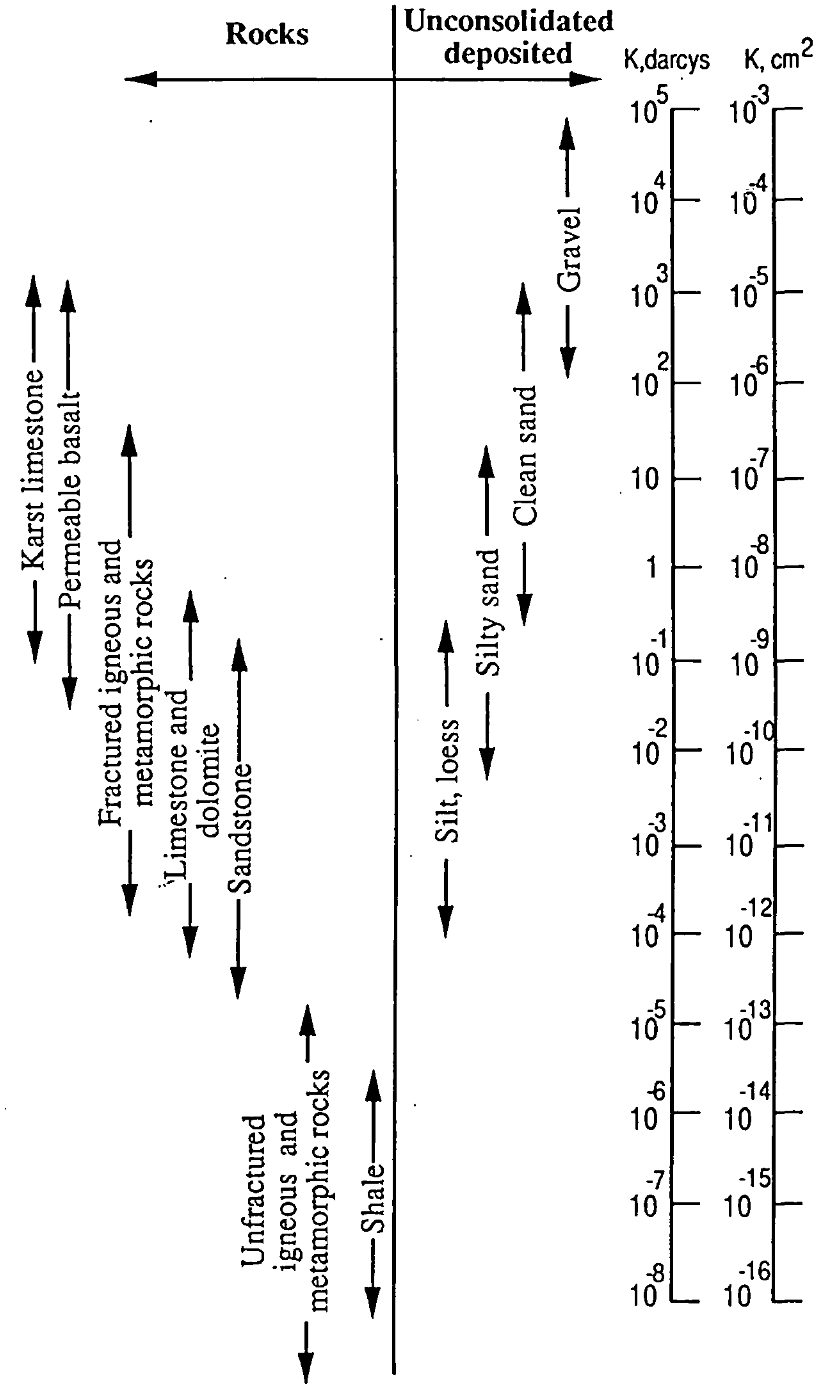


Figure 3 Ranges of permeability values in rocks.
(from Freeze and Cherry, 1979)

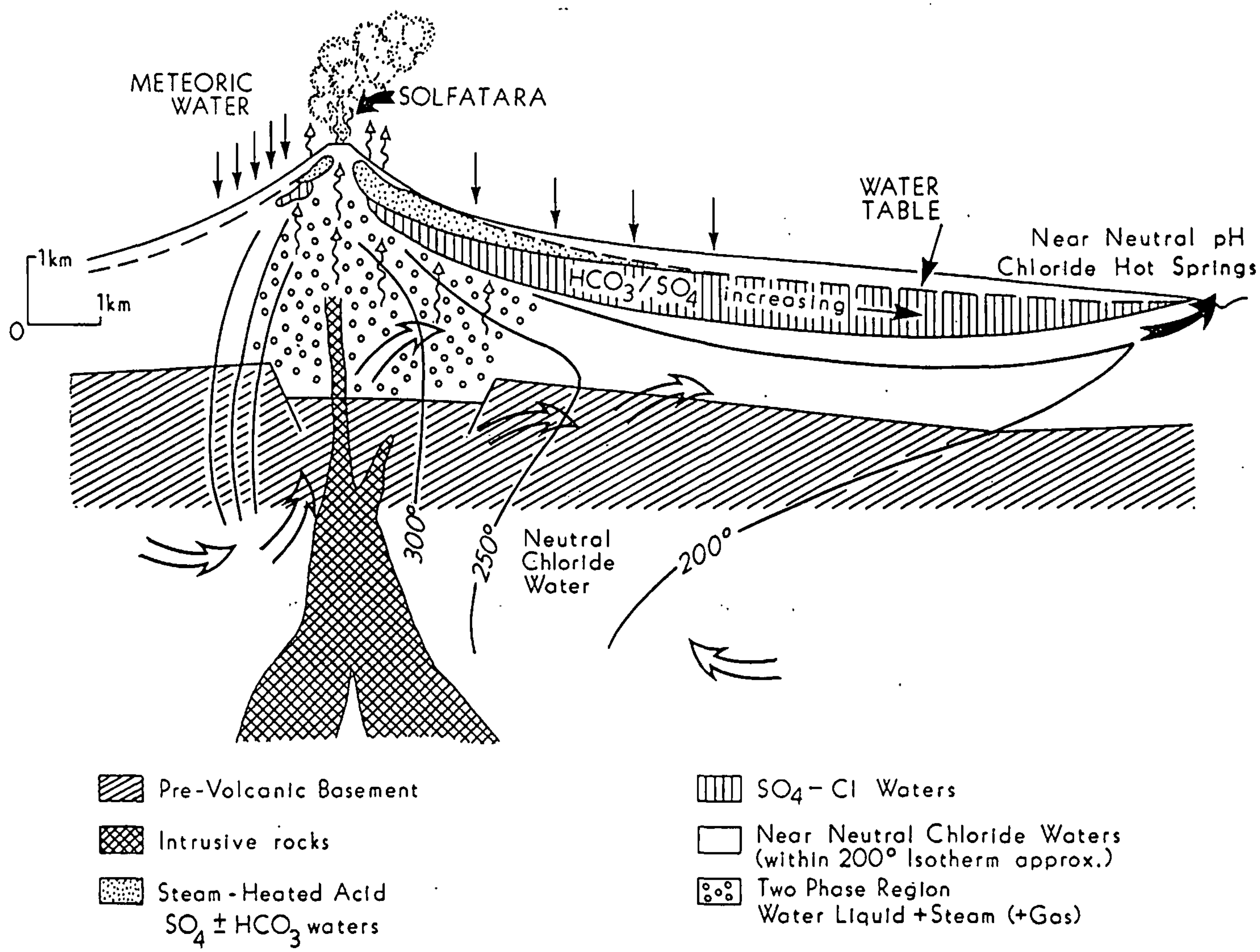


Figure 4. Hydrothermal system in a volcanic terrain. (from Mahon et al.,1980)

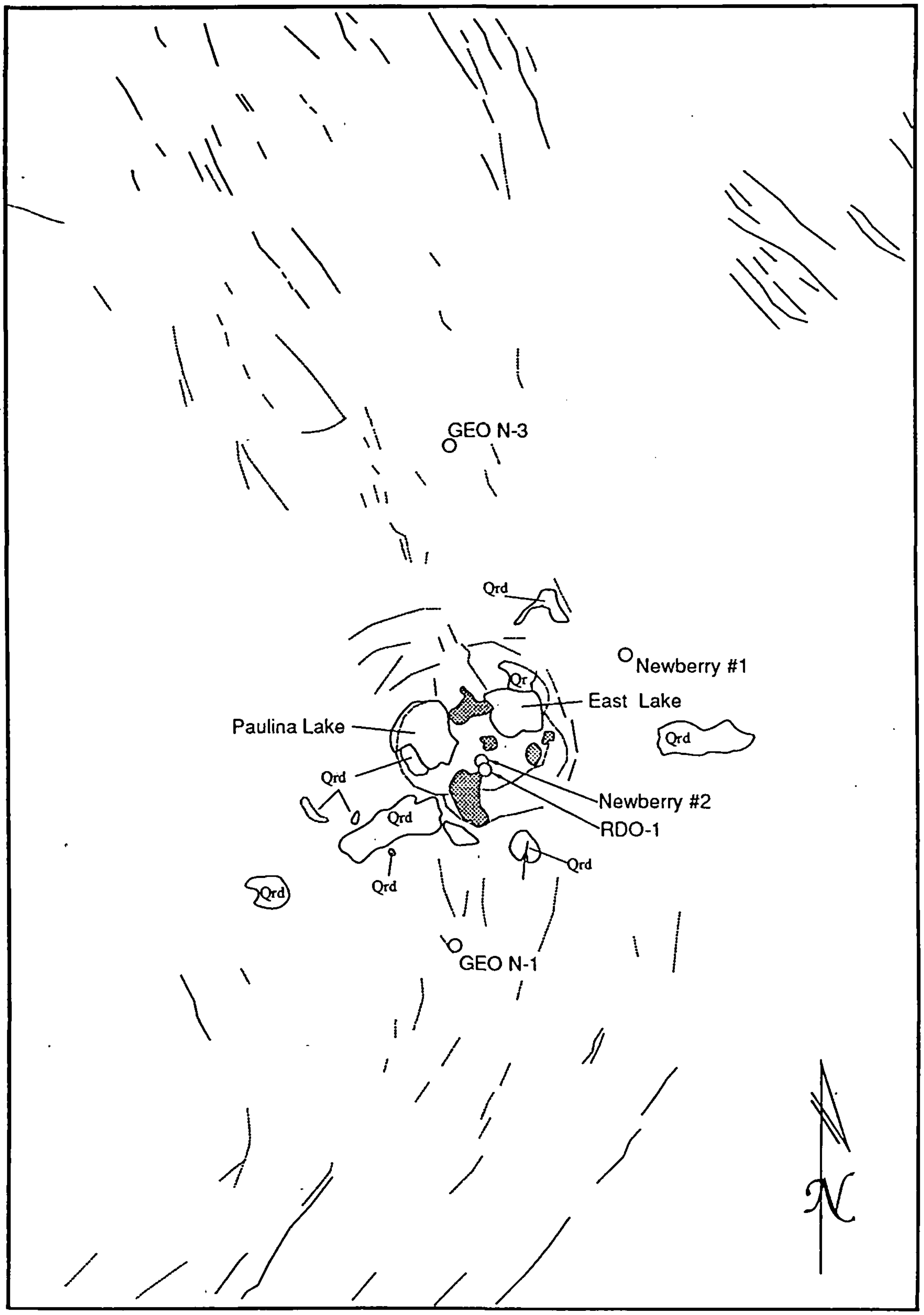


Figure 5. Geologic features in the vicinity of Newberry volcano, Oregon. NW fault trends in the top part of the figure are parallel Brothers fault zone. NE fault trends in the bottom portion of the figure are from Tumalo fault zone. Dark pattern show youngest rhyolite domes. Qrd marks pliestocene rhyolite domes and flows. (from MacLeod et al. 1982)

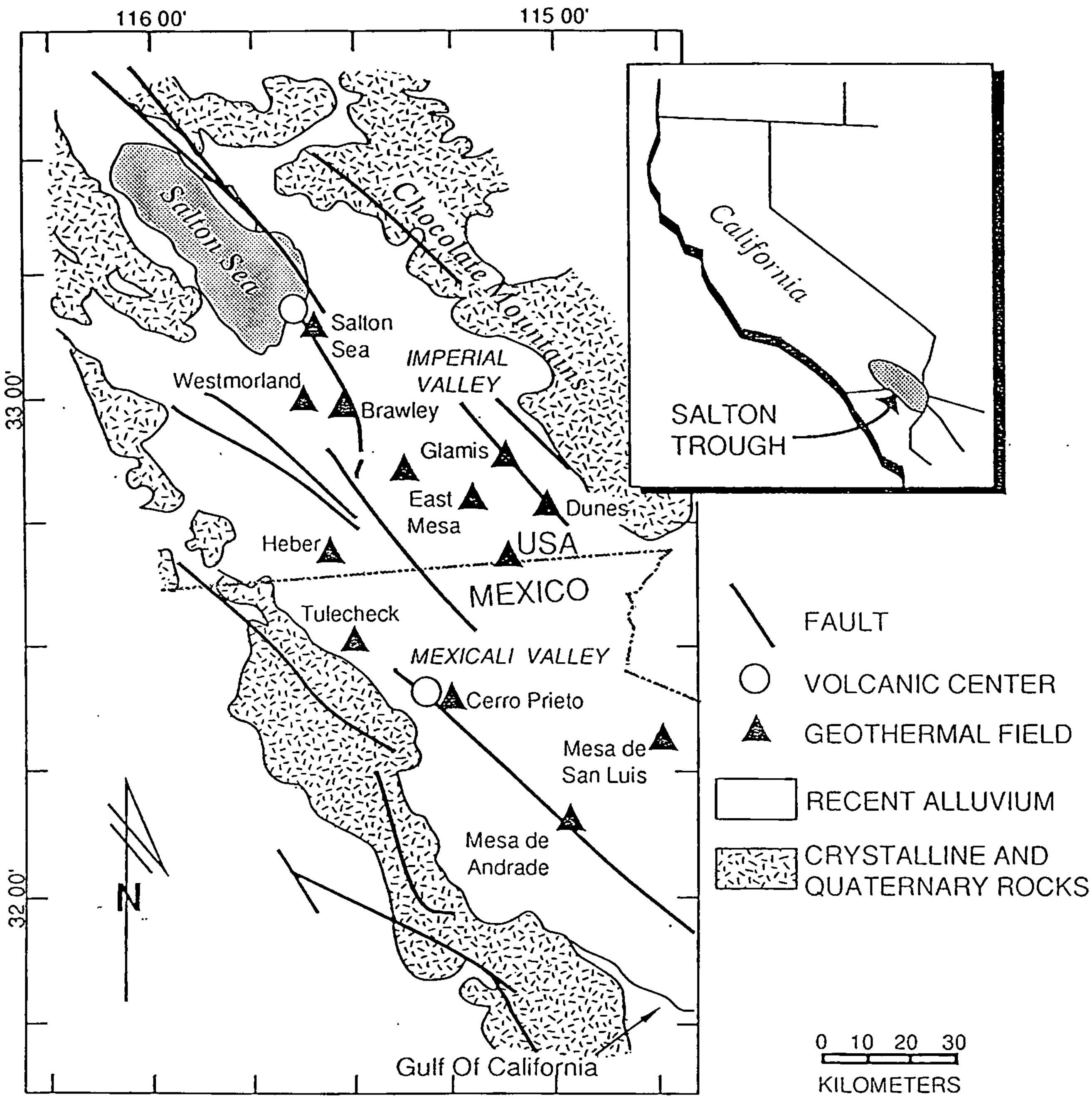


Figure 6. Geothermal systems and structure in the Salton Trough, U.S. and Mexico. (from Elders and Cohen, 1983)

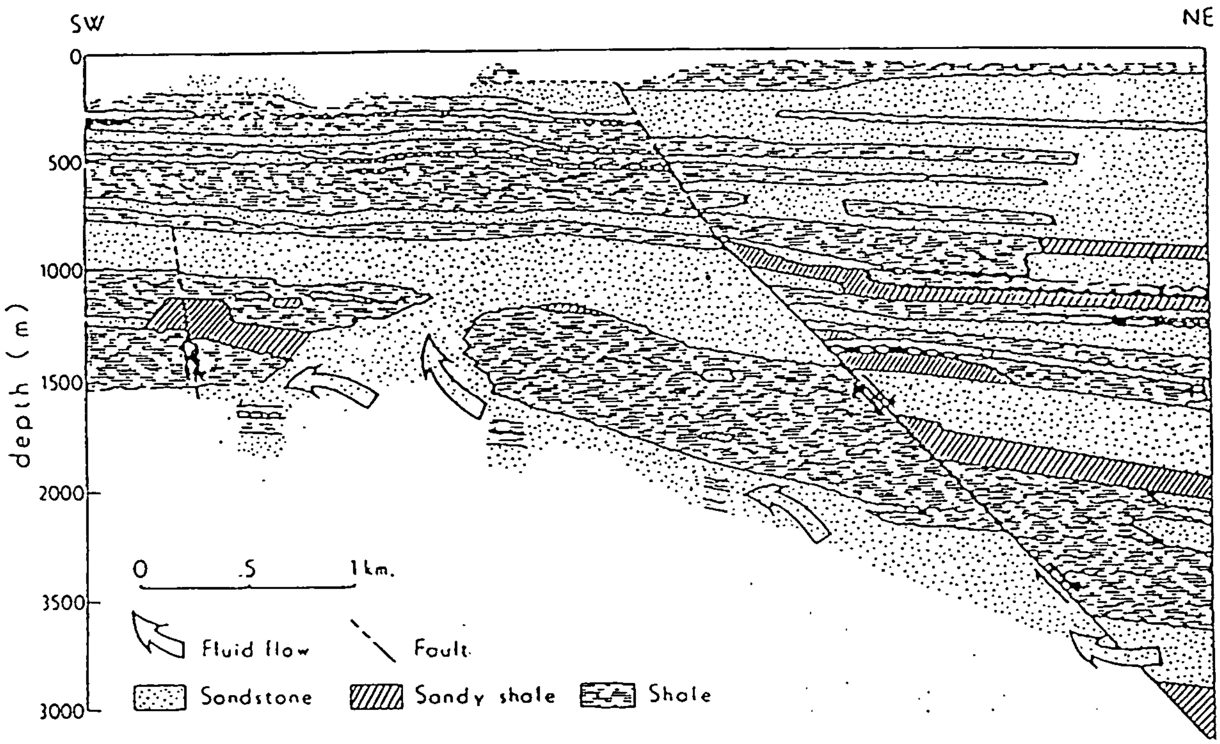


Figure 7. Fluid flow model of Cerro Prieto, Mexico. (from Halfman et al., 1984)

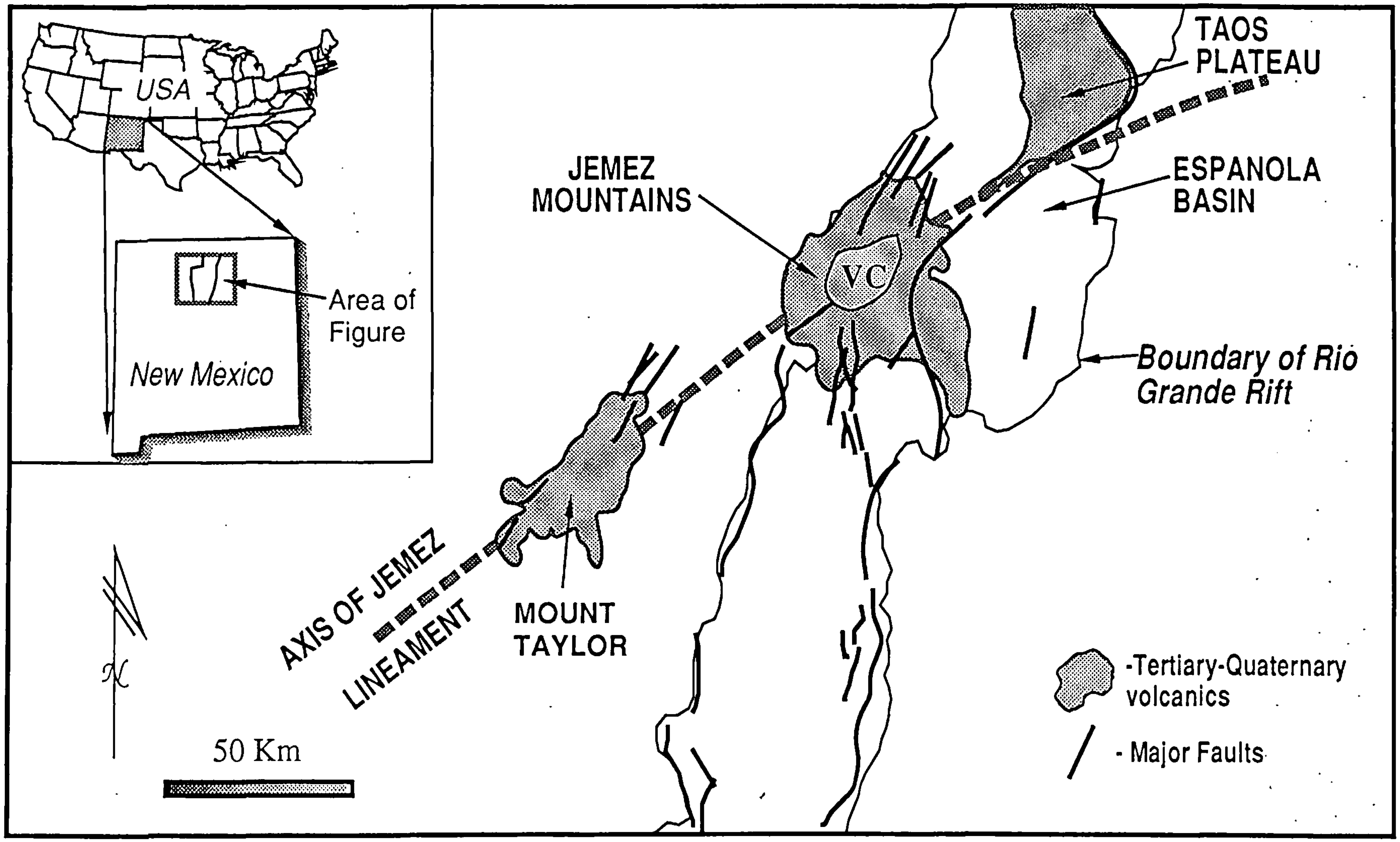
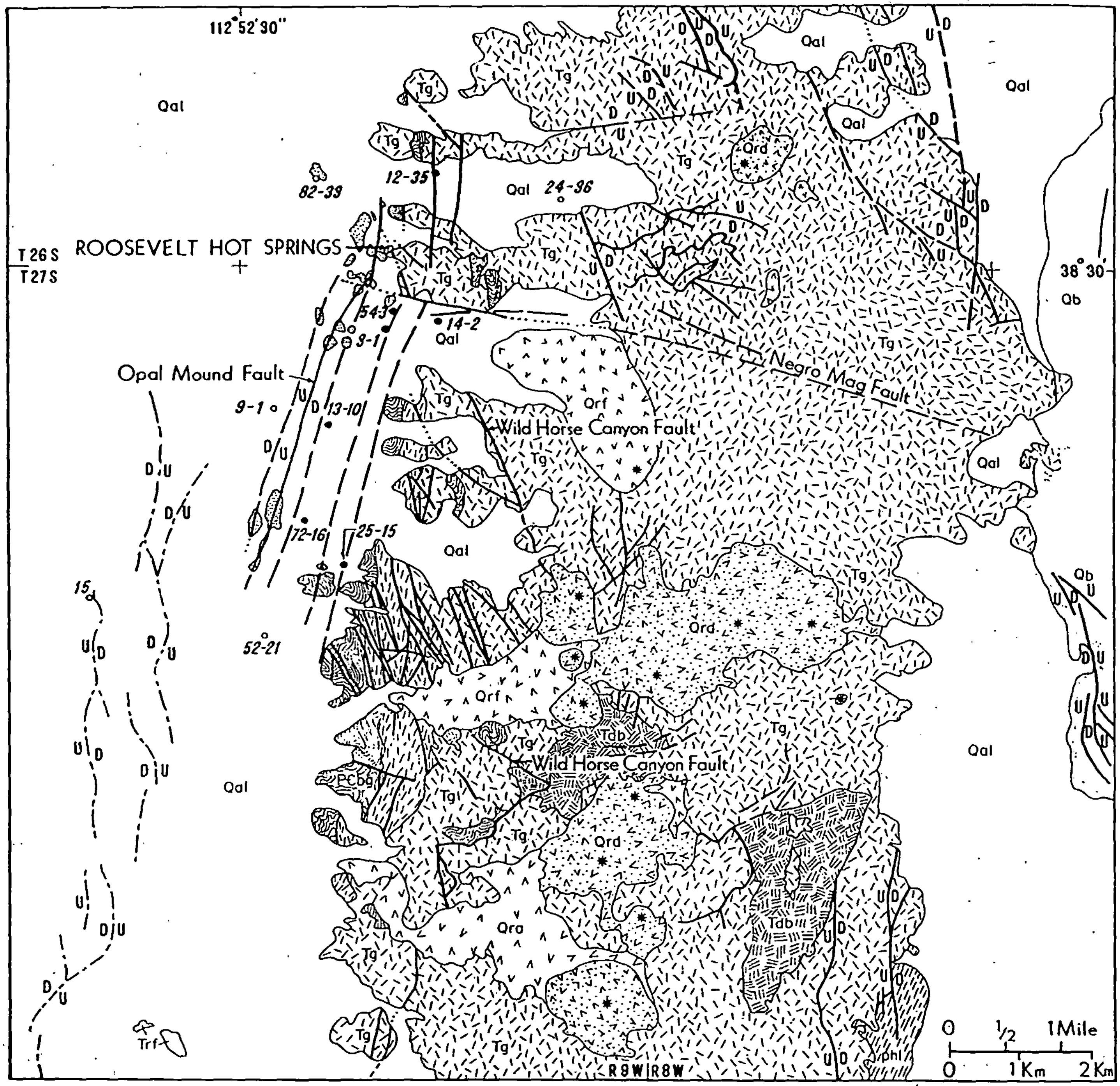


Figure 8 Regional structures in the vicinity of the Valles Caldera, New Mexico.



LEGEND

Qal	alluvium, siliceous sinter	Trf	rhyolite flows
Qb	basalt	Tg	granite, quartz monzonite, & syenite
Qrd *	rhyolite domes, with centers	Tdb	diorite
Qra	pyroclastic deposits	phl	metasediments
Qrf	rhyolite flows	PCbg	banded gneiss

Figure 9. Geologic map of the Roosevelt Hot Springs area, Utah. (from Nielson et al., 1978)

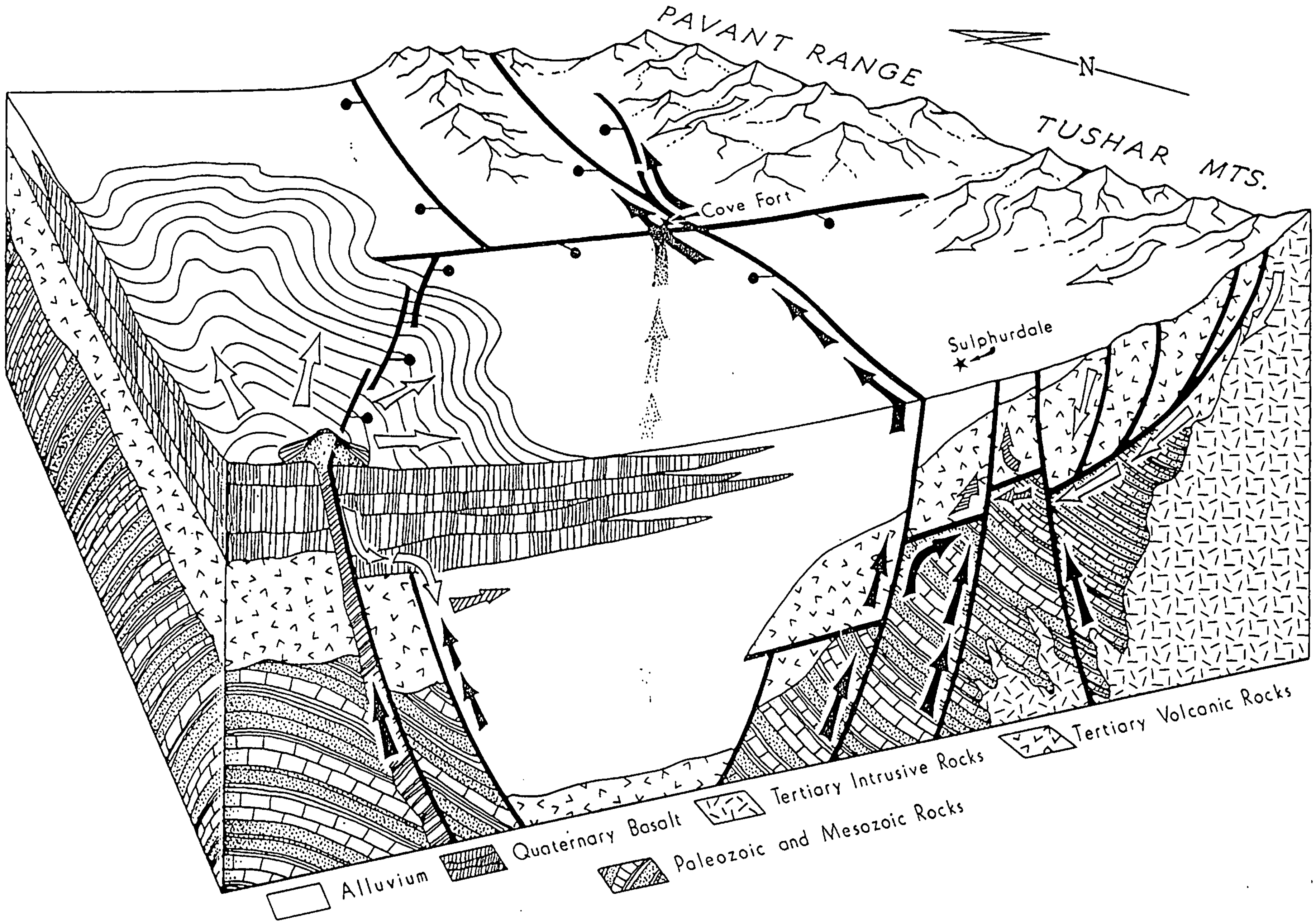


Figure 10. Fluid-flow model for the Cove Fort hydrothermal system, Utah. (from Ross & Moore, 1985)

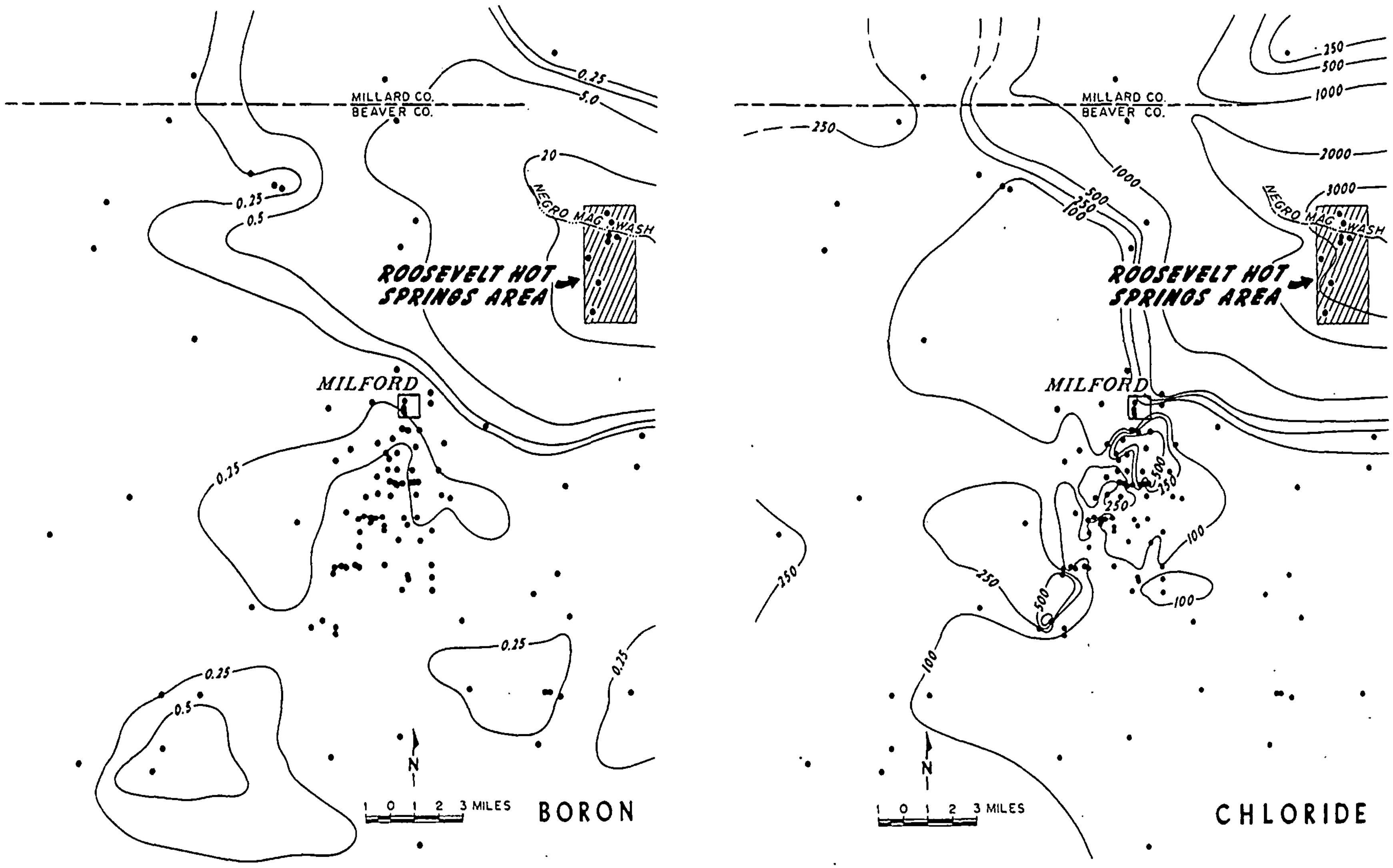


Figure 11. Boron and chloride in groundwater from wells in the vicinity of Milford, near Roosevelt Hot Springs, Utah. (from Cole, unpublished)

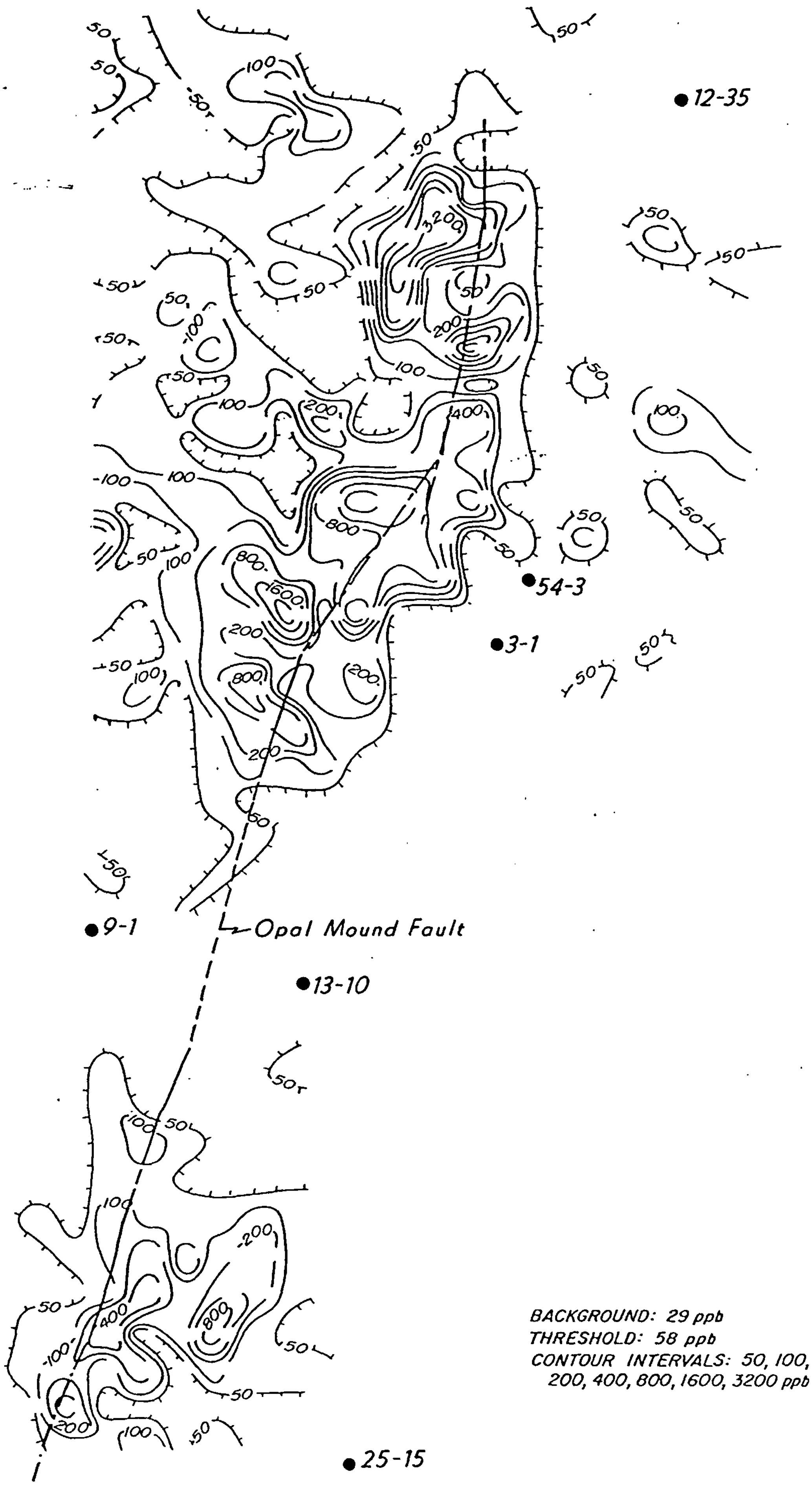


Figure 12. Mercury in soil at Roosevelt Hot Springs, Utah.

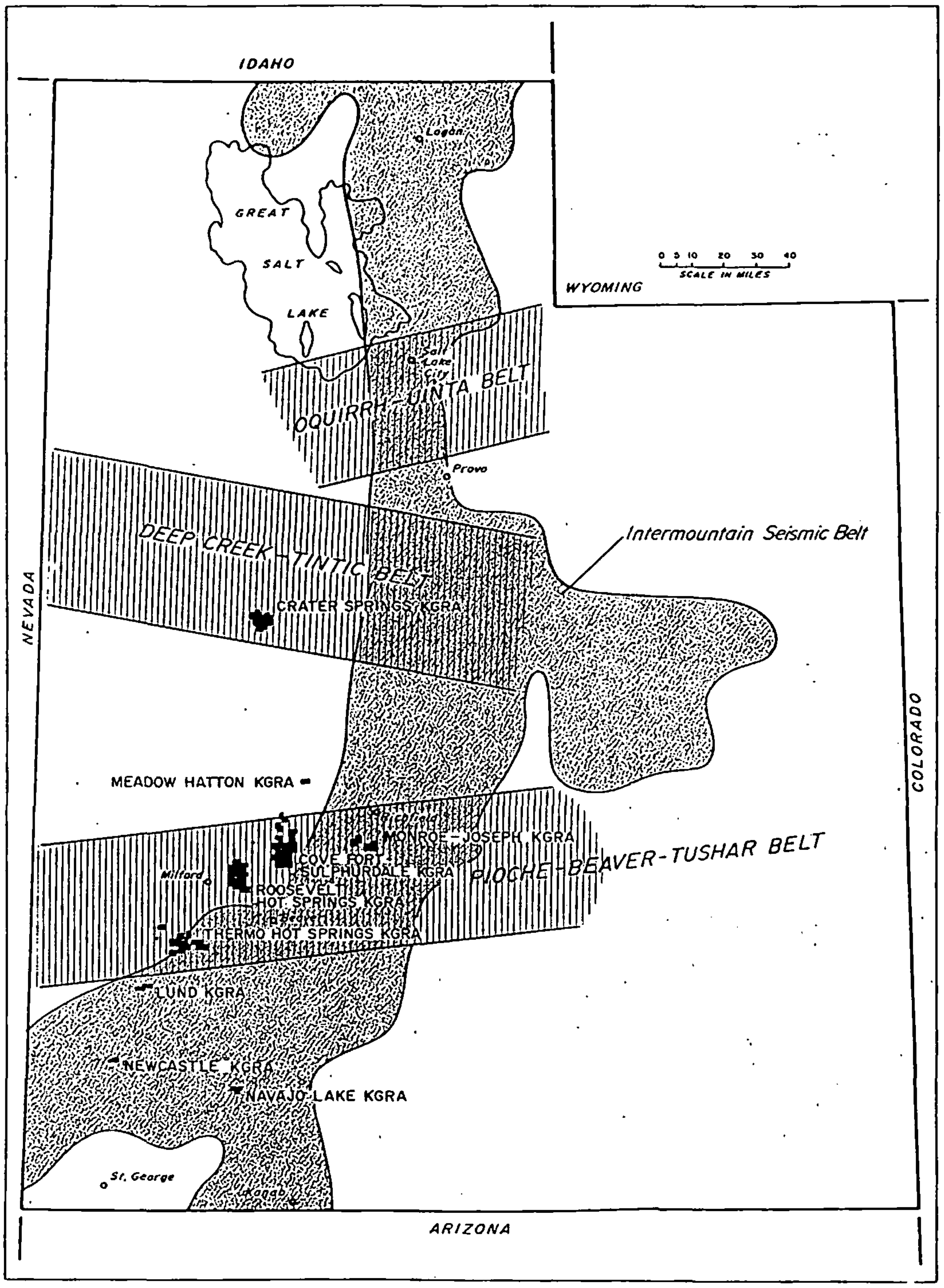


Figure 13. Distribution of tectonic zones and KGRA's in Utah. (Intermountain seismic belt from Smith and Sbar, 1974; mineral belts from Hilpert and Roberts 1964)

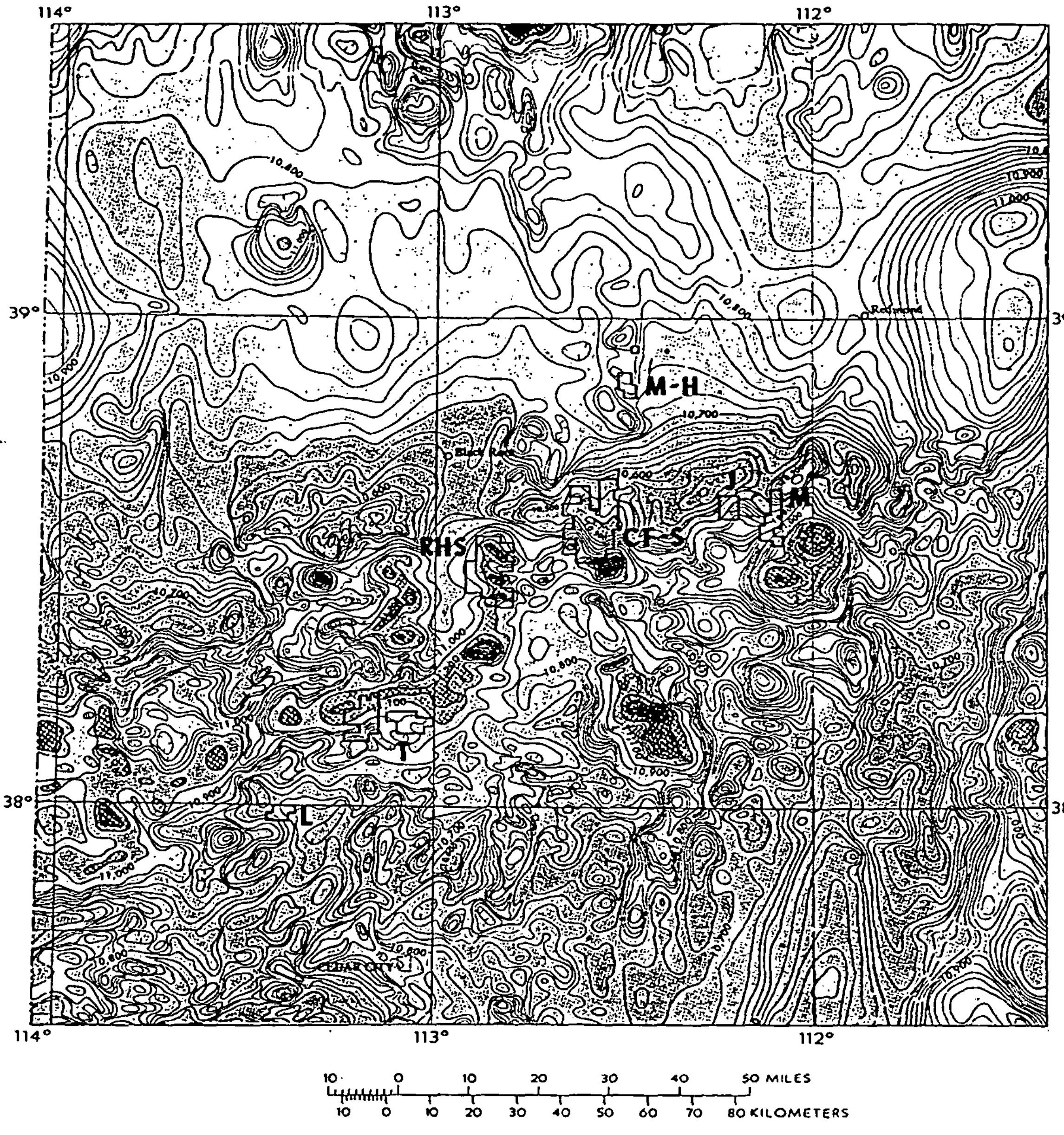


Figure 14 Aeromagnetic data across a portion of the Pioche-Beaver-Tushar trend in Central Utah. (from Zutz et al., 1976)

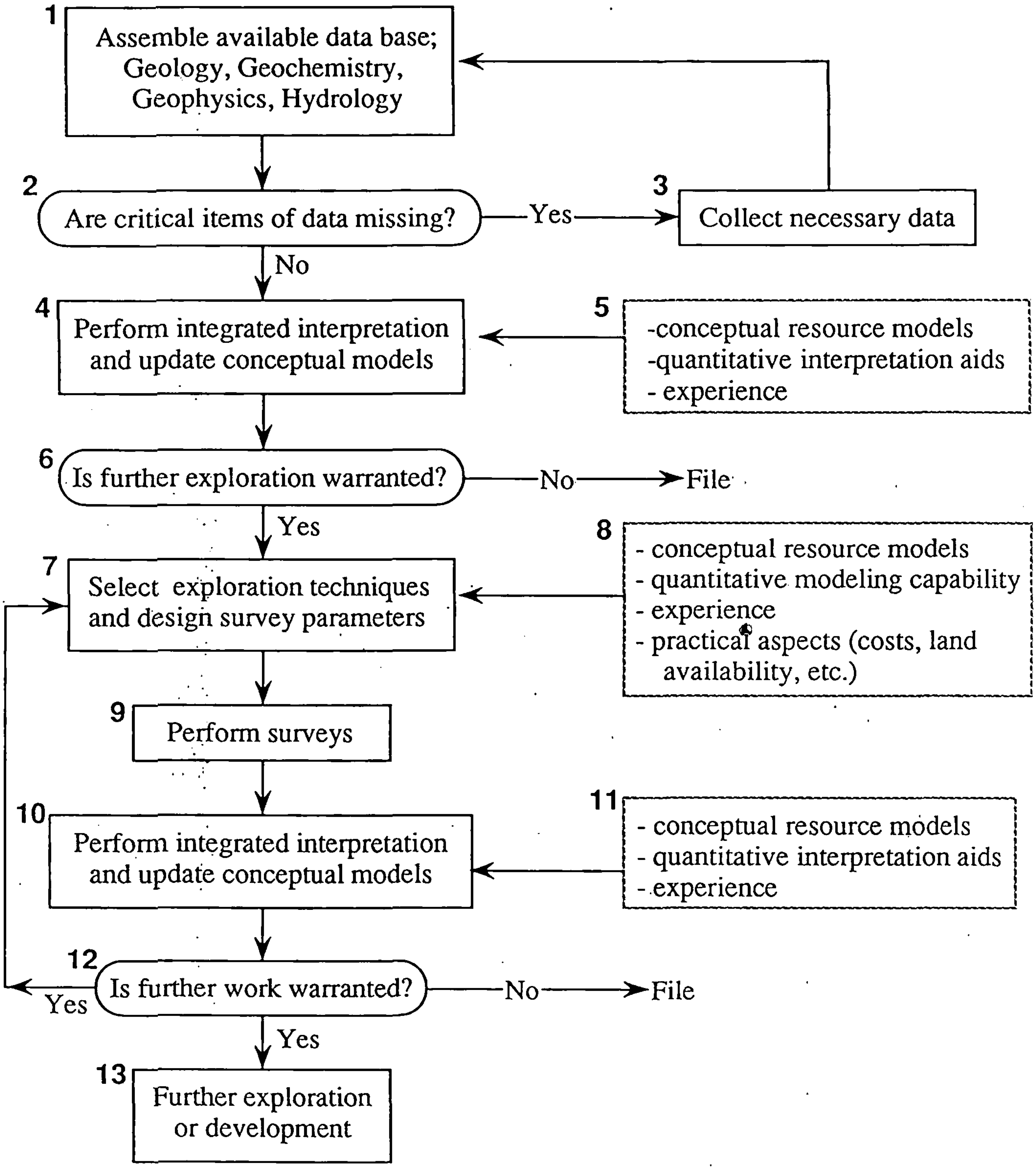


Figure 15 Basic exploration process.

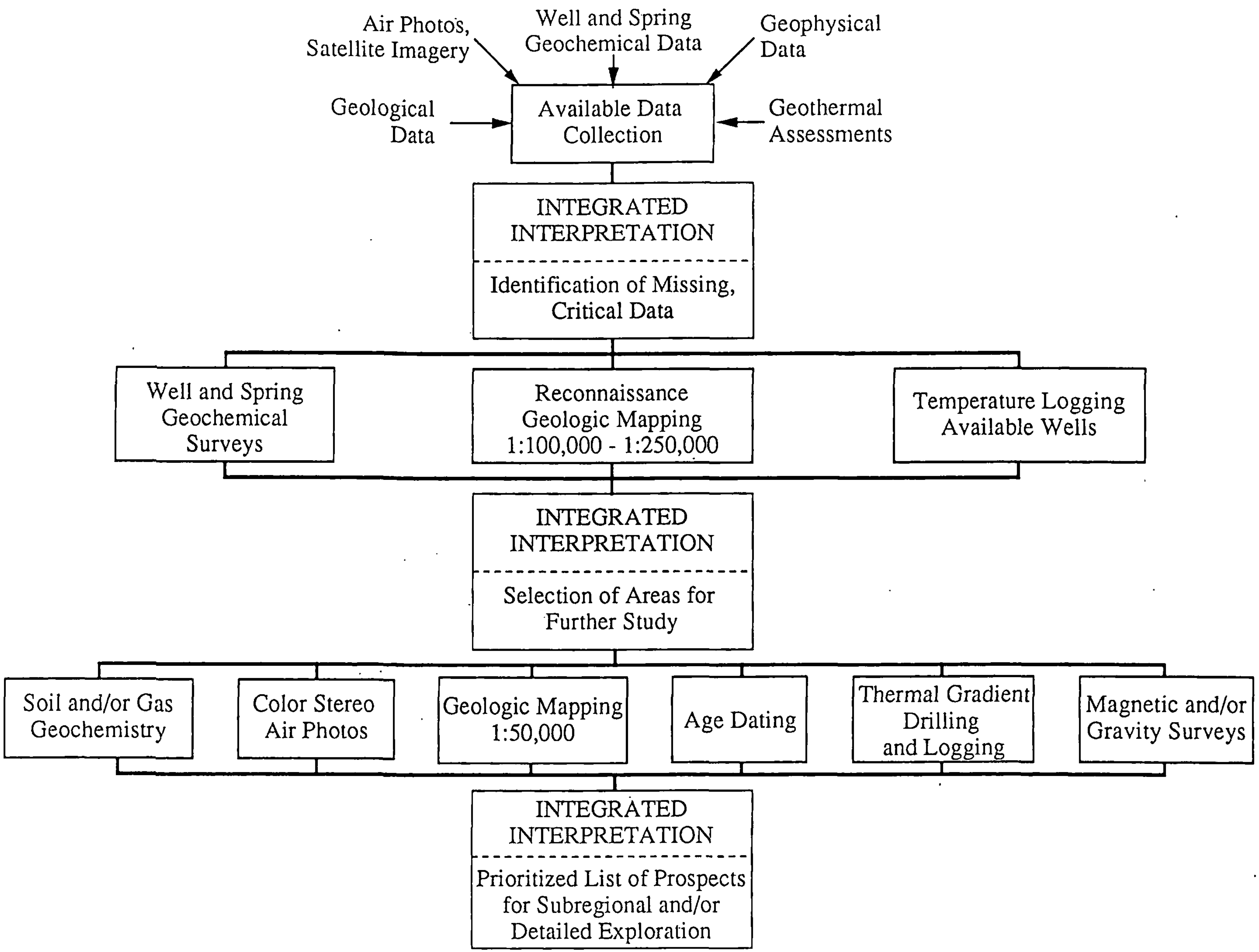


Figure 16 Regional exploration strategy.

Table I Characteristics of geothermal water types.

TYPE	ORIGIN	CHARACTER	OCCURRENCE
NEAR - NEUTRAL SODIUM CHLORIDE	Reaction of hydrothermal fluids with rocks.	-High dissolved solids, generally 1,000 to 30,000 ppm. -Enriched in Na, K, Ca, Cl, SO_4, HCO_3, Si. -Dissolved gases: CO_2, H_2S predominate.	Common deep high-temperature reservoir fluid. NaCl thermal springs.
ACID SULFATE	Steam rises from high-temp. water and condenses in near-surface oxidizing environment.	-Very dilute-enriched in SO_4 from steam. -Other elements derived from acid leaching of country rock.	Near surface, generally above water table.
ACID SULFATE-CHLORIDE	(1) Mixing of sodium-chloride and acid-sulfate waters, or (2) High temp. volcanic steam rich in F, Cl, and S, rises and condenses in near-surface, oxidized water	(1) Composition can vary widely (2) Enriched in Cl, SO_4, F	Near surface
NEAR-NEUTRAL SODIUM BICARBONATE-SULFATE	Steam rises and condenses in reduced ground water	-Low dissolved solids. -Enriched in HCO_3, variable amounts of SO_4.	On periphery or above the high-temperature reservoir.

Table II Plate-tectonic vs volcanic environments of geothermal areas.

	Environment	Quaternary Mafic Rocks	Quaternary Silicic Rocks	Dome Field	Caldera	Strato or Shield Volcano
Meager Creek, B.C.	Conv.	Y	Y	N	N	Y
Newberry, OR.	Conv.	Y	Y	N	Y	Y
Medicine Lake, CA.	Conv.	Y	Y	N	Y	Y
Lassen, CA.	Conv.	Y	Y	N	N	Y
The Geysers-Clear Lake, CA.	Trans (?)	Y	Y	Y	N	N
Salton Sea, CA.	Diverg-Trans	N	Y	Y	N	N
Heber, CA.	Diverg-Trans	N	N	N	N	N
Cerro Prieto, MX.	Diverg-Trans	N	Y	N	N	N
Valles, NM.	Diverg. (?)	N	Y	N	Y	N
Coso, CA.	B & R	Y	Y	Y	N	N
Long Valley, CA.	B & R	Y	Y	Y	Y	N
Steamboat, NV.	B & R	Y	Y	Y	N	N
Desert Peak, NV.	B & R	N	N	N	N	N
Dixie Valley , NV.	B & R	N	N	N	N	N
Beowawe, NV.	B & R	N	N	N	N	N
Roosevelt, UT.	B & R	Y	Y	Y	N	N
Cove Fort, UT.	B & R	Y	N	N	N	Y
Yellowstone, WY.	Plume	Y	Y	N	Y	N
Hawaii, HA.	Plume	Y	N	N	N	Y

Table III. Characteristics of high-temperature Basin and Range geothermal systems.

	Range-Front Fault	Vertical Offset >2km	Tv + Qal Fill >1km	Regional Heat Flow >85 mWm^{-2}	Tertiary Volcanics	Quaternary Volcanics	Quaternary Seismicity	Surface Thermal Features	Shallow Thermal Aquifer	NaCl Waters
Coso Hot Springs, CA.	×		×	×	×	Qr	×	×	—	×
Long Valley, CA.	—	—	×	×	×	Qr	×	×	×	×
Surprise Valley, CA.	×			×	×	—				×
Borax Lake, OR.	B			×	×	—		×	×	×
Cove Fort, UT.	×	×	×	×	×	Qb	×	×	×	×
Roosevelt Hot Springs, UT.	×	×	×	×	×	Qr	×	×	×	×
Thermo, UT.	×	×	×	×	—	—	×	×	×	×
Raft River, ID.	×		×	×	×	—	×	×	×	×
Beowawe, NV.	×	×	×	×	×	—	×	×	×	—
Brady, NV.	×		×	×	×	—	×	×	×	×
Desert Peak, NV.	×			×	×	—	×	×	×	×
Dixie Valley (Ctr.), NV.	×	×	×	×	×	—	×	×		×
Dixie Valley (No.), NV.	×	×	×	×	×	—	×	×		×
Fish Lake, NV.				×	×	—	×		×	×
Humboldt House, NV.	×			×	×	—	×	×	×	×
Salt Wells, NV.	×			×	×	—	×	×	×	×
San Emidio, NV.	×	×	×	×	×	—	×	×	×	×
Soda Lake, NV.	B			×	×	Qb	×	×	×	×
Steamboat Hot Springs, NV.				×	×	Qr	×	×	×	×
Stillwater, NV.				×	×	—	×	—	×	×

× -Characteristic applies on local or regional basis

B -Buried horst bordering fault - no exposed range

— -Feature not present

Qb -Quaternary basalts nearby

Qr -Quaternary rhyolite domes

-Blanks indicate uncertainty and/or insufficient data

Table IV. Regional exploration techniques.

Geological

Mapping	Rock types, structure, spring deposits, alteration
Imagery Interpretation	Structure, rock types, alteration
Structural Analysis	Regional & local structure, fault patterns
Age Dating	Identification of recent igneous rocks
Interpretation	Igneous activity, controls on heat sources, controls on permeability

Geochemical

Fluid Chemistry	Geothermometers, distribution of fluid types
Rock Chemistry	Trace-element zoning, permeable zones
Hydrothermal Alteration	Mineral zoning, permeable zones
Isotope Studies	Geothermometers, distribution of fluid types

Geophysical

Thermal Methods	High temperature, heat flow
Electrical Methods	Hot fluids, hydrothermal alteration, permeable zones
Gravity Methods	Structure, densification, cover thickness
Magnetic Methods	Structure, alteration, rock type
Seismic Methods	Active faulting, hydrothermally generated noise

GEOTHERMAL
SCIENCE AND
TECHNOLOGY

Dr. James C. Bresee
U.S. Department of Energy
Forrestal Building
Washington, D.C. 20585
U.S.A.

July 13, 1988

Dr. Phillip M. Wright
Technical Vice President
University of Utah Research Institute
391 Chipeta Way, Suite C
Salt Lake City, Utah 84108-1295

Dear Mike:

First, I want to thank you and your colleagues for your submission to GS&T. It is a excellent paper, and I am sure we will be publishing your contribution in Volume 2 early in 1989. I further anticipate that a book collecting all of Volume 1 and parts of Volume 2, including your article, will be published by Gordon and Breach next year.

Some modification to the article will be needed, and I am enclosing a marked-up copy of your submitted paper plus the corrected page proofs of Norm Goldstein's article, which will constitute all of Volume 1, Issue 4 to be distributed in the fall. Incidentally, timing dictated the sequence of publication; in the book version, I would expect your chapter to precede his.

It is not too surprising that there are some duplications between the two papers. It was in anticipation of such a possibility that I sent an earlier version of Norm's article to you. However, with Norm's essentially in final now, I think you will see the need for some condensation of your materials, particularly in pages 49 through 71. Earlier references to magma and hot dry rock resources might be modified slightly after you review Volume 1, Issue 2, which should reach you in a few days. Conversely, your brief discussion of exploration strategy at the end of your article could be considerably expanded, since neither your nor Norm's paper addresses that subject in detail.

Some figures need titles and legends. Some tables may not be needed in the final version. We will require glossy photos of the figures for the preparation of page proofs. I would greatly appreciate your best efforts to submit a modified version of your paper before the end of July, if at all possible. Call me to discuss any problems with the marked-up paper. And thanks again for your contribution.

Sincerely,

Jim

James C. Bresee
Editor-in-chief

GEOTHERMAL SCIENCE AND TECHNOLOGY

Dr. James C. Bresee
U.S. Department of Energy
Forrestal Building
Washington, D.C. 20585
U.S.A.

December 23, 1987

Dr. Phillip M. Wright
Associate Director, Earth Science Laboratory
University of Utah Research Institute
420 Chipeta Way, Suite 120
Salt Lake City, Utah 84108

Dear Mike:

As I mentioned when we met several months ago, the geothermal book which Paul Witherspoon and I have been assembling for several years and to which you have contributed has taken an interesting turn. The present plan is for the chapters to be used to launch the new journal "Geothermal Science and Technology". The twelve chapters would be printed during the first year of the journal, at the end of which time they would be assembled and printed as a separate book with a small amount of additional material which Paul and I would supply. We are well on the way to the implementation of that plan, with the first issue of the journal out (enclosed) and the second issue in page proof form.

I would like very much to include your chapter in the third issue, and I would appreciate it if you would consider updating your earlier contribution (also enclosed). The editorial conventions we are using in the new journal are listed on its inside back page. I have done a little editing of the first few pages. There seems to be a missing page at about page 7. Assuming you have the original, perhaps you could solve the mystery. Ultimately, I will need glossy copies of the seven figures. Does the updating look like a task you could complete in the next month or two?

I hope you and your family are having a splendid holiday period. I look forward to hearing from you early next year.

Sincerely,

James C. Bresee
Editor-in-chief

MEMO TO: STAN, DENNIS, JOE, AND HOWARD

FROM: MIKE WRIGHT

SUBJECT: ARTICLE FOR "GEOTHERMAL SCIENCE AND TECHNOLOGY"

DATE: March 3, 1988

You can probably remember back to the time when a high priority was put on producing an article for Jim Bresee entitled "Exploration Strategies for Regional Assessment of Hydrothermal Resources". After we had written such an article and gotten it into pretty fair shape, the project was dropped in Washington. Well, it is back. Jim has been in touch with me and he would like us to update this article for publication in the new journal "Geothermal Science and Technology". One issue of this journal has come out, and a second is in the process. The second one contains an article by LBL on detailed exploration within geothermal systems. Ours is to cover only the regional exploration aspects.

I believe it is to our advantage to update our article and get it to Jim. He needs it by 1 April in order to be in time for Volume 1, Number 4 to be issued in September. Although Jim is no longer in the geothermal division of DOE, he is still a supporter of ours. In addition, it will discharge an old commitment.

Jim tells me that he plans to have our article along with several others published as a text book after their publication in the journal.

Could each of you update one of the sections as follows:

Stan (with my help)	Geophysical Techniques
Dennis	Geological Techniques
Joe	Geochemical Techniques
Howard	Hydrothermal Systems in the U. S.

Would you each please read the attached copy to see that the balance and content of each section is appropriate. If you have suggestions on one of the sections you are not revising, please give them to the person doing the revision. I will put everything together and get it through typing when I have your input. Please take care of your own illustrations with Pat. Also, please check your references and give me any additions. When we get everything put together, say around the third week of March, we can all give it a thorough review before it goes to Jim.

GEOTHERMAL SCIENCE AND TECHNOLOGY

Dr. James C. Bresee
U.S. Department of Energy
Forrestal Building
Washington, D.C. 20585
U.S.A.

March 14, 1988

Dr. Phillip M. Wright
Technical Vice President
Earth Science Laboratory
University of Utah Research Institute
391 Chipeta Way, Suite C
Salt Lake City, Utah 84108

Dear Mike:

It was good to hear by telephone that you and your staff are busily updating your paper for GS&T. Since it will not be available for several more weeks, I am planning in the third issue to put in Norm Goldstein's paper on subregional exploration techniques. Logically, your paper on regional exploration should have preceded Norm's, and it certainly will in the book version.

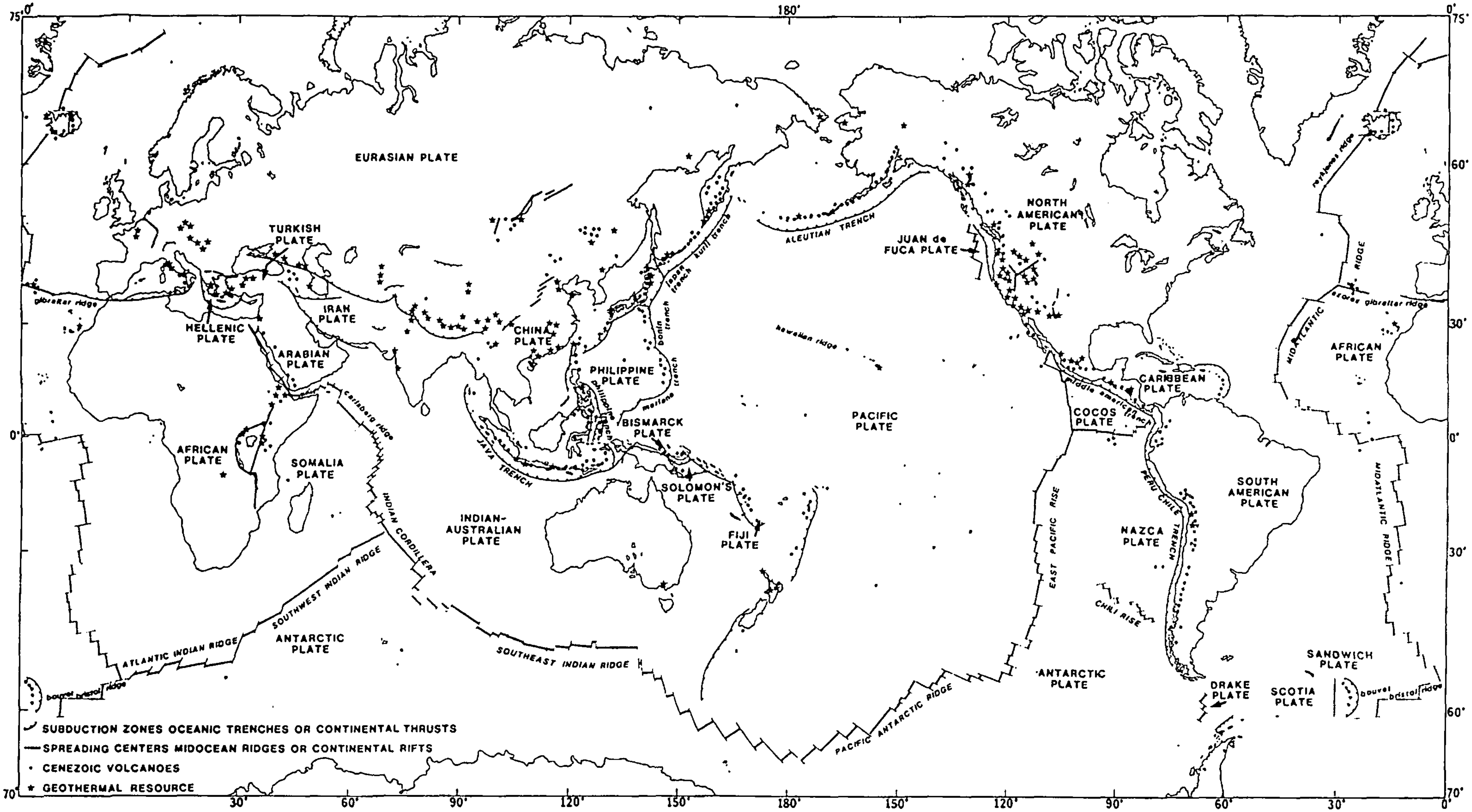
May I suggest that you glance over Norm's paper ("Subregional and Detailed Exploration for Geothermal-Hydrothermal Resources", LBL 18224, May 1986, updated February 1988) to see how well your two papers will fit together. Your library copy would be the original May 1986 version; I am enclosing the table of contents showing certain deletions I made, in one case to omit the remote sensing material, since I expected that you would be covering that. If you need to see any part of the February 1988 version, I would be happy to Xerox it and send it to you.

I look forward to seeing your paper soon and including it in the fourth issue of GS&T.

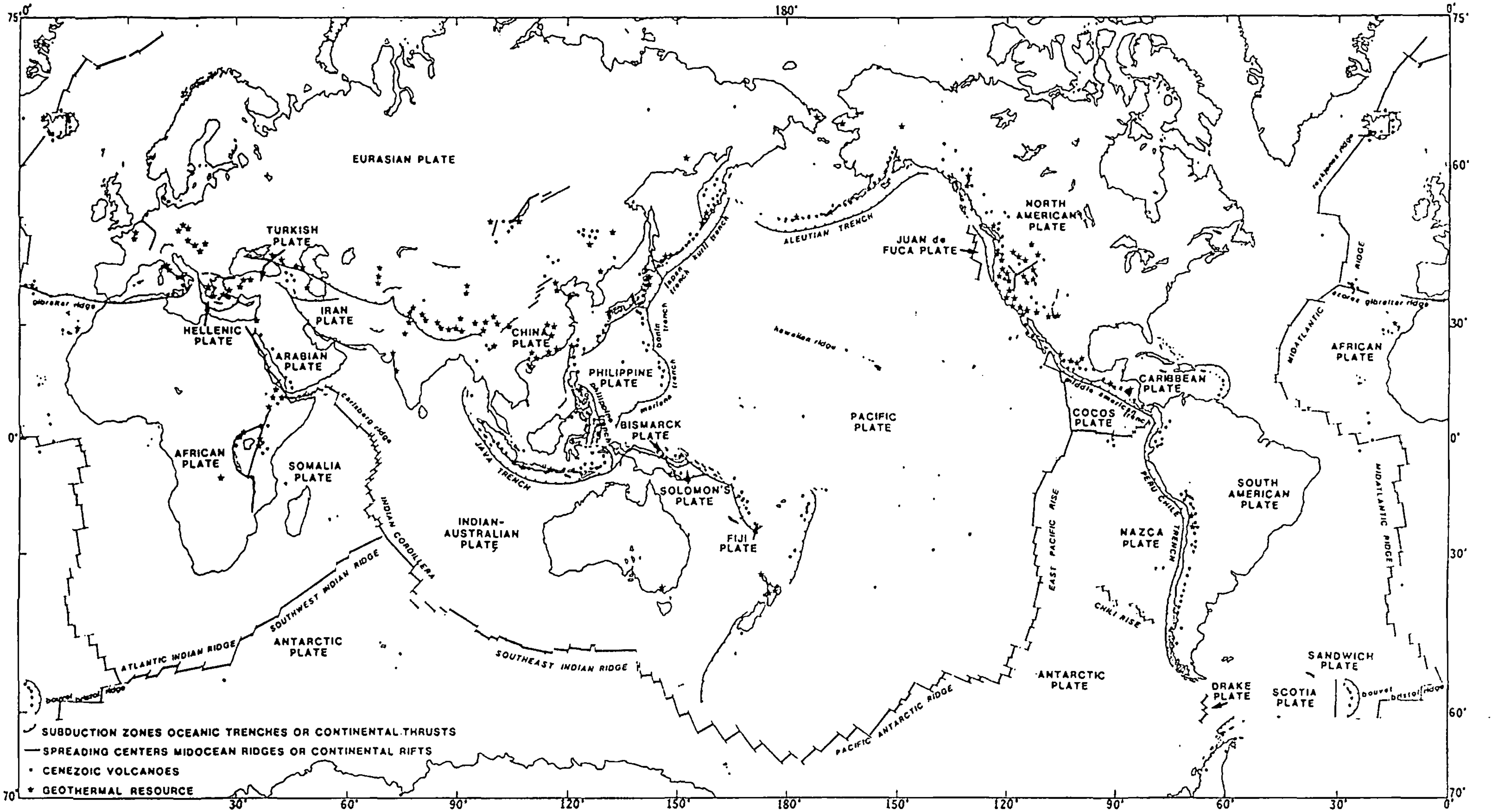
Sincerely,

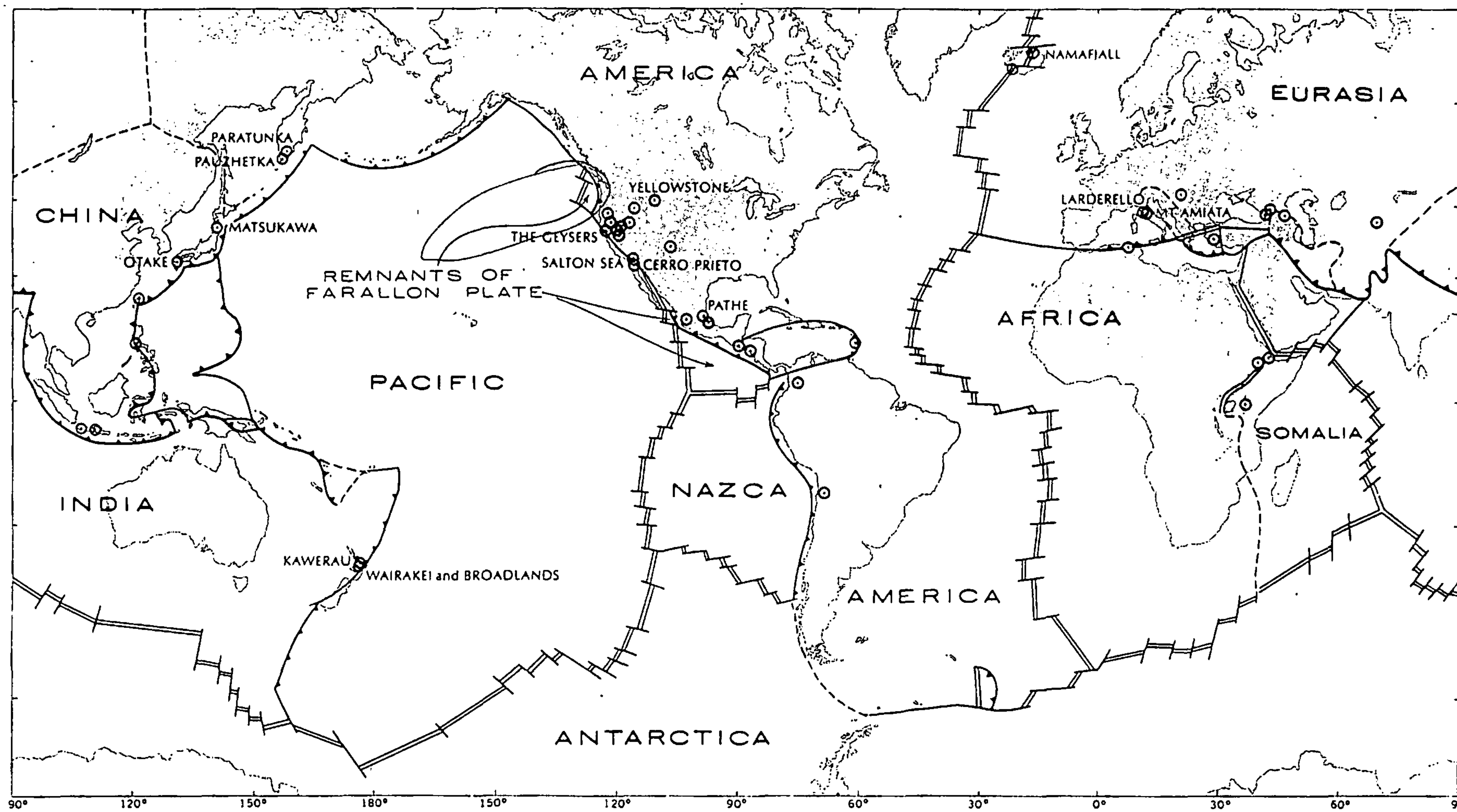
James C. Bresee
Editor-in-chief

cc: John Ford, Gordon and Breach



EURASIAN PLATE
TURKISH PLATE
HELLENIC PLATE
IRAN PLATE
ARABIAN PLATE
CHINA PLATE
PHILIPPINE PLATE
BISMARCK PLATE
SOLOMON'S PLATE
FIJI PLATE
AFRICAN PLATE
SOMALIA PLATE
INDIAN-AUSTRALIAN PLATE
ANTARCTIC PLATE
PACIFIC PLATE
NORTH AMERICAN PLATE
JUAN de FUCA PLATE
CARIBBEAN PLATE
COCOS PLATE
NAZCA PLATE
SOUTH AMERICAN PLATE
DRAKE PLATE
SCOTIA PLATE
SANDWICH PLATE
ALEUTIAN TRENCH
JAVA TRENCH
PERU CHILE TRENCH
EAST PACIFIC RISE
CHILI RISE
MIDATLANTIC RIDGE
SOUTHWEST INDIAN RIDGE
ATLANTIC INDIAN RIDGE
SOUTHEAST INDIAN RIDGE
PACIFIC ANTARCTIC RIDGE
INDIAN CORDILLERA
hawaiian ridge
carlsberg ridge
SUBDUCTION ZONES OCEANIC TRENCHES OR CONTINENTAL THRUSTS
SPREADING CENTERS MIDOCEAN RIDGES OR CONTINENTAL RIFTS
CENEZOIC VOLCANOES
GEOTHERMAL RESOURCE


EURASIAN PLATE
TURKISH PLATE
HELLENIC PLATE
IRAN PLATE
ARABIAN PLATE
CHINA PLATE
PHILIPPINE PLATE
BISMARCK PLATE
SOLOMON'S PLATE
FIJI PLATE
AFRICAN PLATE
SOMALIA PLATE
INDIAN-AUSTRALIAN PLATE
ANTARCTIC PLATE
PACIFIC PLATE
JUAN de FUCA PLATE
NORTH AMERICAN PLATE
CARIBBEAN PLATE
COCOS PLATE
NAZCA PLATE
SOUTH AMERICAN PLATE
DRAKE PLATE
SCOTIA PLATE
SANDWICH PLATE
ALEUTIAN TRENCH
JAVA TRENCH
PERU CHILE TRENCH
EAST PACIFIC RISE
CHILI RISE
MIDATLANTIC RIDGE
INDIAN CORDILLERA
SOUTHWEST INDIAN RIDGE
ATLANTIC INDIAN RIDGE
SOUTHEAST INDIAN RIDGE
PACIFIC ANTARCTIC RIDGE
gibraltar ridge
azores gibraltar ridge
carlsberg ridge
hawaiian ridge
bouvet bristol ridge
middle america trench
japan trench
kuril trench
bonin trench
mariana trench
philippine trench
reykjanes ridge
SUBDUCTION ZONES OCEANIC TRENCHES OR CONTINENTAL THRUSTS
SPREADING CENTERS MIDOCEAN RIDGES OR CONTINENTAL RIFTS
CENEZOIC VOLCANOES
GEOTHERMAL RESOURCE


AMERICA
EURASIA
NAMAFJALL
PARATUNKA
PAUZHETKA
CHINA
MATSUKAWA
OTAKE
YELLOWSTONE
THE GEYSERS
SALTON SEA
CERRO PRIETO
REMNANTS OF FARALLON PLATE
PATHE
LARDERELLO
MT. AMIATA
AFRICA
PACIFIC
SOMALIA
INDIA
NAZCA
KAWERAU
WAIRAKEI and BROADLANDS
AMERICA
ANTARCTICA
90°
120°
150°
180°
150°
120°
90°
60°
30°
0°
30°
60°
90°
60°
45°
30°
0°
30°
60°

	Environ.	Quat. Basic Rocks	Quat. Silicic Rocks	Dome Field	Caldera	Reservoir Rock	Quat. Faulting	Max Temperature °C	Fluid Salinity ppm	Thermal Springs	Fumaroles	Surface Shallow Outflow Plume	Cold water Blanket	Stratovolcano		
Meager Creek, B.C.	Conv	Y	Y	N	N	x/n	?	260	< 5000	A	A	!	N	Y		
Newberry, OR	Conv	Y	Y	N	Y	volc	Y	265	?	A	N	N	Y	Y		check T
Medicine Lake, CA	Conv															
Lassen, CA	Conv	Y	Y	N	N	volc	Y	> 220	?	N	Y	?	N	Y		
The Geysers/Clear Lake, CA	Trans ?	Y	~~N~~		N	mdgm.	Y	> 300	V	Y	Y	N	Y	N		check rhyolites, domes
Salton Sea, CA	Rift-Trans	N	Y	Y	N	Seds.	Y	360	>25%	N	Y	?	?	N		
~~Heber~~, CA	Rift Trans	N	N	N	N	seds	Y	200	15,000	N	N	N	N	N		
Cerro Prieto, MX	Rift Trans	[illegible]	Y	N	N	Seds	Y	360	1500-20,000	?	Y	ch	N	N		check basalt salinity springs
Coso, CA	Exten	Y	Y	Y	N	x/n	Y	340	10,000	N	Y	N	N	N		
Long Valley, CA	Exten	Y	Y	~~Y~~	Y	vol/x/n	Y	~~[illegible]~~	ch	Y	Y	?	N	N		Check Shady Rest
Steamboat, NV	Exten	Y	Y	Y	N	x/n	Y	ch	< 5000	Y	Y	Y	N	N		check
Desert Peak, NV	Exten	N	N	N	N	x/n	Y	?	?	N	N	Y	N	N		
Dixie Valley, NV	Exten	N	N	N	N	sed	Y	ch	< 5000	N	Y	Y	N	N		ch
Beowawe, NV	Exten	N	N	N	N	vol	Y	230	< 2000	Y	Y	Y	N	N		
Roosevelt, UT	Ext	~~Y~~	Y	Y	N	xtn	Y	270	10,000	Y	Y	Y	N	N		
Cove Fort, UT	Ext	Y	N	N	N	vol/sed	Y	200 ~~175~~	V	N	Y	N	N	Y		
Yellowstone, WY	Mantle Plume	Y	Y	N	Y	?Vol	Y	270(e)	< 10,000	Y	Y	Unk	N	N		
Valles...																

GEO-HEAT CENTER

Oregon Institute of Technology • Klamath Falls, Oregon 97601 • 503/882-6321 Ex. 267

Paul J. Lienau, Director

June 10, 1988

Mike Wright
Univ. of Utah Research Institute
391 Chipeta Way, Suite C
Salt Lake City, UT 84108

Dear Mike:

Enclosed is your chapter on Nature of Geothermal Resources (correct title?). Gene has done a major re-write by including information on low temperature resources. This was primarily taken from USGS Circular 893.

We need to wrap this work up. This means we need your review of the two chapters, figures, tables, and references no later than June 30, 1988 and earlier if possible.

I have enclosed a recent Site Data Base for Direct Use in the United States. We would appreciate your comments on this report.

I look forward to hearing from you soon. Should you have questions, please give me a call.

Sincerely yours,

Paul

Paul J. Lienau
Director

Enclosures

cc: B. Lunis

GEO-HEAT CENTER
Oregon Institute of Technology • Klamath Falls, Oregon 97601 • 503/882-6321 Ex. 267

Paul J. Lienau, Director

May 15, 1987

Mike Wright
UURI
391 Chipeta Way, Suite c
Salt Lake City, UT 84108-1295

Dear Mike:

Please find enclosed a sample of a chapter for the Handbook which contains suggested guidelines to establish continuity among authors.

Cindy developed this to give you an idea of appearance for the published handbook, which is similar to the GRC Transactions. In addition, the guideline contains conformity questions which she usually encounters in developing reports and papers by GHC staff. If you or your secretary have questions, please give us a call.

We received your floppy containing a document file using Wordperfect and a textfile compatible with ASC II and DOS. As you can see from the two printed samples of Mikes' outline, the text file printed with returns and for the document file the returns were not recognized. Wordperfect software could have been used to print the document file. Text files are preferred since different authors will be using various software. Our IBM AT MS-DOS operating system will not recognize certain commands, such as returns, underlining and bold for document files. Therefore, please send a hard copy of your completed chapter along with the floppy.

The GHC does have the capability to receive files via the Hayes Modem, which is compatible with Crosstalk and most other modems. Please give me a call if you wish to transmit by this means so we can coordinate baud rate, etc.

Thank you for your cooperation and please let me know if you have questions.

Sincerely,

Paul

PAUL J. LIENAU
Director

/cn

enc

1985 INTERNATIONAL SYMPOSIUM ON GEOTHERMAL ENERGY

INTERNATIONAL VOLUME

GEOTHERMAL RESOURCES COUNCIL

THE CURRENT STATUS OF GEOTHERMAL DIRECT USE DEVELOPMENT IN THE UNITED STATES

Deepak C. Kenkeremath, Robert E. Blackett, James V. Satrape and Gene V. Beeland

Meridian Corporation
5113 Leesburg Pike, Suite 700
Falls Church, Virginia 22041

ABSTRACT

Information obtained on commercial, agricultural, industrial, institutional, and residential projects in the United States that utilize low- to moderate-temperature geothermal water as an energy source is presented in this paper. The annual thermal energy use of a total of 263 projects either on-line, under construction, or under expansion has been estimated to be over 1861 x 10^9 Btu. Of the total annual utilization, space and water conditioning projects (473 x 10^9 Btu) account for approximately 25 percent; district heating projects (426 x 10^9 Btu) are estimated to account for 23 percent; commercial fish farms (396 x 10^9 Btu) comprise 21 percent; commercial greenhouses (328 x 10^9 Btu) contributed 18 percent; while projects involving small resorts (120 x 10^9 Btu) and industrial process heat (118 x 10^9 Btu) combine to make up the remainder. All but one of the identified active projects are located in states west of the Mississippi River, with the bulk of the geothermal energy direct heat utilization occurring in California, Idaho, Oregon, Nevada, and New Mexico. The possible influence of various federal and state incentive programs on geothermal direct use development is discussed.

INTRODUCTION

The survey of current use of geothermal energy in direct heat applications on which this paper is based indicates considerable growth in direct use since the late 1970s. Until incentives were set in motion during that period, the growth observed in Figure 1 is accounted for largely by one-well-one-structure use in Klamath Falls, Oregon, and Reno, Nevada; small greenhouses; and resorts. No municipal geothermal district heating systems had been initiated since the Boise and Ketchum, Idaho, systems were installed before the turn of the century.

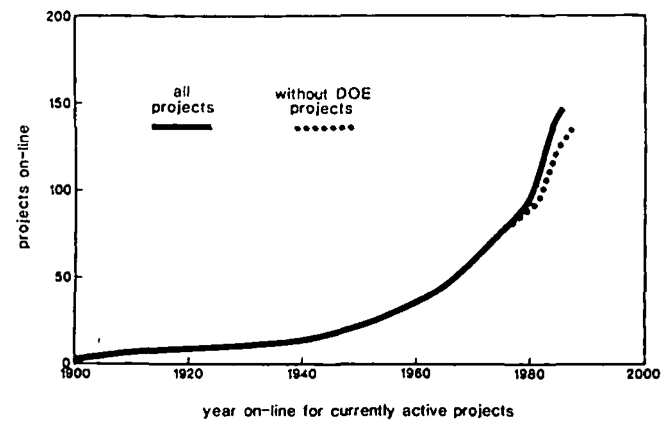


Figure 1. Geothermal direct use activity in the United States

An unforeseen incentive resulted from the oil embargo crisis of 1973 and the rapidly rising fuel prices that followed — geothermal energy suddenly became a much more attractive and competitive alternative source of energy. As a result, large budgets were provided to the U.S. Department of Energy (DOE) to foster early commercialization of the resource, both for power generation and direct use.

In 1978, in order to provide an impetus for early growth in direct use applications, DOE instituted several assistance programs. The Geothermal Program Research and Development Announcement (PRDA) Program consisted of a number of studies of the engineering and economic feasibility of specific direct use applications. Under the Program Opportunity Notice (PON) Program, DOE cost-shared 23 demonstration direct use projects with private companies, municipalities, and other organizations. As a result of the State-Coupled Low-Temperature

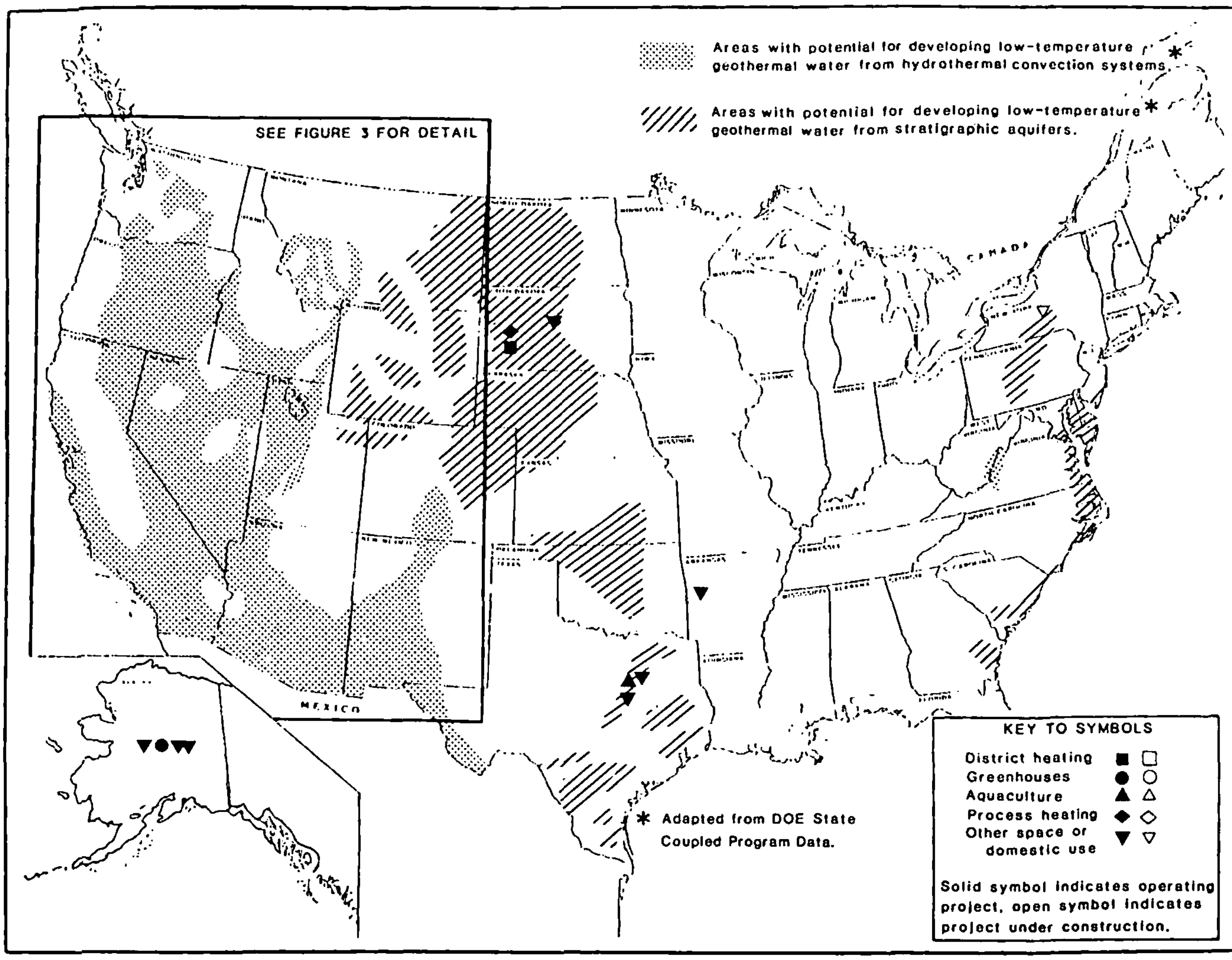


Figure 2. Locations of geothermal direct use projects in the United States

Geothermal Resource Assessment Program, low-temperature and moderate-temperature reservoirs of 16 states were confirmed and mapped. In a related action, DOE began direct support of several research and academic institutions to provide technical assistance on direct use technology and economics, and to disseminate resource information. DOE has also issued loan guarantees to three direct use projects under the Geothermal Loan Guaranty Program.

Also in 1978, the Energy Tax Act provided a tax credit for residential and commercial use of geothermal energy, which was increased in 1980 by the Crude Oil Windfall Profits Tax Act. The credit, of up to $4,000 for residences and 15 percent for business investment over the standard 10 percent, is applicable to the relatively high front-end cost of geothermal direct uses.

More recently, several states and the Bonneville Power Administration have offered incentives for geothermal direct use projects. These incentives take various forms such as income tax credits — residential and business — for installing a geothermal system, property tax assessment exemptions, loans, and grants.

Although it is difficult to assess the degree of influence of any individual incentive — beyond the obvious observation that some of the DOE-assisted projects may not have been realized without financial and technical support — the incentives have apparently worked collectively as intended. The use of geothermal energy in direct heat applications has risen dramatically since they were offered, despite concurrent disincentives such as high interest rates and reductions in the costs of competing fuels. A distinct increase in the rate at which direct use projects have come on-line is shown in Figure 1 to have begun around 1979-80, even excluding DOE's PON projects.

CURRENT STATUS OF DIRECT USE

Geothermal energy is estimated to supply approximately 1625 x 10^9 Btu of heat energy annually through the direct heat applications in current use in the United States. Projects under construction or expansion will utilize another 236 billion Btu annually in the very near future. Large scale projects in California, Idaho, Oregon, and Nevada (Figures 2 through 5) are the major users,

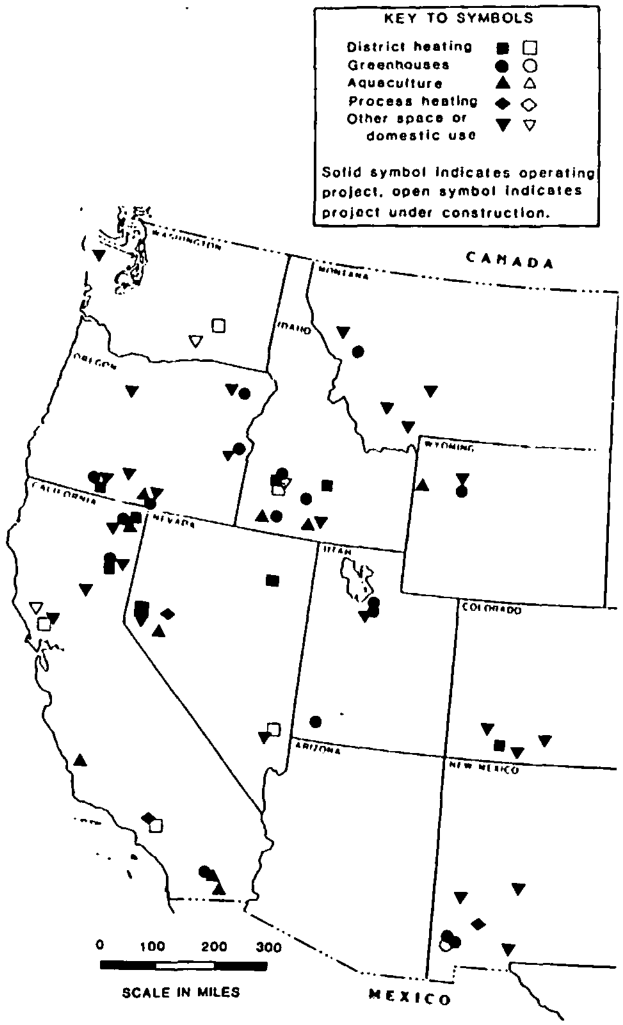


Figure 3. Locations of geothermal direct use projects in the western United States

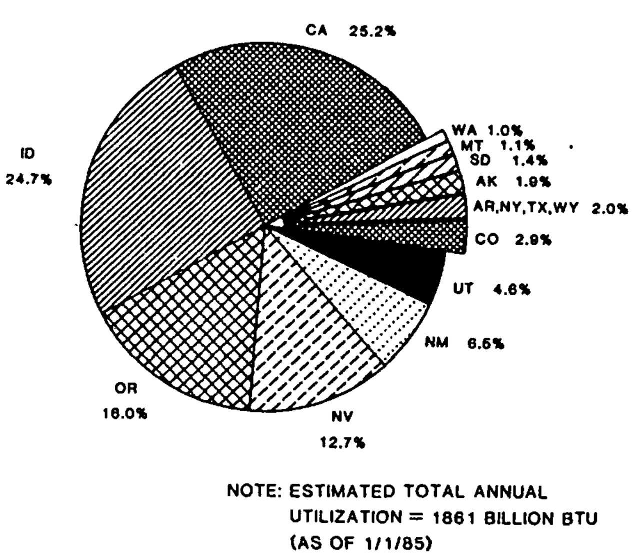


Figure 4. Estimated geothermal direct utilization by state (projects on-line and under construction)

representing nearly 80 percent of the total annual usage. A breakdown of this total by state and application is given in Table 1. The following sections discuss the various uses of geothermal energy in direct heat applications.

District Heating and Space Conditioning

District heating and space/water conditioning projects are the two major direct heat beneficiaries of geothermal energy in the United States. These two categories of use are very similar in many cases, to the extent that district heating could be considered a special subset of space conditioning.*

There are 11 geothermal district heating projects currently on-line, which account for an estimated 228 x 10^9 Btu annually. A large percentage of this energy is utilized in district heating systems at Klamath Falls, Oregon, and Boise, Idaho. An additional 198 x 10^9 Btu annually may be realized from six district heating projects presently under construction. A system under construction at San Bernardino, California, may eventually provide the greatest amount of heat energy of all projects of this type in the U.S.

Space and water conditioning projects (excluding district heating) either on-line or under construction appear to account for the greatest energy usage of the various types of direct heat geothermal projects. Included in this category are numerous individual residences (for example, 550 homes in Klamath Falls, Oregon, and nearly 200 homes in Reno, Nevada), plus institutional heating systems for hospitals, schools, churches, and other public supported facilities and commercial buildings. Other projects considered in this category include thermal water used for domestic hot water supply and water preheating. The estimated total annual end use energy supplied by geothermal for space conditioning and domestic hot water projects is 436 x 10^9 Btu. An additional 37 x 10^9 Btu annually may be realized from projects that are presently under construction.

Aquaculture

Geothermal energy is utilized in aquaculture projects where warm water fish and shellfish are raised for sale as a

*Note: For purposes of this report, district heating systems are narrowly defined as *public utility systems serving multiple end users from a central plant.*

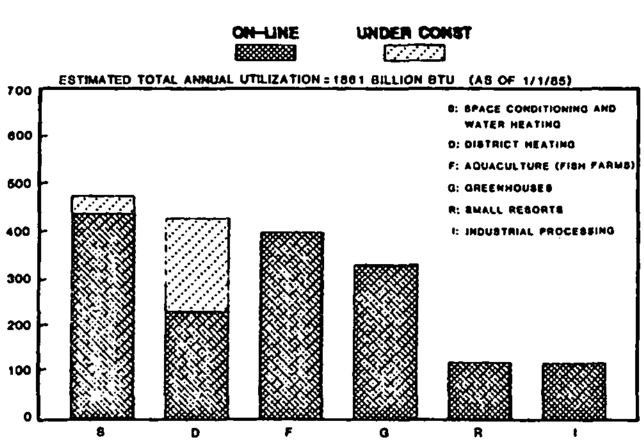


Figure 5. Estimated annual geothermal energy use, by application, for direct heat projects in the United States

Table 1. Estimated Geothermal Direct Heat Utilization in the United States

STATE	ANNUAL UTILIZATION (BTU x 10^9) — APPLICATION: Space and Water Conditioning	District Heating	Greenhouses	Aquaculture	Small Resorts and Baths	Process Heat	TOTAL: On-Line	TOTAL: Under Constr	NUMBER OF PROJECTS: On-Line	NUMBER OF PROJECTS: Under Constr
AK	10	–	–	–	25	–	35	–	9	–
AR	2	–	–	–	–	–	2	–	1	–
CA	50 (5)	11 (134)	130	121	10	8	330	139	19	4
CO	8 (3)	23	–	–	20	–	51	3	17	1
ID	30 (20)	93	72	230	15	–	440	20	76	1
MT	7	–	<1	–	14	–	21	–	15	–
NV	11	39 (59)	–	37	6	85	178	59	17	1
NM	53	1	31 (1)	–	10	25	120	1	20	–
NY	(8)	–	–	–	–	–	–	8	–	1
OR	210	51	24	–	13	–	297	–	52	–
SD	16	10	–	–	–	–	26	–	3	–
TX	16	–	–	–	–	–	16	–	3	–
UT	11	–	68	–	6	–	85	–	14	–
WA	12 (1)	(5)	–	–	–	–	12	6	1	2
WY	1	–	2	8	1	–	12	–	6	–
TOTAL ANNUAL UTILIZATION (BTU x 10^9): On-Line	436	228	327	396	120	118	1625		253	
Under Construction	37	198	1	–	–	–		236		10
Total	473	426	328	396	120	118	1861		263	

Note: Values may not sum exactly due to independent rounding.
Parentheses denote projects under construction.

food source or as ornamentals. Low-temperature geothermal systems are attractive for commerical aquaculture projects since they permit raising certain types of species in climates where fuel costs for such projects would otherwise be prohibitive. Of the estimated total (396 x 10^9 Btu) annual geothermal energy use in nine aquaculture projects, three facilities in Buhl, Idaho; Wabuska, Nevada; and Mecca, California account for nearly 80 percent of the total energy consumption. Geothermal aquaculture facilities require a relatively high rate of flow at a uniform temperature, and geothermal wells are often used in combination with normal water wells. Temperature drop through the facility is usually small; therefore, only a fraction of the total available energy is used.

Agribusiness

Geothermally heated greenhouses are commonplace as many small individual users often cascade thermal water from a primary use, such as space heating, to a greenhouse. Several large commercial greenhouses have also come on-line over the last several years because the geothermal option has become attractive as a fast payback cost to co-located users. There are 28 projects presently in operation with one project undergoing expansion.

Large greenhouse complexes have been constructed at a number of locations. Those located near Susanville and Cochella, California; Animas, New Mexico; Salt Lake City, Utah; Twin Falls, Idaho; and Vale, Oregon account for nearly 80 percent of the estimated total 327 x 10^9 Btu utilized annually. Products produced include decorative flowers, potted plants, and hydroponically grown vegetables.

Industrial Process Heat

Geothermal heat for industrial processing, although potentially very attractive, is used at only a handful of locations in the U.S. Unlike other geothermal direct applications, it is often difficult for a user to relocate a commerical enterprise near a geothermal site since cost effective industrial processing may be governed by factors other than potential fuel savings. The total estimated value for direct heat geothermal utilization (118 x 10^9 Btu) annually through industrial processing was derived from a sewage digestion plant in San Bernardino, California, a vegetable drying facility at Brady Hot Springs, Nevada, and copper processing at Hurly, New Mexico.

Small Resorts

Use of geothermal fluids at resorts, spas, and balneological baths has been the most widespread historical application of geothermal heat. Smaller isolated resorts, numbering approximately 120, have been estimated in this category although large resort complexes were considered under the category of space and water conditioning. The smaller resorts occur at numerous locations throughout the far western states. The total annual energy usage (120 x 10^9 Btu) represented for this category is estimated based upon very limited technical data.

METHODOLOGY

Reference Temperature

Geothermal resource assessments performed by the U.S. Geological Survey in the past have defined thermal water above a convenient reference temperature. For the purpose of this study the reference temperature was arbitrarily assigned at 15 degress Celsius *above* the mean annual ambient air temperature for the region in question. This assignment was made in order to eliminate the majority of groundwater heat pump applications for which resource and end use technical data would be difficult, if not impossible, to acquire.

Data Acquisition

In order to develop a current data base on geothermal direct use projects in this country, it was necessary to use multiple sources for information on project location, type, size, reference temperature, flow rate, and number of wells. A direct use data file compiled by the Mitre Corporation (1981), under contract to the U.S. Department of Energy, was used to construct an initial list of potential projects to be confirmed by new information or modified as necessary.

Where possible, data were extracted from: 1) the Transactions of the Geothermal Resources Council (GRC) annual meetings; 2) the Regional Geothermal Progress Monitor, circulated monthly by the Idaho National Engineering Laboratory; 3) monthly bulletins of the GRC; 4) the *Geothermal Hot Line*, published semi-annually by the California Division of Oil and Gas; and 5) the Quarterly Bulletin of the Oregon Institute of Technology, Geo-Heat Center. In addition, data were drawn from a number of reports prepared in support of programs sponsored by DOE in cooperation with state agencies and recent news articles (Meridian Corp., 1985).

Contacts were established with those state offices responsible for tracking and (or) regulating geothermal projects. Listings of current on-line projects within an individual state were obtained, and, in certain cases, an effort was made to contact representatives at the project sites.

The data gathered were then compiled into a direct heat data base for use in estimating the amount of energy being utilized from low-and moderate-temperature geothermal systems through direct applications.

Estimation Methods

The energy use reported in Table 1 for the individual projects (Appendix A) is an estimation of the annual beneficial heat actually extracted from the geothermal fluids. Methods used to arrive at this estimate were highly diverse due to the varied nature of the data available from project to project. Temperature and flow information from all sources was used where necessary to arrive at an estimation of annual usage.

The estimation methods used can be classified into primary methods (based on reported data and (or) reasonably certain conditions) and secondary methods (estimates based on very limited information). In all cases, "beneficial heat extracted" or end-use energy consumption was the quantity estimated.

Primary methods of estimation include: (1) energy use reported in the literature for specific projects; (2) estimates quoted by state and project contacts; and (3) calculations where flow rates and temperature drop were known. For individual houses, energy estimates are based on heating degree days at the location as determined by the Energy Information Administration (1983), and natural gas was assumed as the fuel displaced. Where quantities of fuel displaced were reported, these were converted to end-use energy using typical residential and commerical boiler efficiency factors:

1) For fuel oil, propane, or natural gas: 70 percent
2) For electricity: 100 percent.

Table 2. Breakdown of Projects by Estimation Methods

	Primary Estimate		Secondary Estimate	
	Projects	Billion Btu/Year	Projects	Billion Btu/Year
On-Line	77	1008	176	617
Under Construction	5	69	5	167
TOTAL	82	1077	181	784

Secondary methods of estimation (indicated by "**" in Appendix A) were used when no information about energy use was available or could be derived. Data available from primary estimates were tabulated and assembled by project type, size, and climate. These "known" points set guidelines whereby an estimate of energy use for other projects could be assigned.

A breakdown of the project summaries by estimation methods is shown in Table 2.

Primary estimation methods account for greater than 50 percent of the total energy use estimated. It was possible to use primary methods on most of the larger projects because data were available in the literature. Secondary methods were primarily needed for resorts and other small projects where detailed engineering studies may never have been performed. Although secondary estimates are at best approximations, they are reasonable relative to the known energy use of other projects. In summary, the total energy use values reported here are based on the best available information and are believed to be a realistic approximation of existing geothermal direct use energy consumption.

COMPARISON OF THE METHODS AND RESULTS OF THIS STUDY WITH PREVIOUS ASSESSMENTS

Since 1981, the Mitre assessment discussed above has provided the commonly used statistics on geothermal direct use applications. The figures derived — 213 operating systems utilizing 13,104 x 10^9 Btu per year — have been widely publicized. The substantial variation in the heat value total calculated by this assessment (1625 x 10^9 Btu) is due to differences in study methods.

First, energy use was totalled only for the applications now customarily considered geothermal direct heat uses. In the totals reported by Mitre, a single water flood enchanced oil recover (EOR) project of 10,000 x 10^9 Btu was counted. Due to the magnitude and level of approximation of this figure, it completely overshadows energy consumption from "traditional" direct use projects. Thus, the Meridian assessment did not consider EOR using oil field brine as "direct use" of the geothermal resource.

There is another difference in study methodology that should be noted since it also produces different results from the Mitre study. That assessment calculated a "replaced energy" value as opposed to estimates of energy extracted to represent total energy use. That approach is perfectly valid and was especially popular during the energy crisis of the 1970s.

Developing both values for a large aquaculture facility located near Buhl, Idaho, will illustrate the difference. The facility has a number of wells to provide a total flow of 504.7 kg/s (8000 gpm) at 28°C (82°F) to growing ponds. The average temperature drop is 3.4°C (6.2°F). Given the temperature drop and the flow rate, the annual end-use energy is 220 x 10^9 Btu/year. To estimate energy replaced, an initial water temperature of 15°C (60°F) is assumed —the average annual ambient temperature in the area —and the energy required to raise the temperature of 504.7 kg/s of water to 28°C is calculated. This results in a "replaced energy" value of 780 x 10^9 Btu/year.

In addition to the operating projects, Mitre tabulated other projects as follows:

Projects	No. Projects	10^9 Btu/year
Under development	42	4,010
On which feasibility studies had been performed	47	6,491
Proposed	197	17,496

Only 69 percent of the projects listed by Mitre as operating appear to be still in operation today. Of the projects under development, those which had been subject to feasibility studies, and those proposed, the percentage of projects in each category that have come to fruition so far are 50 percent, four percent, and eight percent, respectively.

Another study subsequently performed by Engineering and Economic Research Inc. (1982), based largely on Mitre data, did not separate or identify individual projects, but listed the types of applications and energy use for individual counties. In addition to providing a total energy use on-line figure of 12,542 x 10^9 Btu estimates were listed for projects planned through 1985 (12,276 x 10^9 Btu/year) and for projected applications after 1985 (10,653 x 10^9 Btu/year).

Of the total 166 on-line projects included, 64 percent appear to be on-going. Of 150 applications planned through 1985, 12 percent are on-line today, and six percent of the projects projected for beyond 1985 have come to fruition thus far. These numbers illustrate the high degree of optimism that was prevalent within the geothermal industry in the early 1980s.

In summary, the differences in results between this and previous assessments are due to the following factors:

1) abandoned projects
2) replaced vs. end-use energy calculations
3) exclusion of EOR projects.

OUTLOOK

The near future of geothermal direct use will depend strongly on tax credit issues. Unless Congress takes action to extend the federal geothermal energy tax credit, it will end on December 31, 1985. The Idaho state tax deduction of 15 percent also expires on the same day, and the credits in Oregon and New Mexico will be available only through 1989. Only the 15 percent credit in North Dakota has no expiration date. Thus, it appears that with this exception, the only state tax benefits remaining will be the exemption of geothermal systems from property tax assessments in Montana, Nevada, and Oregon.

What effect will the absence of the federal tax credit and other incentives exert on direct use of geothermal energy? This question is not easily answered. However, it appears to be the consensus of many in the direct use community that the tax credit has provided the major incentive for customers of operating geothermal district heating systems. While most customers enjoy considerable savings over their previous natural gas or electricity bills for heating once they are hooked up to a system, the up-front costs are quite large. In addition to the cost, project managers report that potential customers are inhibited from hooking up by their reluctance to accept the new technology until they have seen its effectiveness proven. Yet for a system to succeed economically, an adequate number of customers must be on-line as rapidly as possible.

With no incentive up front, it will be difficult to overcome both the financial and distrust factors. The manager of one district heating system that was initially built with DOE funding, but has expanded since that time and has built-in capacity for further expansion, fears that the effect of the tax credit expiration "could be close to devastating." "Any enticement helps," he said.

Lack of tax credit also removes incentive for the installation of one-well-one-residence space heating systems. The up-front cost of the well to supply the system and system modification may be subject to longer pay-back periods than would be acceptable to the average family without the tax advantage.

On the other hand, the opinions of individuals surveyed by Meridian Corporation in July 1984 just after Congress voted to let the tax credit expire are mixed on the effect of tax credit loss on the commercial/industrial use of geothermal energy. These uses are already inhibited by several factors including the high cost of money, the relatively low cost of competing fuels, and, in some cases, unacceptable simple payback periods — possibly up to 10 to 12 years without the tax credit. One view of this situation holds that lack of the credit "would pretty well stop commercial/industrial use." These uses are "right on the edge" of economic feasibility with the tax benefit, and "they just wouldn't happen without it," according to this school of thought.

The other view point holds that despite widespread complaints about losing the credit, the number of people affected by this action is probably small. While loss of the credit will be detrimental to marginal operations, it is felt by this group that if direct uses for commercial/industrial purposes are not sufficiently efficient to stand on their own, they should not be used. This group feels that it is a waste of the taxpayer's money to subsidize systems with long-term pay back periods, and that federal money would be better used by increasing system efficiency and reducing costs.

Both points of view agree that considerable interest has developed in commercial/industrial use of geothermal energy. They differ only on the degree to which that interest will be affected by the loss of the tax credit.

The various inhibitions to all forms of direct geothermal use may be offset somewhat by the fact that the technology is now proven and well known in "geothermal" areas, and should no longer be viewed with the skepticism of the past. In addition, the attractive economics reported for geothermal greenhouses and aquaculture projects may encourage expansion in those applications.

REFERENCES

Energy Information Administration, 1983, Regression analysis of energy consumption by end use: U.S. Department of Energy, Report DOE/EIA-0431, 116 p.

Engineering and Economic Research, Inc., 1982, Hydrothermal industrialization — direct heat development: U.S. Department of Energy, Report DOE/ID/12186-1, 20 p. Appendix.

Meridian Corporation, 1985, Geothermal direct use in the United States: Unpublished Working Paper MCD-004-85.

Mitre Corporation, 1981., Geothermal Progress Monitor, Report Number 5: U.S. Department of Energy, Report DOE/CE/0009/5, 93 p.

APPENDIX A

Geothermal Direct Heat Projects

Key To Symbols:

D: District Heating
G: Greenhouses
F: Fish Farming (Aquaculture)
R: Resorts
B: Bathing
I: Industrial Process Heat
S: Space Conditioning and Domestic Hot Water

**Assigned annual heat use value based upon secondary estimation methods.

( ) *Numbers in parentheses are estimates for projects under construction or undergoing expansion.*

Project		Utilization			
		Maximum		Annual Usage	
Locality	Type	Flow (kg/s)	Temp (°C)	Btu (x 10^9)	Comments
Alaska					
Melozi Hot Springs	R,G	8.2	55	0.3	Use includes space heating of a lodge, swimming pool, and a small greenhouse.
Circle Hot Springs	R	25.7	58	5**	Springs provide space heat for a 22 room hotel, 13 cabins, a pool, and other structures.
Manley Hot Springs	S,G,B	24	59	10**	Springs provide space heat for a commercial greenhouse, public bath, and a residence.
Chena Hot Springs	R,S	14.7	57	5**	Resort area with swimming pool, theraputic baths, space heated lodge, and a small greenhouse.
Other resorts, bath houses and individual users	R,B,S	NA	NA	15**	5 projects indicated
Total				35.3	9 operating
Arkansas					
Hot Springs National Park	S,R	30.7	68	2**	Space and water heating of administrative buildings, one bathhouse, and one hotel.
Total				2	1 operating
California					
Engler Fish Farms, Niland, Imperial County	F	0.01	71	.1**	Thermal water is used to raise catfish, talapia, and bass.
Crocker Enterprises, Salton City, Imperial County	F	NA	54	.1**	Thermal water is reportedly used for raising oysters.
Calistoga, Lake County	D	(DHHX)	115	(3.3)	A "mini-district heating system" is under construction.
City of Susanville, Lassen County	D	32	79	10** (10.7**)	District heating project includes 15 public and commercial buildings currently on-line plus 23 residences. Flow currently limited by injection well. Project being expanded.
Nursery at Susanville, Lassen County	G	19	60	10**	Discharge permit for greenhouses currently allows for only 300 gpm (18.9 kg/s) at surface. Expansion is planned.
Litchfield Correctional Center, Susanville, Lassen County	S	60	74	45	Facility uses geothermal water to provide space heat and hot water for water supply and laundry. Effluent is proposed to cascade to an adjacent commerce park.
Ramco Resources, Wendel-Amedee, Lassen County	G	NA	102	100**	Approximately 30 greenhouses are operating.
Fort Bidwell, Modoc County	D,G	1.9	47	1	District heating for community building, gymnasium, clinic, apartment complex, and a small greenhouse.
	F	18.9	33	10	Two fiberglass raceways are used for rearing catfish.
Cedarville High School and Elementary School, Modoc County	S	NA	54	(2)	Space and water heating for two school buildings (under construction).
Calistoga High School, Napa County	S	(DHHX)	78	2**	A local high school uses downhole heat exchangers to provide space and water heat.

Project		Utilization			
		Maximum		Annual Usage	
Locality	Type	Flow (kg/s)	Temp (°C)	Btu (x 10^9)	Comments
California (continued)					
Indian Valley Hospital at Greenville, Plumas County	S	3.2	47	1	Geothermal space and water heating system includes a hospital out-patient clinic, and a nearby office building.
Aquafarms International, Mecca, Riverside County	F	158	41	66**	Warm water from 3 geothermal wells combined with cold water from 5 irrigation wells is used in 61 aquaculture ponds that have a combined area of 50 acres. (Flow rate shown reflects *all* wells.)
Lake Elsinore Community Center, Riverside County	S	NA	NA	(0.4)	Project is under construction, production wells have been drilled.
Takashima Nurseries, Coachella, Riverside County	G	NA	49	20**	Approximately 10 acres of greenhouses are used for growing roses (operating 3 months yearly).
San Bernardino, San Bernardino County	I D	11	59	7.6** (120**)	Sewage digester/district heating system is undergoing expansion to include heating of 3 anaerobic sewage digesters and 52 public buildings.
Calaqua, Inc., Paso Robles, San Luis Obispo County	F	NA	49	45**	4 million catfish fingerlings are raised annually.
Indian Springs School, Big Bend, Shasta County	S	17.6	51	1.7	Geothermal space and water heating system supplies classroom, office building, auditorium and swimming pool.
Boyes Hot Springs Village Sonoma County	S	NA	66	(3)	Geothermal system is planned to provide space and water heat to 87 condominum units housed in 20 structures (under construction).
Other resorts, bath houses, and individual users.	R,B,S	NA	NA	10**	5 projects indicated
Total				329.5 (139.4)	19 operating 4 new projects under construction, 2 projects being expanded
Colorado					
Alamosa Shopping Center, Alamosa County	S	63	49	5**	Geothermal water is used to heat a mobile home park and a shopping mall (48,500 sq. ft.).
The Spa Motel, Pagosa Springs. Archuleta County	S	NA	54	2**	Geothermal water is used to heat a small motel, other buildings, a pool, and a dairy.
Pagosa Springs, Archuleta County	D	76	64	23.2	A district heating system provides space and water heat to 9 buildings and a high school.
Redstone Corp., Glenwood Springs, Garfield County	S	76	50	(3**)	Geothermal heat is provided to a 40,000 sq. ft. building and a pool (under construction).
Ouray Hot Springs, Ouray County	S	6.9	50	1**	Thermal water provides space and water heating for a motel, spa, pool, and municipal garage.
Other resorts, bath houses and individual users	R,B,S	NA	NA	20**	13 projects indicated
Total				51.2 (3)	17 operating 1 under construction
Idaho					
Boise City Geothermal District Heating System, Ada County	D	126	77	25	Geothermal water from 4 wells provides space and water heating to 20 government, institutional, and commercial buildings in the city of Boise.
Boise Warm Springs Water District, Ada County	D	52	77	30	A geothermal district heating system provides space and water heating to 250 private residences and a state laboratory. System has been in operation since 1892.
Veterans Administration Medical Center at Boise, Ada County	S	19	72	(20**)	Geothermal wells (2) have been drilled and heating systems for 21 buildings are undergoing retrofit design (under construction).

Project		Utilization			Comments
		Maximum		Annual Usage	
Locality	Type	Flow (kg/s)	Temp (°C)	Btu (x 10^9)	
Idaho (continued)					
Idaho State Capital Mall, Boise, Ada County	D	63	72	27	Geothermal district heating system provides space and water heat to the Statehouse, Hall of Mirrors, Len Jordan Office Building, Towers Office Building, Supreme Court, State Library, and a parking garage.
The Edwards' Greenhouse, Boise, Ada County	G,S	25.2	48	1**	Geothermal well provides space heat for a commercial greenhouse and residences.
Milstead Floral Greenhouse, Boise, Ada County	G,S	NA	47	3**	Geothermal wells provide space heat for a commercial greenhouse and residences.
Ketchum District Heating System, Blain County	D	63	70	11**	Thermal water from Guyer Hot Springs provides space and water heat to about 60 residences and a commercial establishment in the Ketchum-Sun Valley area.
Donlay Ranch Hot Springs, Boise County	G	4.4	55	0.5**	Thermal water is used for space heating at a small greenhouse.
Grimes Pass Hot Springs, Boise County	G	NA	48	0.5**	Thermal water is used for space heating at a small greenhouse.
Warm Springs Creek Hot Springs, Boise County	G	94.6	75	7**	Thermal water is used for space heating at Ward's Greenhouse.
Warm Springs Greenhouse, Garden Valley, Boise County	G	12.9	74	8**	Thermal water is used for space heating at the Warm Springs Greenhouse (40,000 sq. ft.).
Butte City Greenhouse, Butte County	G	NA	60	0.7**	Thermal water is reportedly used for space heating a greenhouse and irrigation.
Corral, Camas County	S,G	1.9	75	0.3	Thermal water is used for space heating 2 residences and a small greenhouse.
Hooper Elementary School, Caribou County	S	NA	27	3.4	Thermal water provides space conditioning to a 35,000 sq. ft. school.
Crook's Greenhouse, Cassia County	G	37.9	90	9**	Thermal water from a shallow well is used to space heat a greenhouse.
LDS Church at Almo, Cassia County	S	NA	34	1**	The Church of Jesus Christ of Latter Day Saints uses thermal water from a 500 ft. well to provide space heat and sidewalk defrosting for a 13,792 sq. ft. church building.
White Arrow Hot Springs, Gooding County	G,F	70.9	63	2**	Thermal water is reportedly used for a small research greenhouse and fish farm.
Malad High School, Oneida County	S	NA	28	5**	Thermal water provides space conditioning to a gymnasium and other areas of a high school.
Express Farms, Owyhee County	G	NA	37	0.7	Thermal water from Given's Hot Springs is used for space heating a greenhouse. Operations have recently begun producing tomatoes. Facilities consist of one 3600 sq. ft. greenhouse operating and another under construction.
Agri-Resources, Inc., Owyhee County	G	NA	43	4**	Low temperature thermal water provided by an artesian well is used for growing hydroponic lettuce.
Cookes Hot Spring, Owyhee County	G	28	83	0.5**	Thermal water is used to heat a greenhouse.
Buhl, Twin Falls County	S,G	22.1	45	1**	Thermal water provides space and water heating of a small greenhouse, residence, and private pool.
Fish Breeders of Idaho, Inc., Buhl, Twin Falls County	F	504.7	32	220	Low temperature thermal water at 32°C is mixed with cool surface water to provide 28°C water to aquaculture ponds for catfish and talapia. Thermal water is derived from multiple wells.
Buhl, Twin Falls County	F	25	32	10**	Robert Lunty uses thermal water for tropical fish raising.
College of Southern Idaho, Twin Falls County	S	75.1	39	18	Artesian thermal well provides water for space heating of 3 campus buildings with water-to-water heat pumps. Savings is estimated at $11,000 annually.
Banbury Hot Springs, Twin Falls County	S	95.5	32	1**	Thermal water from Banbury Hot Springs and shallow well(s) provide space and water heating at a resort.

Project		Utilization			Comments
		Maximum		Annual Usage	
Locality	Type	Flow (kg/s)	Temp (°C)	Btu (x 10^9)	
Idaho (continued)					
Flint Greenhouses, Twin Falls County	G	37.9	46	15**	Thermal water provides space heat for 87,000 sq. ft. of greenhouses.
M&L Greenhouses, Twin Falls County	G	NA	NA	10**	12 greenhouses are reportedly in operation.
Cal Flint Floral, Twin Falls County	G	20.8	71	10**	Thermal water provides space heat for 60,000 sq. ft. of greenhouses
Other resorts, bath houses and individual users	R,B,S	—	—	15**	48 projects indicated
Total				439.6 (20)	76 operating 1 under construction
Montana					
Warm Springs State Hopital Deer Lodge County	W	0.5	68	3	Thermal water is used for domestic water preheating at Warm Springs State Hospital.
Montana Lumber Company, Ennis, Madison County	S	1.8	35	0.5**	Thermal water provides radiant slab heating to a 4500 sq. ft. manufacturing building of a lumber company.
First National Bank, White Sulfur Springs, Meagher County	S	3.8	58	0.5**	A shallow thermal well provides water for space heating a bank.
White Sulfur Springs Motel, Meagher County	S	25	58	3**	Thermal springs heat a 23 unit motel and bath house.
Earth Energy Institute, Avon, Missoula County	G	0.8	32	0.2	Low temperature thermal water is used to heat a 900 sq. ft. greenhouse.
Other resorts, bath houses, and individual users	R,B,S	—	—	14**	10 projects indicated
Total				21.2	15 operating
Nevada					
Brady Hot Springs, Churchill County	A	44.9	132	85	Geothermal Food Processors, Inc. of Reno, Nevada uses geothermal water mainly for vegetable deyhdration. The plant produces 6 million dry pounds during the 6 month season. Energy savings is estimated at over $250,000 annually.
Elko Heat Company, Elko County	D,I	44.2	81	27	Elko Heat Company uses geothermal water to provide space and water heat to a four-story office complex, a hotel/casino complex and process heat for a commercial laundry. A bank and sewage plant have reportedly also become attached to the system.
City of Caliente, Lincoln County	D	69	81	8** (30)	Current uses include space and water heating of a trailer park, hospital, laundry facility, spa, and car wash. Proposed system is for district heating of 320 residences, 90 commercial buildings and 2 industrial sites.
Wabuska, Lyon County	F	37.8	104	37	Alexander Dawson Company is currently growing catfish and prawns. Thermal water at 104°C is mixed with 24°C water for use in growing ponds and raceways.
Moana Geothermal Area, Washoe County	S	NA	98	11	There are 150 homes, 2 motels and 2 churches heated by individual wells.
Warren Estates, Washoe County	D	9.5	96	4.3	Warren Estates is a 60 lot subdivision under development. Downhole heat exchanger system will be replaced by closed loop heat exchanger(s) using a supply and injection well. Energy estimate is based upon 100,000 Btu/hr peak load.

Project		Utilization			
		Maximum		Annual Usage	
Locality	Type	Flow (kg/s)	Temp (°C)	Btu (x 10^9)	Comments
Nevada (continued)					
	D	9.5	102	(29)	To provide district heating system to an additional 400 homes near Warren Estates via a supply well and injection well (under contruction).
Other resorts, bath houses, and individual users.	R,B,S			6**	7 projects indicated.
Total				178.3 (29) (30)	17 operating 1 under construction 1 under expansion
New Mexico					
New Mexico State University, Las Cruces, Dona Ana County	S	25.2	61	36	Geothermal system provides space and water heating of 30 campus buildings. A 12,000 sq. ft. research greenhouse is being designed.
NMSU University Center, Las Cruces, Dona Ana County	S	1.1	48	0.1	Thermal water is used for space heat and water preheat.
Las Alturas, Dona Ana County	W	NA	45	1**	Thermal water is used for domestic hot water supply.
Gila Hot Springs, Grant County	S,B	NA	64	5**	Hot springs are used for space and water heating of 15 buildings, swimming pools, and baths.
Apache Tejo and Kennecott Warm Springs, Grant County	I	NA	34	25**	Kennecott Corporation reportedly uses thermal water in copper processing at Hurly, New Mexico.
Burgett Floral, Animas, Hidalgo County	G	7.3	113	25**	250,000 sq. ft. of greenhouses are reportedly heated by cascading 60° C water from binary electric generators.
Beall Company Greenhouses, Animas, Hidalgo County	G	4.4	91	4.8** (1.2**)	Geothermal system is reported to supply space and water heating to two 24,000 sq. ft. greenhouses. System is presently undergoing expansion.
McCant Greenhouse, Animas, Hidalgo County	G	NA	91	1**	Thermal water is used for space heating a 10,000 sq. ft. greenhouse.
Jemez Springs, Sandoval County	D	1.3	71	1**	Geothermal space and water heating of residences, village hall, fire hall, and police station.
Truth or Consequences, Sierra County	S,B	NA	68	1**	Thermal water is used for space heating a community recreation center, several lodges and motels. Water is also used for swimming pools and baths.
City of Socorro, Socorro County	W	NA	43	10**	Two springs provide the major portion of the city hot water supply.
Other resorts, bath houses, and individual users	R,B,S	NA	NA	10**	9 projects indicated.
Total				119.9 (1.2)	20 operating 1 under expansion
New York					
Cayuga Community College and East Middle School, Auburn, Cayuga County	S	9.5	50	(7.5**)	Geothermal system is being installed to provide space and water heating to school buildings. Production well is complete, piping system is under construction, and an injection well is planned.
Total				(7.5)	1 under construction

Project		Utilization			
		Maximum		Annual Usage	
Locality	Type	Flow (kg/s)	Temp (°C)	Btu (x 10^9)	Comments
Oregon					
Klamath Falls Geothermal Area, Klamath County					Approximately 500 wells provide geothermal heat at 608 locations. 80 wells are pumped, the remainder use downhole heat exchangers. A small portion of the fluid produced from the pumped wells is reinjected. The total annual usage for Klamath Falls is 255 x 10^9 Btu.
	D	45.9	100	51**	A geothermal district heating system presently heats 11 government buildings in the downtown area.
					Geothermal water provides space heating and hot water to the following buildings:
	S	32	88	18**	1 Hospital
	S	NA	NA	2	3 Churches
	S	NA	NA	50**	10 Commercial Sites
	S	NA	NA	15**	13 Apartment Bldgs.
	S	NA	NA	27.5**	7 Schools
	S,F,G	47	89	54	OIT Campus
	S	NA	NA	37	570 Residences
	S	NA	NA	0.5**	2 Municipal Swimming Pools
Hunters Hot Springs, Lake County	S,G	3.2	93	3**	Thermal water is used for space heating at a motel, eight residences, and commercial greenhouses.
Breitenbush Hot Springs, Marion County	S,R,G	3.2	100	2**	Geothermal water is used to space heat resort cabins, swimming pool, bath house, and a small greenhouse.
Private Residences at Vale, Malheur County	S	NA	100	0.6**	Several wells supply geothermal water to private residences at Vale.
Agribusiness at Vale, Malheur County	G	18.9	104	13.7	Oregon Trail Mushroom Company operates a semi-automated mushroom growing and harvesting facility that is expected to reach a capacity of 3 million lbs. per year by mid-1985.
Cove, Union County	G,S	19.3	42	10**	Geothermal heated public swimming pool and a tree seedling nursery.
Other resorts, bath houses, and individual users	R,B,S	NA	NA	13**	8 projects indicated.
Total				297.3	52 operating
South Dakota					
St. Mary's Hospital at Pierre, Hughes County	S	23.7	42	11.4	Low temperature thermal water provides space heat for 2 hospital buildings (coupled with heat pumps).
Philip District Heating System, Haakon County	D	21.5	69	9.5	Thermal water provides space heating for 5 school buildings, 9 community buildings, and a 1-acre greenhouse. It is estimated that 75% to 90% of all building heat is supplied by geothermal water.
Diamond Ring Ranch, Haakon County	S,A	10.7	67	4.7	Thermal water provides space heating for six farm buildings and a grain drying facility.
Total				25.6	3 operating
Texas					
Marlin Chamber of Commerce Building, Falls County	S	1.3	48	1**	An old thermal well provides space and water heating for a 1500 sq. ft. retrofitted Chamber of Commerce building and a local spa.
T-H-S Memorial Hospital, Marlin, Falls County	S	10.1	67	11.5	Geothermal water from a production well supplies space and water heat to a retrofitted 130 bed hospital. It is estimated to have reduced consumption of natural gas by 61%.
Navarro College, Corsicana, Navarro County	S,F	NA	52	3.6	Geothermal water from a well provides space heating for campus buildings. Water is then cascaded to aquaculture ponds for raising prawns and catfish.
Total				16.1	3 operating

Project		Utilization			
		Maximum		Annual Usage	
Locality	Type	Flow (kg/s)	Temp (°C)	Btu (x 10^9)	Comments
Utah					
Newcastle, Iron County	G	NA	110	3**	Geothermal water is supplied to four 30 ft. x 130 ft. hydroponic greenhouses from a 500 ft. well.
Utah State Prison, Salt Lake County	S	32	81	10.9	Geothermal water is supplied from well(s) to heat a maximum security building at the Utah State Prison.
Utah Roses at Bluffdale, Salt Lake County	G	25	88	32	Geothermal water is supplied to a 130,000 sq. ft. greenhouse of Utah Roses, Inc. Disposal into an injection well.
Utah Roses at Sandy, Salt Lake County	G	11	51	33	Geothermal water is supplied to 260,000 sq. ft. of greenhouses of Utah Roses, Inc. from a deep (5000 ft.) production well.
Other resorts, bath houses, and individual users	R,B,S	NA	NA	6**	10 projects indicated.
Total				84.9	14 operating
Washington					
Sol Duc Hot Springs, Clallam County	S	9.3	50	11.8	Thermal springs provide space heating of a lodge.
City of Ephrata, Grant County	D	NA	29	(5**)	A grant from HUD has provided funds to develop a demonstration space heat projects utilizing heat pumps for a district heating system. System is under construction.
City of Yakima, Yakima County	S	NA	36	(1**)	Planned project using heat pumps to provide water heating to the county jail at Yakima.
Total				11.8 (6)	1 operating 2 under construction
Wyoming					
Jackson, Teton County	F	8.5	26	7.9**	Low temperature geothermal water is used at Jackson National Fish Hatchery, with ground water source heat pumps to provide space heat. Thermal water is mixed with cooler water for fish rearing.
Thermopolis, Hot Springs County	S,G	63.2	54	1**	Geothermal water is used to provide space heat to two private homes and a commercial greenhouse.
Lander, Fremont County	G,F	32	37	2**	Thermal water is used for space heating a 40'x80' commercial greenhouse, and water heating for tropical fish raising.
Other resorts, bath houses and individual users	R,B,S	NA	NA	1**	3 projects indicated.
Total				11.9	6 operating

Recommendations from Handbook Meeting, 5/5/87, Portland, OR

Format:
Book will be two columns per page, figures and tables one or two columns wide as required. We will use oversized paper for final copy which will then be reduced to about 70%. Specifics will be listed in the forthcoming sample page(s). Ben Lunis will have EG&G do graphics and **must have an estimate (ball park figure) from each author by 1 June '87 as to the number of graphics** in each chapter so he can attempt to procure funding from DOE for these. There may be a large folded resource map in a pocket inside the cover of the book perhaps taken from the USGS resource map. Other maps, charts, graphs, etc will be on regular sized paper, i.e. no fold out pages in the book. All chapters will be numbered individually, along with their tables, figures, etc. So, Chapter II pages will be numbered II-1, II-2, and the figures or tables will be Table II-1, Figure II-1, Figure II-2. To make it easier to find certain figures in a given chapter, each chapter will be identified by tabs possibly on the upper outside corner of front facing pages possibly in black background with white numbering. At any rate, some method of tabbing will be employed. An example page or two will be sent out to the authors outlining methods for equations, how units will be displayed, column widths, indention of paragraphs, spacing, references, etc.

Delivery dates as follows:
August 1, 1987 chapters to peers for review. Each author to choose peers, number of reviews basically based on material contained in the chapter, available peers for reviewing and time restrictions. Authors should have back from review around the first of September to allow time for updates and changes to be made.

October 1, 1987 peer edited proofs in for review by Paul Lienau and Ben Lunis.

Book to be printer ready by end of December 1987.

Technical Level:
The contents will be practical/technical. The chapters should answer the questions most commonly asked without going into too much detail. It will be designed to put the work back into the hands of the private sector. It should tell enough for the lay person to understand what needs to be done and what considerations should be made, what agencies to contact initially for information on procedures and permitting within their individual state.

Glossary:
There will be an index of glossary terms in the back of the book and

an individual glossary in each chapter. Authors initially responsible for identifying terms and definitions for their glossaries.

Content:
Each chapter will basically contain the following information -
- Introduction
- Description
- Explanation - text
- Vendors
- References (journals, text, personal contact)l
- Glossary

There is no page limit per section, but authors are cautioned to be brief and concise, don't go into minute detail. There will be no specific temperature limits set, but the entire handbook should deal basically with temperatures <90°C, and for sure under <150°C.

Units
There will be a conversion table in the front of the book for the whole handbook. The basic units of measure will be English.

Printing/Publishing
Printing will initially be done similarly to the GRC Special Report #7 handbook with perfect binding.

Word Processing Compatability
Ideally, we want to authors to have IBM compatible systems. The chapters can be sent from the author station to the Geo-Heat Center as we have a modem to accept files from your floppy onto our floppy. As a great deal of special symbols, like bold, underlining, etc will be lost due to differences in software, please forward a printed copy of your material through the mail at the same time the file is transmitted or with your floppy disk if it is mailed so Cindy can start cleaning up the file. The final processor will be a Panasonic KX-E828. Paul is working on having a program written that will allow IBM documents to be transferred onto floppies from the GHC's IBM to the Panasonic. Each author is requested to send a sample transmission, i.e. outline of their chapter to test the system at their earliest convenience.

General Discussion
Basically, only the western states identified below will be addressed: Alaska, Washington, Oregon, California, Idaho, Nevada, Arizona, New Mexico, Hawaii, Colorado, Utah, Wyoming, North Dakota, South Dakota, Montana and Texas.

It was generally agreed on that there needs to be a preface to the book separate from the Intro chapter that will be the "political appeasement statement" stating the intent of the book.

Chapter 17 will be modified to include industrial applications rather than just drying ag products. Something like: Industrial/Agricultural Applications.

There will be a published request for response of vendors to be included in the handbook. This will be done right away by Paul who will enter the notice in the GRC Bulletin, GHC Quarterly Bulletin and perhaps other related publications. All vendors responding will be included in a listing in the pertinent chapter. There will be a disclaimer stating all vendors who responded to our request are included but inclusion does not constitute recommendation or approval of products or services. Also, that this is by no means a complete list and that those listed at the time of the writing are not necessarily still in operation, etc. Ben know the parameters on this from DOE's viewpoint.

District heating was discussed. It was decided that Ben's PONS section will address district heating and possibly refer to the District Heating Handbook.

No color availability for graphics. Everything will be black and white.

The heat pump chapter, #13, will be industrial - not for individual home type situations.

Each chapter stands alone, but can refer to other chapters within the handbook. Your chapters can be addressed as a question and answer type of situation, etc. Gordon's chapter on Institutional, Legal and Permit will be addressed by State.

GEO-HEAT CENTER

Oregon Institute of Technology • Klamath Falls, Oregon 97601 • 503/882-6321 Ex. 267

Paul J. Lienau, Director

April 16, 1987

Mike Wright
UURI
391 Chipeta Way, Suite C
Salt Lake City, UT 84108

Dear Mike:

As a follow up to my letter of 12 March 1987, a meeting of writers for the "Geothermal Direct Heat Applications Handbook" has been scheduled at Portland, Oregon.

Place: Benson Hotel
309 SW Broadway
Portland, OR 97205
Date: Tuesday, 5 May 1987
Time: 9:00 am

The purpose of the meeting is for you to present an outline of your chapter(s) and to coordinate timetables, editing and publishing of the document. The agenda will include the following items and any others that you may bring to the meeting.

Agenda

a. Outline of contents
b. Project status to date
c. Review of chapter outlines by writers
d. Technical level
e. Format
f. Units
g. Graphics
h. Glossary
i. References
j. Services and equipment vendors
k. Word processing compatibility
l. Editing
m. Draft review
n. Printing/publishing
o. Timetable - completion, and
p. Publication by December 1987

q. Cascading applications?

If travel funds are a problem to attend this meeting, please let me know.

A Pacific Northwest Section GRC workshop will be held at the same hotel 6-7 May, 1987. A program is enclosed.

We are looking forward to seeing you in Portland, 5 May 1987.

Sincerely,

Paul

PAUL J. LIENAU
Director

/cn

enc

Who Should Attend

The workshop is organized to meet the needs of gas and electric utility management, planning, and marketing staffs; engineers; developers; and potential system owners and operators.

Location

The workshop will be held at the Benson Hotel in downtown Portland on May 6 and 7, 1987. The workshop location allows for site visits to operating water source heat pump systems. A block of rooms is reserved at the Benson Hotel. When you call the Benson, (503) 228-2000, to make your room reservations, mention that you are attending the Pacific Northwest Section Geothermal Resources Council. **The block of rooms will be held until April 22.**

Schedule

The workshop begins with registration at 9:00 a.m. on Wednesday, May 6.

REGISTRATION

Registration for the workshop is $25 per day (including lunch). Those wishing to register for both days may do so for $45. *Reservations for the workshop should be made no later than May 1. Although registrations will be accepted at the door, we cannot guarantee the availability of lunch unless you are registered by May 1. To register, call Marshall Gannett at (503) 378-8456, or R. Gordon Bloomquist at (206) 586-5074.*

Washington State Energy Office
400 Union S.E., First Floor
Olympia, WA 98504-1211

FIRST CLASS

OLYMPIA WASH APR -9 '87

U.S. POSTAGE .22

3319973

Paul Lienau
Geo-Heat Center
Oregon Institute of Technology
Klamath Falls, OR 97601

Water Source Heat Pumps

For Commercial, Industrial, and District Heating Applications

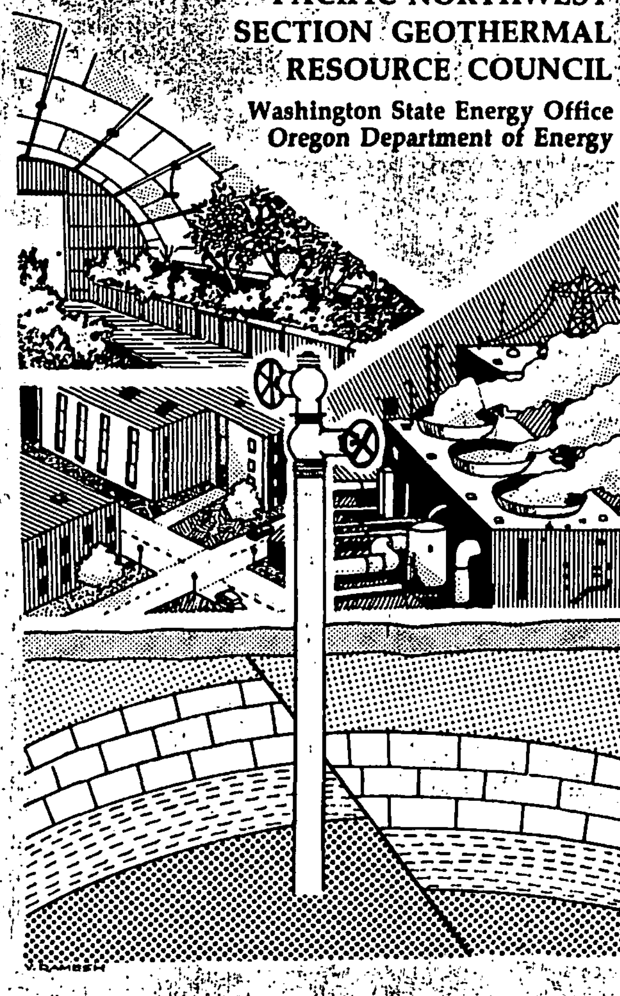
Technical and Regulatory Aspects

May 6-7, 1987

Benson Hotel, Portland, Oregon

PACIFIC NORTHWEST SECTION GEOTHERMAL RESOURCE COUNCIL

Washington State Energy Office
Oregon Department of Energy



The Workshop

The workshop focuses on the rapidly expanding use of large water source heat pumps (500 kW to 30 MW) in the commercial and industrial sectors, and as the primary heat source for district heating.

Large scale water source heat pumps have proven themselves in applications throughout Europe. In Sweden over 2,000 MW of heat pump capacity has been put on-line since the early 1980s. The U.S., though behind many of the European countries in reliance on heat pump technology, is rapidly discovering the advantages that heat pumps can provide. The main advantage is conservation of conventional energy resources while at the same time providing a means for orderly load growth by both electric and natural gas utilities.

The Program

Day one of this workshop will focus on the technical aspects of water source heat pump utilization, hydrothermal resources available for use with heat pumps, special resource considerations that apply to the use of geothermal resources, and operating experience from systems in the Northwest.

Day two of the workshop will emphasize regulatory aspects related to water source heat pump utilization and district heating. Topics include groundwater regulations, public utility implications of district heating operations, contracts and franchises, financing, and marketing.

Site Visits

Two short field trips to operating water source heat pump installations are planned for Wednesday afternoon, May 6, and Thursday afternoon, May 7.

MAY 6 PROGRAM

Marshall Gannett, Program Coordinator

9:00-10:00	Registration
10:00-10:30	PNWGRC Section Business and Roundtable
10:30-11:15	Introduction: *Why Heat Pumps?* Ed Knipe, Brown and Caldwell (tentative)
11:15-12:00	Groundwater Resource Considerations Marshall Gannett, Oregon Dept. of Water Resources
12:00- 1:00	Lunch (included in registration)
1:00- 1:30	Hydrothermal Sources for Large Heat Pumps Gunner Bringel, VBB-Allen
1:00- 2:00	Large Electrically Driven Heat Pumps ASEA STAL (speaker to be announced)
2:00- 2:30	Large Gas Driven Heat Pumps Dr. A.R. Maret, Gas Research Institute
2:30- 2:45	Break
2:45- 3:15	Lessons Learned: *Do's and Don'ts of Heat Pump Applications* Kevin Rafferty, Oregon Institute of Technology
3:15- 3:45	Operating Experiences: *A Case Study of the Commonwealth Building* Chet Bowlsby, Landsing Property Corporation
3:45-	Site Visit
5:00	Hosted Reception, VBB-Allen

MAY 7 PROGRAM

R. Gordon Bloomquist, Program Coordinator

8:30- 9:00	Registration
9:00- 9:30	Introduction: *Does District Heating Have a Role in the Northwest?* Alex Sifford, Oregon Dept. of Energy
9:30-10:00	Community Energy Planning: *Why District Heating?* Eliot Allen, VBB-Allen
10:00-10:30	Groundwater Regulation Marshall Gannett, Oregon Dept. of Water Resources
10:30-10:45	Break
10:45-11:15	Public Utility Implications: *Should District Heating be Deregulated?* John Nimmons, John Nimmons and Associates
11:15-11:45	A Developer's Perspective Jerry Huttrer, Geothermal Management Company
11:45- 1:00	Lunch (included in registration)
1:00- 1:30	Financing District Heating Projects Wally McOuat, McOuat and Associates
1:30- 2:00	District Heating Contracts and Franchises John Nimmons, John Nimmons and Associates
2:00- 2:30	Marketing District Heating: *The Key to Success!* Jim Barnes, Central Heating Distribution Limited

Meeting Notes:
Ben Lunis, Kevin Rafferty, Gordon Bloomquist, Paul L.
(5, 8, 9, 11, 14, 15 are draft or done)
1. Introduction & State of the Art — OIT
2. Program Opportunity Notices - Proj Summary — INEL

Audience for book:
- consulting engineers
- not necessarily geothermal background
- want to answer typical questions -- one big item is <u>costs</u>. w/ <u>vendors</u>.
- directed to technical audience.

<u>vendors</u> - its ok to cite specific suppliers, but <u>do not</u> <u>indicate</u> <u>preferences</u>.

Give <u>costs</u> <u>for</u> <u>various</u> <u>techniques/</u> <u>approaches</u>.

Project status to date - <u>Kevin</u> : Example <u>Absorption chapt</u>
Introduction
Short Description
Performance under various conditions
hot water req., energy consumed, etc.
Specific Example - chiller on OIT campus -
ties performance together w/ real $.
Description of Small vs Large Equipment
Commercial Refrigeration
Included Cost curves vs capacity - compare
to total electric chiller
Current Research.
Materials Selection
Software - put in <u>Appx.</u>? - in keeping w/ handbook approach.

Each chapter will stand alone re references, etc.
- want to emphasize practical approach.

Intro, Brief Description, state of Art, etc

chapter on Sampling is oriented to materials selection.

Possible examples
Haw Thorne
Utah State Prison
Red Hill -

also exploration / services vendors
Separate table @ chapter's end.
must not show preference -- must really dig
to get everyone.

Advertise for vendors in GRC Bulletin, GeoHeat Center Bull, Green Sheet, etc.

Occurrence & Description Chapter -

Restricted to 14 western states.
WA, OR, CA, AZ, NV, UT, MT, WY, CO, NM, SD, ND, HA, AK, ID, TX, NB

Graphics -
Lewis will chose DOE for $ for graphics so will look same. Maps will require special care.

Talk w/ DOE about reprinting some of state maps.

Get list of graphics to Lewis as soon as get chapter outlined.
Lewis will call in about ___ for this list just general so he can get feel for magnitude of job.

Units -
- in engineering, go with units,
- have conversion table for SI
- So we will use [customary common] units w/ conversion table.
- Only one conversion table -> so feed stuff to OIT
can have conversion table by category.

Figures & Tables: (Chapter. Number)
Fig 2.1, 2.2, etc.
Table etc.

Use 2-column format. Figs, Tables can be either 1 column or across page.

mail floppies to OIT w/ texts on.

Glossary - have one for each chapter
- also References.
glossaries by chapter w/ index to items in the Index.
glossary detail - things likely to be misunderstood or need spec def.

References stand alone each chapter.
names & date for reference. OIT will furnish format or use WGS.

word processing compatability -
Send floppy + printed copy

Send a Text file on floppy.

We each send floppy for test ASAP.

Editing -
@ draft stage, peer review, then come back for final edit.

So Cindy will put stuff on her Panasonic after re-edit after peer reviews.

We pick our reviewers. Peer review will be by chapter, selected by ~~us~~ authors.

Then after peer review + corrections, OIT will take charge of final reviews.
Then ~~editors~~ authors will see galley proofs.

Equations can be numbered.

Time tables - Shooting for Dec. 87 for sending to printer.

write
peer review
re-write → peer edited drafts

1 Oct. to OIT
1 Aug. to Peer review

PMW peer reviewers —
Isherwood
Huttner
Priest

pages
(double sp)

Nature of Low- and Moderate-Temperature Resources

3 - controls by structure and lithology

2 - geologic description

- conceptual models

3 - geochemistry of waters

solinities, common constituents, pH, etc.

+ environmentally sensitive elements

1 - range of temperatures

2 - recharge mechanisms - isotopic studies

11

Exploration for Resources

3 - ~~Surface expressions - springs, tuffa, sinter etc.~~

3 - ~~geothermometers~~

3 - ~~how to collect fluid samples~~

6 - geophysical expressions (typical)

6 - Selection of drill targets - expl strategy

3 - heat flow studies

- data compilation

2 - description of ~~[illegible]~~ "typical" systems

24

GEO-HEAT CENTER
Oregon Institute of Technology • Klamath Falls, Oregon 97601 • 503/882-6321 Ex. 267

Paul J. Lienau, Director

March 12, 1987

Dr. Phillip M. Wright
University of Utah Research Institute
391 Chipeta Way, Suite C
Salt Lake City, UT 84108

Dear Mike:

The Geo-Heat Center is coordinating the publication of a document on the technical aspects of geothermal direct use projects titled **"Geothermal Direct Heat Applications Handbook"**.

I wish to thank you for agreeing to contribute your expertise by writing a chapter on "Nature & Dist. of Geo. Resources" and on "Explor. for Geo. Resources".

In 1979, **"Direct Utilization of Geothermal Energy: A Technical Handbook"** was published. Since that time, a great deal of additional information and knowledge has been developed, including lessons learned, from the many projects that have come on line.

It is the feeling among GHC staff and USDOE that the new handbook should be directed towards a more technical reader (i.e. engineers and technical designers of future projects) than the original publication.

Preliminary guidelines that have been established for development of the handbook include:

a. Convene a meeting of writers to discuss an outline of contents and technical level - 5 May, Benson Hotel, Portland, OR.

b. Document will be directed towards technical readers.

c. Document will answer questions most frequently asked.

d. A plan for development of renewable energy projects, i.e. general steps.

e. Document may contain lists of services and equipment vendors with no endorsements.

f. Graphics may be done by EG&G Idaho, Inc.

g. In order to maintain continuity, a technical writer will compose the document from drafts of the various writers by mid-September. Ben Lunis, EG&G Idaho, Inc. and myself have accepted the responsibility as co-editors.

h. Document draft could be reviewed by experts in the field.

i. Completion and publication by December, 1987.

Chapters initially selected for the handbook include:

Geothermal Direct Heat Applications Handbook

	Chapter	Responsible
1.	Introduction & State of the Art	GHC Staff
2.	Program Opportunity Notice (PON) Project's Lessons Learned	Ben Lunis, EG&G Idaho, Inc.
3.	Nature & Distribution of Geothermal Resources	Mike Wright, UURI
4.	Exploration for Geothermal Resources	M. Wright, UURI
5.	Water Sampling Techniques	GHC Staff
6.	Drilling & Well Construction	GHC Staff
7.	Well Testing	Sally Benson, LBL
8.	Materials Selection	GHC Staff
9.	Well Pumps	GHC Staff
10.	Piping	GHC Staff
11.	Heat Exchangers	GHC Staff
12.	Space Heating Equipment	GHC Staff
13.	Heat Pumps	GHC Staff
14.	Absorption Refrigeration	GHC Staff
15.	Greenhouses	GHC Staff
16.	Aquaculture	GHC Staff
17.	Drying Agriculture Products	GHC Staff
18.	Engineering Cost Analysis	GHC Staff
19.	Institutional, Legal & Permit Requirements by State	Gordon Bloomquist, WSEO
20.	Environmental Aspects	G Bloomquist, WSEO

Please prepare an outline of your chapter's contents for review at the writers meeting tentatively set for 5 May 1987. If travel funds are a problem to attend this meeting at the Portland Benson Hotel or you have a conflict with 5 May 1987, please let me know. This meeting coincides with the GRC Section meeting on May 6.

Thank you again for your cooperation and we are looking forward to working with you on this project.

Sincerely,

Paul

PAUL J. LIENAU
Director

/cn

GEO-HEAT CENTER

Oregon Institute of Technology • Klamath Falls, Oregon 97601 • 503/882-6321 Ex. 267

Paul J. Lienau, Director

June 1, 1987

Phillip M. Wright
UURI
Earth Science Laboratory
391 Chipeta Way, Suite C
Salt Lake City, UT 84108-1295

Dear Mike:

Enclosed is a draft of the downhole heat exchanger portion of the heat exchanger chapter for the handbook. I'm requesting that you be a peer reviewer. While you may think review is more in the engineering area than geology or hydrology, the limitation on output and suitable resource characteristics is certainly geology/hydrology related. Please check my terminology, hydrology related calculations, etc and give me some ideas on improvements or additional ways to estimate outputs - especially the heat storage/recovery problem. I don't know how to solve that problem.

John Lund returned from Italy with very sketchy information about a use in Italy for greenhouse heating with GMW_t output and use in Turkey (Izmir) and Hungary for district heating. When/if I get more detailed information, it will be included.

Sincerely,

GENE CULVER
Associate Director

/cn

EXPLORATION FOR DIRECT HEAT RESOURCES

by

Phillip M. Wright
Earth Science Laboratory
University of Utah Research Institute

INTRODUCTION

Geothermal exploration may be divided into two types:

1. Exploration _for_ geothermal resources, that is, locating new geothermal resources; and,

2. Exploration _within_ geothermal resources, that is, defining the lateral and vertical boundaries and the internal properties of the reservoir.

In each case, one objective of the exploration work is to site wells that intersect the resource. Because drilling is usually expensive and because the present economics of most direct-heat applications will not support an extensive exploration program, it is important to design and execute exploration programs in the most efficient way possible. Many earth science techniques are applicable both to exploration for and exploration within geothermal systems, and in the discussion that follows, examples of both types of exploration will be given.

Siting successful geothermal wells is far from easy. Even within such well-known geothermal areas as The Geysers, California, where the experience of locating and drilling hundreds of wells is available, the success rate for production-well drilling is only about 80 percent. For wildcat geothermal drilling in relatively unknown areas, the success rate is much lower, perhaps only 10 to 20 per cent. The problem usually is not so much in finding heat as it is in finding fluids in amounts that are sufficient to supply a utilization system and to repay the costs of well drilling, testing and system installation and maintenance. In many geothermal reservoirs, this means drilling into one or more fractures that are connected to the source area for the geothermal fluids. Although large blocks of rock in nature are typically broken by fractures and faults, most of these breaks are not continuous enough to be connected with the source of fluids, and are thus not part of the reservoir per se even though they may be filled with thermal water. Because there is no known way to detect from the surface the particular permeable zones at depths of hundreds of meters that are connected to the reservoir, exploration techniques are mostly indirect and provide only circumstantial evidence of the existence and location of a reservoir.

Exploration consists of the application of various methods and techniques from the fields of geology, geochemistry and geophysics, with assistance from hydrology. Each of these fields is highly specialized, and because exploration can be quite expensive, it is important for the developer of geothermal resources to obtain the best consultants possible in these fields. The key is to be certain that the prospective consultant has both education and experience in locating and defining geothermal resources. The purpose of this chapter is to outline the principal methods used in geothermal exploration, how exploration strategies are devised and the costs for typical exploration programs. The methods and techniques we will consider can be grouped under the general headings of geological methods, geochemical methods and geophysical methods. We will consider techniques for study of resources of all temperatures, with emphasis on low- and moderate-temperature resources because these are the one usually considered for direct applications.

GEOLOGICAL METHODS

Collection of geologic data through surface geologic mapping, study of drill cuttings and core and laboratory work on surface and subsurface rock samples provides the basic information required for interpretation of geochemical, geophysical and hydrological data. Development of an adequate understanding of the regional and local geology should be the first step undertaken in any geothermal exploration program.

Geologic Mapping

Often ignored or shortchanged, geologic mapping and field evaluation of existing geologic maps is the important first step. The field geologist:

(1) identifies and locates on an air photo or a map the various rock units in the area (e.g. sedimentary rocks, plutonic rocks, volcanic rocks);

(2) maps the relationships among rock units (e.g. normal contacts, fault contacts);

(3) maps the structural elements of the geology (e.g. faults, fractures, folds);

(4) studies the relative age relationships among rock units as shown by their mutual field relationships;

(5) searches for evidence of geothermal activity, which evidence may range from obvious thermal springs, geysers and fumaroles to very subtle indications such as hydrothermal alteration of rocks or ancient or modern spring deposits of sinter (SiO_2) or travertine ($CaCO_3$);

(6) relates the geology of the particular prospecting area to the regional geology;

(7) collects samples of rocks and minerals for microscopic examination, age dating, geochemical analysis or geophysical characterization; and,

(8) collects samples of fluids from wells and springs for geochemical studies.

This work helps provide answers to such questions about the prospective geothermal area as:

(1) is there direct evidence of geothermal activity in the area?,

(2) are there young volcanic rocks, less than 1 million years old, in the area that would indicate an underlying molten or recently solidified rock mass to provide a source of heat?,

(3) are there porous and permeable rock units or are there active faults or open rock contacts that could together constitute a plumbing system?, and

(4) does the area have high potential for discovery of a geothermal resource and, if so, what exploration strategies and techniques should be used next?

Typically, geologic mapping will be done on high-quality black and white or color, stereo air photos. The mapped information will then be transferred to topographic map at an appropriate scale. Figure 1 is an example of a geologic map from the Roosevelt hot Springs geothermal area in Utah (Nielson x d). Roosevelt Hot Springs is an area of diverse rock types and complex faulting. A major high-temperature geothermal system underlies the central part of the mapped area, and a plume of hot water of decreasing temperature moves from the center of upwelling, in the general vicinity of the Opal Mound and Negro Mag faults, toward the northwest through the alluvium of the valley. This map is only a generalization of the detailed geologic map produced by the field geologist, but serves to indicate the type of mapping that should be done at the outset of detailed exploration in any geothermal prospect.

Study of Drill Samples and Information

In many geothermal exploration areas, exploration holes or wells have already been drilled, and samples of subsurface core or drill chips as well as driller's reports and well logs may be available. Such samples, reports and logs are usually stored with the state geological survey, the state water rights division and or the state oil and gas commission, and they become public information either soon after collection or after a specified confidential period. Some states require the bottom-hole temperature to be reported in all holes, providing information especially important for geothermal exploration. The geologist seeks and obtains all such available information in the initial stages of an exploration program. If drill chip samples are available, they are logged by the geologist and correlations of subsurface rock type made among holes and the surface. This work yields information on the three-dimensional distribution of potential reservoir rocks and on the geologic structure.

During the drilling phase of any exploration program, drill chips are cut from the bottom of the hole by the drill bit and are brought to the surface by the circulating drilling fluid. Many drillers will not take samples of the drill chips as they are removed from the drill mud stream at the shaker table unless instructed to do so. It is important to collect representative drill chip samples from any hole, and it is better to collect too many samples rather than too few. The driller should be instructed in how to take the samples by the geologist, who will be cognizant of the sampling requirements in different geologic environments. Typically, 1 to 2 lbs of sample will be collected each 10 to 20 feet of drilling. These samples will be carefully washed to remove the drill mud and then placed in geologic sample bags whose labels record information on the drill hole name, location, date and footage from which the sample was drilled.

Stratigraphic Studies

A thorough knowledge of the rock types in the prospecting area is fundamental, and is obtained through stratigraphic studies. The geologist analyzes both surface outcrops and rock samples from drilling. He strives to identify rocks in the area that would make a good reservoir rock at depth, i.e. rocks that have adequate primary permeability or in which secondary permeability may be developed. In a volcanic environment, for example, sequences of young lava flows often are highly permeable whereas air-fall or water-laid volcanic tuffs are easily altered to clay minerals and become impermeable. The permeability in flow sequences usually exists at the upper and lower boundaries or contacts of individual flows--the center portions of flows tend to be massive, with little primary permeability. In areas such as the Salton Trough, permeability is controlled by the type of rock (permeable sandstone or impermeable shale) and by its degree of metamorphism. High-temperature metamorphism causes the rocks to be brittle and to fracture under tectonic stress whereas low-temperature metamorphism may not induce brittleness. In some environments, permeability is developed in carbonate rocks (e.g., limestone, dolomite, marbled) by solution of the carbonate minerals by moving groundwater. It is obvious that an understanding of effects such as these is important to the success of a geothermal exploration project.

Structural Analysis

A thorough knowledge of the geologic structure of an area is important. Moore and Samberg (1979), for example, showed that at Cove Fort/Sulphurdale, Utah, much of the surface is covered by large landslide blocks that have moved into place from the east along an underlying, nearly horizontal faults. Subsequent movement has occurred along vertical faults, and the area now consists of numerous separate fault blocks. An obvious implication from this discovery is that the geology underlying the fault along which land sliding took place can not be determined from the surface geology. Realization of this fact has had a great impact on subsequent exploration in the area.

Faults can form zones of permeability if they fracture rock and create open spaces, or alternatively they can be filled with gouge, a rock flour generated during fault movement that is quite impermeable. Gouge developed along faults can isolate the aquifers in individual fault blocks and decrease hydrologic communication across an area. In places where faults intersect, permeability may be especially enhanced. It is important to determine the relative ages of faults and especially to be able to distinguish young faults and fractures from older ones and to distinguish faults that have had recent movement from those that have not. Older faults are more likely to have had their open spaces filled through deposition of minerals from circulating hydrothermal fluids. Relative ages of faulting can usually be determined through detailed geologic mapping.

Structural analysis is also important in other respects besides faulting. Recognition of volcanic structures such as calderas and vents is important in understanding the geologic evolution of an area and can suggest where subsurface heat sources and permeable zones may be found. In basin environments, knowledge of the shape and size of the basin and the location of faults can be used to predict depth to permeable horizons containing thermal water in advance of drill testing.

Age Dating of Rocks

Certain minerals contain the element potassium (K), and a small percentage will be the naturally radioactive isotope K-40. This isotope decays radioactively to argon (Ar-40), with a half-life of about 1.2 billion years. By measuring the amount of Ar-40 relative to the amount of K-40 in certain minerals, the time since Ar-40 began to accumulate can be determined. In this way, the time interval since formation, i.e. the age, of certain rocks can be determined. There are also other radioactive isotopes that can be similarly used for dating.

Age dating has obvious use in geothermal exploration in terms of helping to locate young igneous rocks. If volcanic rocks are found in an area that are less than about 1 million years old, they would indicate the likelihood of a heat source for geothermal energy.

One must be careful about interpretation of the dates derived by these methods. In the case of K-Ar dating, for example, if the mineral being used for the dating has been heated sufficiently by a thermal event subsequent to its formation, the gaseous Ar may escape, thus resetting the radioactive clock to the date of the thermal event. Determination of age dates is a highly specialized field. Most equipment for age dating is at universities. The University of Arizona has one of the most advanced laboratories for dating of samples of rocks younger than about 1 million years, which is the age range of most interest in geothermal exploration. The geologist should work with scientists in the laboratory that will do the dating in order to obtain advice on methods for collecting the most appropriate field samples of the rocks and minerals for dating.

GEOCHEMICAL STUDIES

A number of important exploration and reservoir production questions can be answered from studies of the chemistry of geothermal fluids and reservoir rocks, and so geochemistry plays a relatively important role in geothermal exploration and development (Henley and Ellis, 1983). Geochemical reconnaissance involves sampling and analyzing waters and gases from hot springs and other geothermal manifestations in the area under investigation. The data obtained are then used to help locate a geothermal system, to determine whether the geothermal system is hot-water or vapor-dominated, to estimate the minimum temperature expected at depth, to predict the homogeneity of water supply, to infer the chemical character of the waters at depth, and to determine the source of recharge water. Geochemical principles can also be applied to interpretation of chemical data from producing wells and may yield information on formation of scale in pipes or a gradual chemical change in the geothermal fluids that could indicate an impending change in production temperature. We will discuss some of the more important geochemical applications in this section.

Overview of Geothermal Geochemistry

Geothermal fluids contain a wide variety and concentration of dissolved constituents. The simplest chemical parameters often quoted to characterize geothermal fluids are:

1. Total dissolved solids (TDS) in parts per million (ppm) or milligrams per liter (mg/l). This gives a measure of the amount of chemical salts dissolved in the waters; and,

2. pH. The pH of a fluid is a measure of the acidity or alkalinity of the fluid. Neutral fluids have pH = 7 at room temperature. Acid fluids have lower pH values and alkaline fluids have pH values greater than 7.

These two parameters can be measured in the field by use of a conductivity meter and a pH meter. The conductivity meter measures the TDS of a fluid by measuring its electrical conductivity. The more dissolved salts, the higher the electrical conductivity.

The amount and nature of dissolved chemical species in geothermal fluids is a function of temperature and of the local geology (see Table 1 and Table 2). Lower-temperature resources usually have a smaller amount of dissolved solids than do higher temperature resources, although there are exceptions to this rule. TDS values range from a few hundred to more than 300,000 mg/l. Many of the high-temperature resources in the West contain 6,000 to 10,000 mg/l TDS, whereas a portion of the Imperial Valley, California resources are essentially saturated with salts at 300,000 mg/l. The pH of geothermal resources ranges from moderately alkaline (8.5) to moderately acid (5.5). The dissolved solids are usually composed mainly of sodium (Na), calcium (Ca), potassium (K), chlorine (Cl), silica (SiO2), sulfate (SO4), and bicarbonate (HCO3). Minor constituents include a wide range of elements with mercury (Hg), fluorine (F), boron (B) and arsenic (As) being toxic in high enough concentrations and therefore of environmental concern. In general, each state has regulations governing the use and disposal of waters that contain toxic or otherwise harmful constituents, and local regulations should always be consulted in planning the use of any geothermal resource. Dissolved gases usually include carbon dioxide (CO2), hydrogen sulfide (H2S), ammonia (NH4) and methane (CH4). Hydrogen sulfide (H2S) is a safety hazard because of its toxicity to animals. Effective means have been and are still being developed to handle the scaling, corrosion and environmental problems caused by dissolved constituents in geothermal fluids.

As geothermal fluids move through rocks, they react chemically with the rocks, which themselves are usually chemically complex. Certain minerals in the reservoir rocks may be selectively dissolved by the fluids while other minerals may be precipitated from solution or certain chemical elements from the fluid may substitute for certain other elements within a mineral. These chemical/mineralogical changes in the reservoir rocks may or may not cause volume changes, i.e., may or may not affect the permeability and porosity of the rocks. Obviously, if the mineral volume increases it must be at the expense of open space in the rock, which causes a decrease in permeability. In locations where pressure, temperature or rock chemistry change abruptly, minerals may be precipitated into the open spaces, resulting in plugging of the plumbing system. Silica and calcium carbonate (CaCO3) are the principal minerals usually involved. The solubility of SiO2 decreases with a decrease in temperature, with pressure changes having very little effect. SiO2 can be precipitated into open spaces such as fractures or pores in the rock in regions where the subsurface temperature changes abruptly and at the surface where hot springs discharge. Calcite (calcium carbonate) has a retrograde solubility, i.e., it is more soluble at low temperatures than at high temperatures. Other carbonate species such as dolomite (MgCO3), as well as sulfate species such as anhydrite (CaSO4), show similar retrograde solubility relationships with temperature. In addition, the solubility of carbonate minerals decreases rapidly with a decrease in the partial pressure of carbon dioxide. Thus, as fluids which are saturated with carbonate approach the surface, carbonate minerals such as calcite are deposited as a result of the loss of CO2, which exolves from the solution with the decrease in hydrostatic pressure.

The chemically complex hydrothermal system is dynamic through time, that is, for any given volume element in the reservoir, the fluid composition varies slowly with time, bringing about variation in the rock composition, porosity and permeability. However, because the rate of fluid circulation is perhaps only a few centimeters per year, in most hydrothermal systems a state of chemical equilibrium or near-equilibrium is observed to exist between reservoir fluid and reservoir rocks (Capuano and Cole, 1981; Helgeson, 1969). The assumption of chemical equilibrium is made in the application of several of the geochemical techniques discussed below. Lack of equilibrium could be evidence for rapid movement of fluid through the reservoir.

Chemistry of Geothermal Fluids

By taking appropriate samples of fluids from surface springs and from well discharges, a great deal can be learned about presence or absence of a geothermal resource and about the resource itself. The most important information can probably be obtained in the following topic areas:

1. Reservoir Fluid Types. Various fluid types evolve from typical geothermal systems, and identification of the fluid type can have important implications on the existence of other fluid types in the vicinity and, thus, on exploration;

2. Geothermometry. Chemical data can be used to the estimate the maximum subsurface fluid temperatures to be expected in a given area;

3. Reservoir Processes. The extent of mixing of thermal and non-thermal waters and boiling in the subsurface can be determined; and,

4. Production Monitoring. In a producing geothermal resource, monitoring of the concentrations of chemical species over time can lead to information of the nature of the recharge to the system and to the prediction of adverse temperature changes in advance of their

manifestation in the well.

Geothermal Fluid Types. As discussed in the chapter on the nature and occurrence of geothermal systems, the chemical compositions of geothermal fluids (Table 3) is a product of their mode of formation. Normal ground waters are usually near neutral in pH and slightly bicarbonate in character. When they are heated in a geothermal system, they tend to become more sodium chloride in character, with dissolved salt contents that can range from a few hundred mg/l to more than 300,000 mg/l. If the fluid boils at depth, gases (e.g. CO2, H2S) are partitioned into the steam phase and migrate independently toward the surface. The gas-rich steam phase may encounter cool groundwater, which is heated. Oxidation of H2S produces acid-sulfate waters which react with the rocks to produce characteristic advanced argillic alteration assemblages. Bicarbonate-rich geothermal waters are produced where groundwater dissolves CO2, rising with steam from the deeper geothermal system. Any of these water types may be diluted with low salinity ground water before being sampled from a thermal spring or by a drill hole. By study of the chemistry of the various waters found in a geothermal area, the nature of the (independent) geothermal reservoirs can be determined.

In the usual reconnaissance application, water samples are taken for analysis from springs and wells in the vicinity of the prospect. For detailed reservoir studies, fluid samples can also be taken from producing wells or from wells recently drilled but not yet producing. Great care must be taken to ensure that the samples contain only pristine reservoir, well or spring fluid. In the case of a recently drilled well, the well must be flowed until all traces of the drilling fluid have been removed. Often repeated samples are taken at intervals of hours, and when the analysis of these samples becomes constant, the fluid is assumed to represent reservoir fluid. Proper sampling technique is very important and should be entrusted only to someone with experience. The samples must be filtered and properly acidified for preservation until analysis. The sampler is cautioned to work with a chemist at the laboratory where the analyses will be done to design the sampling program. At each sample location, pH and temperature are measured at the time of collection.

Various systems have been devised to diagram water chemistry for better visual presentation (Hem, 1970). One of the most popular in geothermal work is a plotting method given by Piper (1944). This method is based on the relative amounts of Na + K, Mg, Ca, Cl + F, SO4 and HCO3 + CO3 in a fluid. These components are the major ions in thermal and non-thermal waters, and classifications based on them agree well with observations on the formation of various geothermal water types. To construct a Piper plot, also called a "trilinear plot", the concentrations of the cations and anions are transformed from units of ppm or mg/l into the units of milli-equivalents and the percentages of the cation and anion combinations as given above are plotted on a diagram similar to that shown in Figure 2. Any water analysis will contain cations that yield one point on the lower left portion of the diagram and anions that yield one point of the lower right portion. The cation and anion percentages are combined by projecting them onto the central rhombohedron, as illustrated with the water analysis shown as point A in Figure 2. The diagram can be used to plot all of the waters from a prospecting area, and a classification of water types then developed by comparing the result to the general classification diagram shown in Figure 3.

Figure 4 shows a classification of water types found at the Meager Creek geothermal area in southwestern British Columbia (Moore x). In this study, the authors showed that the several water types were chemically independent, i.e. that they had not evolved from a common water type nor had one from any other. This result implies that there is no through-going permeability in the part of the Meager Creek area explored, and that the several waters originate in unconnected, probable small reservoirs. Murray at al. (1985) show a Piper plot of waters from the Calistoga geothermal area in the Napa valley of California (Figure 5). Their results suggest that thermal water rises along a central fault in the valley. The thermal water is progressively diluted with non-thermal groundwater, gradually becoming enriched in iron, sulfate and bicarbonate. The several water types in the valley can all be related chemically to the thermal water seeping up the central fault, and water geochemistry can be used to trace and map geothermal waters in the valley.

Chemical Geothermometry. Chemical analyses of geothermal fluids can sometimes be used to estimate subsurface reservoir temperature. This information is of obvious interest during exploration, when information from measurements in drill holes may be unavailable, but it is also very important during drilling because (1) accurate temperature measurements cannot be made in a well until after thermal effects of the drilling process have been dissipated, weeks to months after drilling is finished, and (2) chemical geothermometry may indicate that temperatures higher than those found in the drill hole may be found elsewhere. Chemical geothermometry is not able to determine the location or depth of the highest-temperature reservoir fluids, but only that they exist somewhere in the vicinity.

Both quantitative and qualitative geothermometers are available. The basic assumptions in application of quantitative geothermometers are that equilibrium has been reached in temperature-dependent chemical reactions between the reservoir rock and the fluid, and that no changes occur in the fluid after it leaves the reservoir and is sampled by a well or at the site of a natural spring. In this case, the chemistry of the fluid will reflect the chemical equilibrium at reservoir temperature, and analysis with subsequent interpretation yields an estimate of this temperature.

Several major-element geothermometers have been used successfully for estimating subsurface temperature, and reviews of these geothermometers were given by Fournier (1981) and by Henley et al (1984). In certain geothermal areas, the silica content of geothermal fluids appears to be limited above about 180 deg C by the solubility of the mineral quartz (SiO_2) and to be limited below 180 deg C by the solubility of amorphous silica. Both solubilities are temperature dependent, as shown by Figure 6, which gives a graph of the solubility of various silica phases versus temperature. Table 4 gives some of the silica geothermometer equations. Other silica geothermometers are based upon equilibrium with the minerals chalcedony, a-cristobalite, or b-cristobalite, and it is obviously of importance to know which silica minerals exist in the reservoir rocks. If drill information is not available on this point, as it usually the case early in an exploration program, one must rely on the geologic mapping and inference to provide this information.

A second system of geothermometers is based upon the equilibrium reached among sodium (Na), potassium (K) and calcium (Ca) where reservoir rocks contain abundant quartz and feldspar (Fournier and Truesdell, 1973). Common geothermometers of this class are also shown in Table 4.

Different geothermometers frequently give different results when applied to the same fluid. Use of other data may help shed light on the relative reliability of the various geothermometers in specific geologic situations. For example, silica concentration can be affected by the pH of the fluid, and temperatures calculated from the Na-K-Ca geothermometer may be in serious error if the CO_2 or magnesium concentrations are too high or if there has been addition of any of these elements through interaction of the fluid with sedimentary rocks or ion-exchanging minerals such as clays or zeolites. Mixing of the thermal reservoir waters with normal groundwater can also change concentrations of the critical elements in a geothermometer, and can result in a calculated temperature that is either too high or too low. In addition, some geothermometers do not work well where reservoir temperatures are below about 150 deg C.

Care must obviously be taken in interpretation of chemical geothermometer data, and in this matter there is no good substitute for experience. Anyone can apply the geothermometer equations to the chemical analyses, but the interpretation of the results can be extremely involved, and is best left to appropriate experts.

Age Dating of Geothermal Waters. Radioactive isotope chemistry has been used to attempt to determine the age of the water in geothermal systems using techniques similar to those for dating of rocks. The most successful applications have used tritium (H-3, see Table 5) which has a half life of 12.26 years. Minor amounts of tritium are naturally produced continually in the stratosphere by the action of cosmic radiation on hydrogen in the air. However, major amounts of tritium have been put into the atmosphere by thermonuclear weapons testing. Tritium concentration is expressed in terms of the Tritium Unit (TU) which is equivalent to a ratio of tritium to hydrogen-1 of 1 x 10 E-18. In continental climates in the temperate zone, cosmic radiation produces about 10 TU. As many as 10,000 TU were measured in 1963 following extensive atmospheric weapons testing. Ambient tritium levels thereafter decreased until about 1968, and since then have remained fairly constant. The following generalizations can be made concerning the age of geothermal water in the absence of mixing. A tritium content of less than 3 TU indicates that no water younger than 25 years is present. Values of 3 to 20 TU suggest that some amount of thermonuclear tritium is present, which indicates that the fluids entered the groundwater environment in the 1954-1961 time frame. If more than 20 TUs are found, the water entered the system after 1963. Many geothermal reservoir waters are older, some much older, than the 25 to 50 years of useful dating range available with tritium. Typically, convecting hydrothermal fluids move at speeds measured in feet or tens of feet per year. However, tritium dating of water can indicate rapid movement in a system.

Production Monitoring. It is important to obtain and analyze samples from geothermal production wells on a periodic basis beginning at the start of production. By collecting a history of production chemistry data, processes and changes in the reservoir can be more easily understood and predicted. Samples should be taken and analyzed from each production well monthly in large, high-production systems, and quarterly in smaller systems. These intervals are meant to be simple guidelines. The geothermal developer should obtain competent consulting help in designing and carrying out a sampling and analysis program.

Geochemistry of Rocks

Rocks contain a variety of chemical elements that make up the minerals. Most minerals are made from only a few elements, and these are known as the "major elements" in the rock. Elements that occur in the approximate range 0.5 per cent to 0.05 per cent are usually termed "minor elements", whereas elements that occur in the parts-per-million range are termed "trace elements". Analysis for major elements is sometimes done for rock identification purposes. Minor and trace elements are studied for their geothermal exploration implications.

Major-Element Chemistry. At times, the major-element chemistry of rocks is used to identify the rock type. If there are no good samples of the whole rock available, as when drilling produces only tiny chips, or in cases where rock type can not be pinned down by hand-specimen or microscopic observation or from X-ray identification of minerals, a complete chemical analysis of the rock may be made for the purpose of rock type identification. The major rock-forming elements analyzed include silicon, aluminum, iron, magnesium, sodium, and potassium. Analytical results for these elements are generally expressed as percentages of oxides of the metals. Table 6 shows typical analyses for several rock types.

Minor- and Trace-Element Zoning. As hydrothermal fluids circulate in a geothermal system, they pick up, carry and then deposit (i.e., they redistribute) elements. The mobilities of various elements varies, so that some elements are not mobilized at all by the geothermal fluids whereas others become highly mobile in the fluids and are carried long distances. The more mobile elements can be used in exploration by analyzing for them in rock or soil samples and using their presence to indicate geothermal activity in the vicinity. For example, mercury, arsenic, manganese and zinc are all quite mobile in geothermal fluids, even those of low and moderate temperature. Soil sample surveys of a prospect area are sometimes used to determine where the geothermal potential is highest and to locate faults and fractures along which geothermal fluids have moved. Figures 7a and 7b show the distributions of mercury and arsenic, respectively, in soil samples from the Roosevelt Hot Springs geothermal system (Capuano x). Arsenic concentrations seem to outline areas where subsurface transport of geothermal fluids has brought this element into the shallow subsurface along faults. Mercury, however, is more restricted in its distribution than arsenic. Mercury is much more volatile than arsenic and decreases in quantity in the soil if fluid flow stops in the faults below. These results are interpreted to indicate that arsenic and mercury are transported together in the geothermal fluid, and deposit in trace amounts in the soil above geothermal conduits. Arsenic remains in the soil after the subsurface supply ceases, but mercury concentrations decrease with time due to the higher volatility of this element. Thus, mercury appears to indicate areas where the faults are open and carrying fluids in the shallow subsurface today. In this respect, mercury could be a powerful exploration tool for geothermal resources of all temperatures.

Trexler et al (1980) found that a soil mercury survey of the Caliente, Nevada area did a reasonable job of outlining areas known to be thermally anomalous from well temperature and 2 meter temperature surveys (see Geophysics - Thermal Methods in this chapter). Matlick and Buseck (1975) also outlined uses of mercury geochemistry in geothermal exploration. They show data for mercury distribution in four geothermal areas -- Long Valley and East Mesa, California, and Summer Lake basin and Klamath Falls, Oregon. In each area the mercury survey work outlined known and suspected areas of present-day geothermal activity quite well. Figure 8 is a profile of the mercury anomaly over and adjacent to the area of high heat flow at Klamath Falls. Note that mercury values increase about ten-fold over the high heat flow area relative to background values.

In conducting soil geochemical surveys, one must be careful to sample consistently a certain, chosen soil horizon in order to obtain the most meaningful results. Noise in the survey results can be generated from inconsistent sampling. Matlick and Buseck (1975) advocate the consistent use of the A soil horizon for mercury surveys. It is advisable to work with a geochemist who is experienced in soil sampling surveys for collecting and interpreting survey results.

Soil gases can also be used for locating faults, fractures or other permeable horizons that are open and carrying geothermal fluids at depth. Radon, carbon dioxide, hydrogen sulfide, mercury and the noble gases all move freely in permeable zones, gradually working their way to the surface, where they are naturally discharged into the atmosphere. Sensitive detectors can be used to measure their abundance, which should increase above a fluid-carrying fault or other zone. Exploration and production drill holes can sometimes be sited using such survey information.

Soil gas surveys are made by plunging a stainless steel tube into the soil to a specified depth and extracting a measured amount of soil gas by suction. The gas is then analyzed. Radon is a naturally radioactive gas that migrates to the surface through transport in waters and movement as a gas in open spaces. Its presence can be quantitatively determined using a track etching method. A strip of cellulose nitrate film is taped inside a plastic cup and inverted cups are placed in shallow holes and left for periods up to several weeks. Radon gas reaches the cup and collects, and as it decays radioactively, the nitrate film records the passage of subatomic particles as microscopic tracks. Acid etching of the film after the cups are retrieved makes the tracks visible, and they are counted under a microscope. One advantage of this technique is that is provides an integrating effect over the time the cups are in the ground, which tends to reduce the effects of changes in atmospheric pressure and soil moisture on radon content in the soil. Because radon is so highly mobile in the geologic environment, such surveys can be used to locate faults and zones of upwelling fluids (Nielson x).

<u>Mineral Zoning.</u> Study of minerals deposited in the subsurface plumbing of geothermal systems helps to map the boundaries of such systems, determine the temperatures at which the minerals formed and locate zones of upwelling and recharge. It is usually the assemblage of minerals rather than the occurrence of individual minerals which is most diagnostic of the zoning in geothermal systems. Different mineral assemblages are

formed in response to changing temperatures, changing rock chemistry (mineralogy), and other factors, but the major effect is one of temperature. The hydrothermal mineral assemblages of active geothermal systems are dominated by clays or zeolites at relatively low temperatures, and by chlorite, illite, K-feldspar and epidote or wairakite at higher temperatures (Table 7). Quartz, calcite, pyrite and anhydrite are frequently associated with these minerals, and appear to form readily at both high and low temperatures. The distributions of the clay and silicate minerals is strongly temperature-dependent. At the lowest temperatures, below about 180 deg C, the stable assemblage consists of dolomite, kaolinite, smectite and interlayered illite/smectite. With increasing temperature and depth, smectite, dolomite, kaolinite, and interlayered illite/smectite disappear, and at temperatures above about 150-180 deg C, the typical assemblage is illite, chlorite, potassium-feldspar and quartz. The calcium-aluminosilicates, wairakite and epidote appear only in rocks above 230-250 deg C. One very important result of this mineral zoning is that the higher-temperature mineral assemblages cause the rocks to become brittle, and they fracture easily under the influence of tectonic movement and stress. Faulting of brittle rock creates and renews fracture permeability in the higher-temperature parts of some systems.

The interpretation of the mineral assemblages found in many thermal systems is complicated by the presence of minerals formed during earlier, frequently unrelated, hydrothermal events. The Roosevelt Hot Springs thermal system provides a situation where at least two distinct hydrothermal events can be recognized; an earlier event related to intrusion of the Tertiary Mineral Mountains pluton, and the present hydrothermal system (Nielson et al., 1978). Cross-cutting veins, identified in drill chips suggest that the depositional histories of these events was complex.

Acid sulphate springs are typically a surficial feature produced by the oxidation of hydrogen sulfide to sulfuric acid in the near surface zone. Altered ground surrounding the acid springs and fumaroles provides a striking example of reactivity of the waters some geothermal waters. The altered areas are typically bleached and converted to a siliceous residue containing native sulfur, cinnabar (mercury sulfide), yellow sulfate minerals, yellow, red and brown iron oxide minerals and clay minerals. Such acid altered areas would be recognized by the geologist as an indication of past or present geothermal activity. Similar acid alteration can also be formed at depths where steam heating of groundwaters occur.

<u>Fluid Inclusion Studies.</u> In the process of the formation of minerals from hydrothermal fluids circulating in the fractures and pore spaces of geothermal systems, tiny amounts of the fluids themselves become trapped when the mineral grows around them. These fluids thenceforth exist as microscopic bubbles in the

minerals known as "fluid inclusions". Study of fluid inclusions is useful in geothermal exploration because they contain a sample of the fluids that formed the mineral in which they are found, and the information derived can tell about the formation and evolution of the geothermal system.

Typical measurements in study of fluid inclusions involve heating and cooling of the mineral specimen. The mineral is first hand-selected from a sample of drill chips or core and mounted on a microscope slide. Under the microscope, fluid inclusions appear as flaws in the mineral that are bounded by an outline that may be regular or irregular and which generally contain a bubble. The bubble is a vacuum bubble resulting from fluid contraction as the mineral cooled from its temperature of formation. By heating the mineral under the microscope and measuring the temperature at which the bubble disappears (i.e., the fluid inclusion becomes completely filled), the temperature for formation of the mineral can be determined. Also, by cooling the mineral under the microscope until the fluid freezes, the salinity of the fluid in milligrams per liter can be determined, since dissolved salts lower the freezing point of a solution by known amounts. The temperatures of formation yield information on, among other things, whether the system has cooled down or warmed up since the mineral formed. Information of fluid salinities provide information on system evolution when compared to present-day fluids.

Isotope Studies

Several stable isotopes are used in chemical studies of geothermal systems. Isotopes of a chemical element are separate species of the same element that have different numbers of neutrons in the nucleus. Stable isotopes are those that do not decay radioactively. The stable isotopes most often of help in geothermal studies are hydrogen-2 or deuterium and oxygen-18. Table 5 lists these isotopes and shows how they relate to the other isotopes of these elements. The percentages of each isotope distributed in nature are well known from many measurements. However, there are geochemical and geological processes that can cause the relative percentages to change. Among these processes are boiling in a geothermal system, chemical reactions between water and rock, mixing of different fluids, filtration through shales and changes in the state of oxidation. Figure 9 illustrates on a plot of change in oxygen-18 versus deuterium the expected direction of change for the hydrogen and oxygen in water due to these processes. We see from this plot that the effects of chemical interaction between geothermal waters and reservoir rocks is expected to lead to enrichment in O-18 in the water, with little or no change in deuterium. The explanation is that there are few hydrogen-containing minerals in reservoir rocks, and thus the deuterium in the water has few minerals to interact with, and remains constant. However, there are many oxygen-containing minerals in reservoir rocks, and these minerals are typically enriched in O-18. Figure 10 illustrates observed O-18 enrichment in waters from some typical geothermal systems having a variety of temperatures. The data points for O-18 extend right from the meteoric water line. The meteoric water line shows the average composition in deuterium and O-18 for rainwater in the area. It has been observed that the deuterium content of rain water is dependent on the altitude at which the rain falls. It has also been observed that the fluids of hydrothermal convection systems are predominately meteoric in origin (Craig, 1963). These observations, coupled with measurements of the deuterium content of hot springs and reservoir waters discharged from wells, can sometimes be used to help determine the altitude at which the reservoir waters originated, and thus the source area for the recharge waters.

Isotopic studies can also help answer questions on bulk reservoir permeability. Figure 11 illustrates the concept of this application. If the ratio of water to rock in the system is large, i.e the system is highly permeable, the isotopes in the water will not be changed much because of the relatively large amount of water, whereas the isotopes in the rock will show large shift. However, if the ratio of water to rock is small, i.e. the system is of low permeability, the isotopic composition of the water will be shifted markedly while that of the rock will not be appreciably changed. Thus, by characterizing the O-18 and deuterium compositions of both the water and the reservoir rock, a crude estimate of the bulk permeability of the system can be obtained.

GEOPHYSICAL STUDIES

Geophysical exploration is the use of physical measurements either to detect a resource directly, to provide indirect evidence of its existence and location or to determine and map its physical and chemical characteristics. Such physical parameters as the distribution of temperature over the surface of the earth and at depth, the electrical, magnetic or density properties of the ground, and the manner in which elastic waves are propagated in the earth all respond in their own way to the presence of a geothermal resource (Ward, 1983; Wright et al., 1985) or to an ore body, a coal deposit or a petroleum reservoir. Geophysical surveys are used in geothermal exploration and reservoir mapping to help locate resources that have no evident surface expression, to help site production and injection wells and to monitor production from a reservoir.

Reliable interpretation of geophysical survey data requires an understanding of the geology of the prospecting area. Interpretation is always a two-step process:

1. The geophysical field data are interpreted in terms of subsurface variations in the physical or chemical

property to which the method responds; and,

2. Subsurface physical or chemical property variations are interpreted in terms of the geology.

The first step requires an experienced geophysicist who has access to computer-based interpretation aids. From it, one can determine the locations and depths beneath the surface to changes or contrasts in the physical or chemical property to which the particular geophysical responds. The second step is one to be taken by the geophysicist working closely with the geologist. Often, interpretation procedures must be repeated several times until the subsurface model both explains the field geophysical data and agrees with the geologic data. Reliable interpretation is as much an art as it is a science. Understanding of the method, care and experience are all ingredients to successful interpretation.

Thermal Methods

Thermal methods involve the measurement of subsurface temperature and heat flow in drill holes. Under suitable circumstances, geothermal resources can be detected directly by application of these methods.

Thermal Gradient and Heat Flow Studies. Apparatus to measure subsurface temperature consists of a sensitive thermometer probe capable of measuring temperature differences of about 0.01 deg C (0.005 deg F), several hundred to several thousand feet of logging cable for lowering the probe down a borehole and a winch to handle the cable. Small units for shallow holes can be highly portable whereas more sophisticated, deep-hole units must be truck mounted. One commercially available unit with 4,000 ft of cable has a total weight under 50 lbs and can easily be used by one person.

Making a temperature log of a bore hole consists of lowering the probe down the hole and making temperature measurements at certain intervals. Holes are generally logged from the top downward to avoid the temperature effects of mixing of the water as the probe and cable descend. A typical interval between measurements is 10 to 30 feet. At each measuring depth, the probe is left motionless for a time of a few seconds to several minutes to allow the thermometer in the probe to come to thermal equilibrium with the surroundings at that depth. The operator can determine when equilibrium is reached because the measured temperature will cease to change with time. Some units measure temperature continuously as the probe descends. For these types of surveys, one must be sure that the logging speed in slow enough to allow the probe to reach thermal equilibrium at each point.

The drill holes themselves require a certain amount of time to come to thermal equilibrium after drilling, because the circulation of drilling muds and other drilling processes cause severe disturbance to the temperature. Temperature logs taken before the hole has reached equilibrium after drilling can be useful in locating zones of inflow and outflow, but one must remember that the absolute values of the temperature will not be correct. Generally, circulating drill mud causes the rock temperature to be lowered at the bottom of the hole because the mud temperature will be lower than the ambient rock temperature and the mud will remove heat from the bottom of the hole. Since the mud is heated at the bottom of the hole, it tends to heat the rocks in the upper portions of the hole as it rises toward the surface. Thus, directly after drilling, the upper parts of the hole will be warmer than their equilibrium temperature while the lower parts of the hole will be cooled. The amount of temperature disturbance depends on the length of time required to drill the hole, among other parameters. These relationships are complex, and it is not always possible to predict exactly the length of time required for a hole to reach equilibrium. Repeated logging at intervals of a week to several months allows one to obtain an equilibrium, temperature profile.

Figure 12 shows temperature profiles taken by logging in several holes in the Newberry volcano area, near Bend, Oregon. Several aspects of these profiles are of note. The uppermost part of each hole is disturbed by seasonal temperature changes. Below depths of about 100 ft, these seasonal effects are damped out and cause no further problem. (Describe x).

One basic parameter of interest in geothermal exploration is the heat flow, the rate at which heat flows upward toward the surface. We have seen in the chapter on nature and occurrence of geothermal resources that outward flow of heat from the earth's interior is a worldwide phenomenon. In geothermal areas, the heat flow is higher than the worldwide background, and anomalously high heat-flow values may be clues to underlying geothermal resources.

The vertical heat flow in an interval in a drill hole is determined by multiplying the measured value of the temperature gradient in that interval, $(T2-T1)/(D2-D1)$, by the thermal conductivity, K, of the rocks in the interval. Here, T1 refers to the measured temperature at depth D1 and T2 refers to the temperature at depth D2. The thermal conductivity must be measured in the laboratory on rock samples from the hole as there is no suitable down-hole probe for its determination. In the absence of hydrologic effects, which tend to distort the pattern of heat flow, the calculated heat flow should be the same in different intervals in the hole, although the temperature gradient and the thermal conductivity may both vary. Thermal conductivity is a function of the minerals and the porosity in the rock. Table 8 summarizes thermal conductivity ranges for typical rocks, but the particular rocks in a borehole may vary from these values. Thus, in making heat flow surveys, measurements should always be made on actual rocks from the borehole. We can see from the heat-flow equation, Q = K x (T2-T1)/(D2-D1), that if we assume a constant heat flow, and if the thermal conductivity in an interval is high, the temperature gradient in that interval will be low. If the thermal conductivity is low, the temperature gradient will be high. Since Table 8 shows thermal conductivities for rocks to range over more than a factor of two, temperature gradients can have a range of more than a factor of two in an area due to the effects of varying thermal conductivity.

An often applied but dangerous shortcut to heat flow surveys is to forgo measurement of thermal conductivity, perhaps to save the cost of good sample collection and laboratory measurement, and to obtain only the thermal gradient data. Lateral as well as vertical variation of the temperature gradient could be due either to genuine changes in the heat-flow field or simply to changes in rock type that affect the thermal conductivity and are unrelated to any geothermal resource. Temperature gradient surveys by themselves are never as useful as taking the extra step of determining thermal conductivity and calculating heat flow values, too. Furthermore, one often sees temperature data extrapolated below the bottom of the hole as a prediction of how deep one must drill to achieve a specified temperature. We conclude from this discussion that extrapolation of an observed temperature gradient to levels below the borehole generally can not be done reliably because one will not generally know or be able to account for the variations in thermal conductivity beneath the hole. Most predictions of temperature at depths below the drill hole only lead to disappointment, and we can see the reasons why such predictions are not to be relied upon.

Drilling can be expensive, and so the natural tendency is to use thermal gradient or heat flow holes that are as shallow as possible. It is desirable to make the temperature measurements below the level affected by seasonal air temperature variations, and one is usually safe on this account with holes that are deeper than about 100 feet, as mentioned above. Perhaps the biggest problem with shallow gradient holes, and deep holes in certain geologic environments, is movement of ground water. In some areas of sufficient topographic relief and abnormal precipitation, aquifers tens to hundreds or thousands of feet deep may carry large quantities of non-thermal water which sweep away any anomalous amount of heat coming from depth and obliterate a high heat-flow or temperature pattern over a geothermal resource. It is imperative that one understands the hydrology of the prospecting area in order to predict and cope with the complications likely to be introduced into a heat flow or thermal gradient survey by the local geology and hydrology.

Shallow Temperature Surveys. Several workers have shown the utility in a few geothermal areas of the use of very shallow

holes for measurement of temperature. Such surveys at the Coso Hot Springs area in California show a +2 deg C anomaly over the reservoir in holes 2 m deep (LeShack and Lewis, 1983). Trexler et al used shallow temperature surveys to outline areas where thermal water flows close to the surface at Caliente, Nevada (Trexler, x). Careful corrections must be applied for slope of the land, surface soil or rock type and vegetation (which affect the way the surface reflects solar radiation), and for surface hydrology, topography and other factors. These corrections require care and understanding.

Infrared Surveys. Existence of shallow temperature anomalies implies that airborne or even satellite imagery in the thermal infrared region of the spectrum may be helpful. In practice, these methods have not been widely successful to date. Soil temperature fluctuations induced by sun angle variations, vegetation, ground slope and water table variations, to name a few variables, cause a high level of background noise against which one must try to resolve the rare geothermal anomaly. Of course in specific areas, depending upon the geologic situation, infrared airborne surveying may be helpful, but it would probably not constitute a first step in any exploration program.

Snow-Melt Surveys. In some areas, photographs have been made in an area after a snowfall, and the areas of first snow melt mapped. This can be a quick and inexpensive way of doing a reconnaissance thermal study of an area. x

Electrical Methods

Electrical geophysical surveys are used to measure the electrical properties of the earth and help determine the rock type, nature of pore fluids and the temperature in the subsurface. Most electrical geophysical methods are based on measurement of the electrical conductivity (or its reciprocal, the resistivity) of the earth. Conductivity and resistivity are measures of how well the earth conducts electrical current. In areas of high conductivity (low resistivity), the earth conducts electricity well, and in areas of low conductivity (high resistivity) the earth is a poor conductor. Geophysicists tend to use both terms in discussions of electrical geophysics.

With the exception of a few metallic minerals, dry rock-forming minerals do not conduct electricity well. However, electricity is conducted in the earth by chemical species dissolved within ground waters that occupy the pore spaces in the rock. As we have seen, ground waters and geothermal waters invariably contain dissolved chemical salts, and the ions in solution respond to an applied voltage difference between two points by moving through the water, thus sustaining a current. Measurement of the quantity of current induced by a given voltage drop constitutes a basic parameter of many, but not all, of the electrical geophysical methods.

Several parameters affect the value of resistivity of the rocks in the subsurface. Among them are:

1. The concentration of dissolved salts in the pore fluids in the rocks. We have noted that the current is carried by the movement of dissolved ions in the groundwater. The higher the concentration of these ions, the more current is carried and the lower the resistivity of the ground water;

2. The temperature of the subsurface. The higher the temperature, the faster the ions in solution are able to move. Thus, the higher the temperature, the lower the resistivity of the ground water;

3. The porosity of the rocks in the subsurface, i.e. the quantity of groundwater held in pore spaces and open fractures. By intuition, one would expect that the higher the porosity, the lower the resistivity of a rock because there would be more current-carrying ground water contained within it;

4. The degree of saturation of the rock in the subsurface. This factor is coupled with (4). If the rock is only partly saturated, there will be less fluid to carry current and the resistivity will be higher; and,

5. The mineralogy of the rocks. We mentioned above that most minerals do not conduct. However, a few metallic minerals to conduct electricity, and if their proportion is high enough, the rock will have low resistivity. More importantly, however, is a class of minerals whose mechanism of conduction is different from the metallic minerals. The clay and zeolite minerals generally have loosely-held ions in their crystal structure, and these ions can migrate under the influence of an applied voltage. Thus, rocks that contain clays or zeolites generally have low resistivity.

Factors (1) through (4) above are related through Archie's Law, given in Table 9 (Archie, x). The clay factor has not been successfully quantified and can not be included in Archie's Law at present.

On the basis of the foregoing and from what we already know about the temperature, salinity and hydrothermal alteration minerals within geothermal systems, one would expect geothermal resources to display good electrical conductivity. Indeed, low resistivity (high conductivity) has been discovered by surface surveys over many geothermal systems, and geophysical techniques that measure resistivity are in use worldwide in geothermal exploration (Ward and Sill, 1983).

There are many ways in which the resistivity of the subsurface can be measured using surface electrical geophysical surveys. Two basic divisions of the methods can be recognized -- the galvanic methods and the electromagnetic methods. In the galvanic methods, current is introduced into the ground through electrodes placed in shallow pits for surface surveys or placed in drill holes for logging. In the electromagnetic methods, either naturally occurring electromagnetic signals are used or current is induced to flow in the ground by creating an electromagnetic field with a coil of wire placed on the surface or in an airplane. We will consider these methods and also some variations upon them. The choice of which method to use in a given exploration problem is best left to an experienced geophysicist. Each of the methods varies in the type of information it will yield in certain geologic situations, in the difficulty of interpretation of the results, and in the speed and cost of the survey.

Galvanic Resistivity Surveys. In this technique, two grounded electrodes are used to introduce a current in the earth, and the voltage resulting from the current flow is measured between two separate grounded electrodes. There are several ways to deploy the electrodes as indicated in Figure 13. Perhaps the most useful configuration is the dipole-dipole array. Using this technique, an effective depth of exploration of approximately two times the electrode separation (a in Figure 13) can be achieved, and because the maximum practical value for separation is perhaps 1,500 to 2,000 feet, the dipole-dipole method can detect low-resistivity zones to depths of 3,000 to 4,000 feet.

Computer-aided interpretation methods are available and are easily applied. The geophysicist first performs a preliminary interpretation using rules of thumb for the particular electrode array being used. This first guess is entered as a model of the subsurface into the computer, and the expected resistivity expression of the model is calculated. The expected resistivity based on the model is compared with the observed resistivity, and the model is changed based on the differences. This process is repeated until a satisfactory fit is achieved between the calculated resistivity of the subsurface model and the observed values of resistivity.

The method can be very useful for obtaining detail on a geothermal system. Figure 14 shows the dipole-dipole resistivity amonaly over the Red Hill low-temperature geothermal system at Monroe, Utah. Note that there is low resistivity associated with the zone of known geothermal occurrence at the Red Hill hot spring. Low resistivity values persist both north and south of the hot spring. and are believed to show the total area prospective for drilling. With the exception of one hole drilled near the hot spring, to be discussed later as a case study, no further drilling along this resistivity anomaly has taken place.

Electromagnetic (EM) Methods. In the electromagnetic methods, the loop of wire used to create the electromagnetic field can be very large or small enough to fit on an airplane, and it can be placed vertically or horizontally. An alternating current is put into the loop, and the loop creates a magnetic field that alternated with the same frequency as the current. Part of this alternating magnetic field penetrated the earth and induces currents to flow in any conductor in the subsurface. The alternating currents flowing in the subsurface conductors create an alternating magnetic field of their own, and part of this secondary magnetic field cuts the surface of the earth, where sensitive receiving equipment is deployed. The receiver detects the magnitude and phase of the secondary magnetic field, which is due to current flow in conductors at depth, and this is the basic measurement technique in the electromagnetic method. Factors that can be varied include the amount of current in the transmitter loop, the frequency of the current and the spacing between the transmitter loop and the receiver.

In the electromagnetic methods, geophysicists tend to speak of the conductivity rather than the resistivity. Figure 15 shows a map of subsurface conductivity variation from the Newberry volcano area near Bend, Oregon.

Electromagnetic methods and galvanic resistivity methods each enjoy certain advantages. Resistivity methods usually use simpler equipment and the results are perhaps easier to interpret because a wider variety of computer aids is available than for the electromagnetic methods. On the other hand, electromagnetic methods do not require the long lengths of wire to be placed on the ground that galvanic methods do, which is an advantage in mountainous or heavily vegetated country.

Magnetotelluric (MT) Surveys. In this method, natural magnetic and electrical signals are used, i.e. it is an electromagnetic method using natural signals (Ward and Wannamaker, 1983). The natural signals originate in the ionosphere high above the earth. It can be shown that a measure of resistivity of the earth below is given by the ratio of the electric field to the perpendicular magnetic field on the surface. Now, an electromagnetic field, natural or man-made, will penetrate into the electrically conducting earth to a depth dependant on its frequency. One can define the "skin depth" as that depth at which the electromagnetic field is attenuated in strength by the factor 1/e from its value at the surface, where e is the base of the natural logarithms, and equals approximately 2.72. Thus, lower-frequency waves penetrate to deeper depths than do higher-frequency waves, and by making simultaneous measurements of Ex and Hy for a range of frequencies, a depth sounding may be effected, the lower frequencies yielding information from deeper depths.

The MT method has been used a great deal in geothermal exploration with generally disappointing results (Ward, 1983). By far the biggest problems appear to be misapplication and inadequate interpretation. Most MT data have been interpreted using one-dimensional inversion to a layered-earth resistivity structure. This method is totally inadequate in most geothermal exploration and usually produces misleading results. Full three-dimensional modeling is needed. The MT method has many subtleties, and must be applied with a great deal of care by geophysicists who are well experienced. It would generally not be applied in exploration for direct-heat resources, although MT data may be available for a prospect and should be reviewed by the geophysicist.

Audoimagnetotelluric (AMT) and Controlled-Source Audiomagnetotelluric (CSAMT) Surveys. The principle of this method is exactly the same as in the MT method discussed above. MT equipment can be considerably simplified if its range of operation is restricted to frequencies between about 1/10 Hz to 10,000 Hz, loosely called the audio range. This frequency range covers the depth range of usual interest in geothermal exploration. Therefore, the AMT or audiomagnetotelluric method, has seen some geothermal exploration. Most reported AMT surveys are scalar AMT, that is, only one component of electric field and one component of magnetic field are measured during the survey. It can be demonstrated that, in purely layered geologic terrains, this scheme is adequate for obtaining resistivity structure. However, if resistivity also varies in either or both of the horizontal directions, as it does in the vast majority of geothermal prospecting areas, scalar AMT is inadequate and is not recommended for exploration.

New equipment has recently become available to perform controlled-source audiomagnetotelluric surveys. In this method, a loop of wire on the ground is used to create an artificial source of electromagnetic waves instead of using the natural AMT and MT signals. Use of an artificial source gives the geophysicist much more control over the survey. CSAMT surveys in which all three components of the magnetic field (two horizontal and one vertical component) and both components of the electric field (there is no vertical component across the earth-air interface) are measured are called tensor CSAMT surveys, and they yield very much more information than the scalar AMT surveys discussed above. Although tensor CSAMT surveys are not thoroughly tested for geothermal application, indications are that they will yield good results and may displace the use of the galvanic resistivity method in certain applications.

Self-Potential (SP) Methods. Self-potential or spontaneous-potential surveys are one of the electrical geophysical survey types wherein the resistivity or conductivity of the earth is not measured. Instead, natural voltages are measured over the surface of the earth, and these natural voltages are related to chemical or physical processes in the subsurface. There is a process called the "electrokinetic effect" whereby water flowing in the subsurface can generate a voltage. Another process called the "thermoelectric effect" expresses the coupling between temperature variations in the subsurface and development of a voltage. SP anomalies over convective hydrothermal systems, then, arise from the electrokinetic and thermoelectric effects, which couple the generation of natural voltages with the flow of fluids and the flow of heat, respectively (Corwin and Hoover, 1979; Sill, 1983).

SP surveys are simple and quick to run, and are inexpensive. However, SP surveys are not diagnostic in detecting geothermal resources and are difficult to interpret in terms of a subsurface model that assists in geothermal exploration. SP surveys have been used successfully in certain areas. On Hawaii, Zablocki (1976) found a large SP effect over the East Rift zone, which is currently producing geothermal electricity and is an area of continued exploration. An SP anomaly has been found over the East Mesa geothermal field, and it is shown in Figure 16. On the basis of the physical mechanisms by which SP effects arise, one would expect that SP anomalies should be found over zones of upwelling of geothermal waters and perhaps over zones of recharge to geothermal systems as well.

Seismic Methods

Elastic mechanical waves are transmitted through rocks and their measurement can be used to help determine the structure and mechanical properties of rock bodies. Two types of waves are most useful:

1. The compressional or primary (P) wave, in which the particle motion in the rock is back and forth along the direction of travel of the wave. Of course, each rock particle only moves a millimeter or so, but the rock particle is in contact with surrounding rock particles and transmits its motion to its neighbors. In this way, the wave moves outward in all directions from a source. P-waves are ordinary sound waves in rocks. They travel at velocities that vary between about 3,000 ft/sec and 20,000 ft/sec; and,

2. The shear or secondary (S) wave, in which the particle motion is perpendicular to the direction of travel of the wave. S-waves have no analog in air as P-waves do because fluids (liquids and gases) do not support shear. In rock, S-waves travel at velocities

about 70 percent of the velocities of P-waves.

There are a variety of seismic methods, and they can be applied to different exploration and mapping problems. The selection of which seismic method might be helpful is best left to an experienced geophysicist. Seismic techniques can be classified into active techniques and passive techniques. In the active seismic techniques, dynamite detonated in a shallow borehole or some other source of mechanical energy is used as a source for the signals that are received and interpreted. The active techniques can be further subdivided into reflection and refraction surveys. The passive techniques use naturally occurring signals. Natural signals arise from earthquakes and from the movement of water or molten magma in the subsurface. We will consider some of the common seismic methods below.

Seismic Reflection Surveys. In this method, an artificial source is used to create seismic waves which travel downward in the earth, are reflected from a boundary at which the mechanical properties of the rock change (a velocity discontinuity) and return to the surface. At the surface, their arrival is detected by sensitive geophones spaced along lines at known distances apart. This method has proven to be very effective in exploring for petroleum, and is used extensively by the large oil companies. However, it has not been as successful in the geothermal environment. Geothermal areas seldom have the flat geological structures or layering that the method detects best. Swift (x) discussed the results of a trial reflection survey at Beowawe, Nevada.

Seismic Refraction Surveys. In refraction surveys, waves refracted along subsurface boundaries and then returned to the surface are detected. Refraction surveys can be helpful in determining the geologic structure in the shallow subsurface, and are perhaps more easily applicable to direct heat prospecting than are reflection surveys. The thickness of unconsolidated alluvium over bedrock can usually be mapped with refraction. Such information is useful in planning drilling and other exploration activities. Applegate et al (1981) discuss applications of both reflection and refraction to geothermal work.

Earth Noise Surveys. There is limited evidence that hydrothermal processes, including boiling and the rapid movement of water in geothermal resources, can generate seismic waves in the frequency band 1 to 10 Hz (Liaw and Suyenaga, 1982). Noise also arises in such sources as traffic, trains, rivers, canals, and wind. Liaw and McEvilly (1978) have demonstrated that field and interpretive techniques for earth noise surveys require a great deal of understanding and care. These surveys can provide a guide to hydrothermal processes provided the data quality is good and careful interpretation is done.

Microearthquake Surveys. Microearthquakes frequently are related spatially to major geothermal systems. These microearthquakes appear to originate in faulting and fracturing at depth, processes that are needed to keep the plumbing system of hydrothermal resources open. Accurate locations of these earthquakes can provide data on the locations of active faults that may channel hot water toward the surface. Microseismic activity in most geothermal areas has been observed to be episodic rather than continuous. This characteristic limits the technique in its geothermal exploration applications because the surveys are not inexpensive and one may need to survey for months or years to get useful information.

Figure 17 shows the locations of microearthquakes at the Roosevelt Hot Springs geothermal area in Utah. The figure shows the locations in both plan map and vertical section. The microearthquakes are believed to be associated with movement on the Negro Mag fault, one of the main zones of upflow of geothermal fluids in the district. Significantly, a six-station seismic net was run at Roosevelt for more than two years and it detected only a few events. During the last two months of the survey, more than 1,000 microearthquakes were detected to provide the data in Figure 17. This is a demonstration of the episodic nature of seismic activity in this area.

P-Wave Delay and S-Wave Shadowing. Seismic methods have been proposed for use in detecting molten magma in the subsurface to depths of 10 miles or more. Magma has a lower P-wave seismic velocity than consolidated rock and, being a liquid, it will not pass shear waves. Thus, if one observes seismic waves from distant earthquakes that have passed through a magma body, the P-waves should be slowed down and the S-waves should be removed. Surveys that compare these parameters with nearby seismic waves that have not passed through the magma have been attempted with apparent success at Newberry volcano in Oregon, for example (x). The success is apparent, of course, because there is no proof through drilling that the magma predicted to occur there on the basis of the survey actually exists. Such work has indirect bearing on exploration for direct heat resources in that surveys of this type would normally not be carried out specifically for direct-heat exploration, but such data might have been collected in the prospect area for other purposes, and it should be reviewed and made part of the direct-heat exploration picture.

Magnetic Methods

The earth has a main magnetic field whose shape is similar to that which would be produced by a large bar magnet near the center earth. The earth's magnetic field is believed to arise from electrical currents flowing deep within the earth, in the electrically conducting, fluid core. This magnetic field induces a magnetic response in certain minerals at and near the earth's surface. Principal among the magnetic minerals is magnetite (iron oxide) and pyrrhotite (iron sulfide). Although pyrrhotite is not common, magnetite is a mineral that is found in small amounts in many rocks of the earth's crust. The magnetism in rocks adds to the earth's main magnetic field, and by detecting spacial variations in the earth's total field, the variations in distribution of magnetic minerals may be deduced. This information, in turn, can be related to geology.

The earth's magnetic field can be mapped on the ground by use of a sensitive instrument known as a magnetometer. Magnetometers can also be installed in aircraft and magnetic maps created much faster and more cheaply from the air. Aeromagnetic surveys are widely used by industry in petroleum and mineral exploration in attempting to map subsurface geologic structure and changes in rock type. The use in geothermal exploration closely follows that in mineral exploration, for most geothermal resources are located in geologic environments that are similar to or the same as those in which mineral deposits are found.

The physical property of the rock that quantifies its response to the earth's magnetic field is called the magnetic susceptibility. Susceptibility varies over several orders of magnitude, but most rocks have magnetic susceptibilities in the ranges given in Table 10. Note that igneous and volcanic rocks are usually highly magnetic (high susceptibility). The process of hydrothermal alteration, discussed in the section on geochemistry, tends to destroy the magnetite in a rock, and to render it nonmagnetic or only weakly magnetic. Thus, some geothermal systems in magnetic rocks are expressed as magnetic lows. However, there are many other causes of magnetic lows, so one must be very careful in interpreting magnetic data.

Regional aeromagnetic data are often available as part of state or federally sponsored surveys. These data often show major structural features and aid in geologic mapping in areas where the surface is covered by alluvium. Regional data are generally too widely spaced and too high in altitude however, to constitute a data base appropriate for detailed interpretation on the scale of a geothermal prospect. In certain geologic environments, therefore, the geophysicist may want to collect detailed airborne or ground magnetic data on a geothermal prospect. The locations of geologic structures (faults, fracture zones), intrusions, volcanic rocks, and other features of interest in forming a geologic model of a geothermal prospect may be evident of magnetic maps. Figure 18 is an aeromagnetic map of the Cove Fort geothermal area in Utah. Explain x.

Gravity Methods

The earth's gravity field is caused by the mass of the earth itself. Since the density of a material is determined by the mass per unit volume, variations in the density of subsurface rocks cause minute variations in the earth's gravity field. In order to detect these gravity variations, very delicate instruments are required. The modern gravity meter measures 1 part in 1,000,000,000 of the earth's gravity field. These instruments are among the most sensitive mechanical instruments ever made by man.

Gravity data are often acquired or compiled in the early stages of an exploration program. Regional data, with station densities of 1 station per square mile, may be available as the result of surveys by governments or universities. Available data are generally the starting point for detailed surveys suitable for geothermal prospecting.

The contribution from gravity surveying to geothermal exploration is much the same as from aeromagnetics, that is, structural, lithologic and other geologic information. However, three notable successes of the gravity method stand out. In the Imperial Valley, California, gravity surveys have proved useful in locating areas where hydrothermal alteration and metamorphism have caused the rocks to become densified (Rex et al., 1971). Deposition of minerals in the pore spaces in rocks above convecting hydrothermal systems has increased the density of the rocks enough to be detectable with the gravity meter. Gravity surveys have been used in conjunction with temperature gradient and heat-flow surveys to locate subsurface geothermal systems that have absolutely no surface manifestation.

A second application of gravity surveying has particular importance to exploration for low- and moderate-temperature geothermal resources. Such resources are often found on the active faults that bound many of the mountain ranges in the western United States. Range-front faults are particularly common in the Basin and Range province. Gravity surveys can generally be used to map the locations of these range-front faults. The faults have thrust the mountain blocks up while dropping the valley blocks down. The valleys become filled with unconsolidated erosional debris from the mountains. There is a marked contrast between the density of rocks on the mountain side of the fault and on the valley side.

TECHNIQUES FOR DIRECT HEAT EXPLORATION

The first part of this chapter discusses essentially all of the geological, geochemical and geophysical methods commonly applied in geothermal exploration. Not all of the techniques presented would be used in a typical exploration program for direct heat resources, however, since some of them are normally applied only in exploration for high-temperature resources. The reason a more complete discussion was presented above is that some low- and moderate-temperature resources occur in the fringe areas of high-temperature resources. Data and results of the application of any of the methods above may be available to the direct heat developer, and if available they would be a valuable asset. In this section we will present a little more detail on these methods commonly applied in direct-heat exploration, including costs and expected rates of progress.

Geothermal development is an interdisciplinary endeavor. Figure 19 shows some of the components of the team that must work together successfully if a site is to be developed. Because geothermal resources are geological phenomena, earth science information is needed for all phases of the development. This involvement of the earth sciences is similar to that required for development of petroleum and mineral reserves.

Over the years, the petroleum and minerals industries have spent billions of dollars developing earth science tools and techniques to solve their particular exploration problems. By contrast, relatively little has been spent in developing earth science tools and techniques especially for geothermal exploration. Geothermal developers have resorted to use of existing tools, which are not optimum for geothermal application. In some cases, there are simply no tools or techniques to solve a particular problem.

Risks in Geothermal Exploration

Like all natural resource exploration, geothermal exploration is risky. There is never a guarantee that a test hole or a production well will intersect the geothermal fluids desired. The developer must be mentally and financially prepared to accept exploration as a risky venture. The degree of risk can often be assessed semi-quantitatively by analysis at the outset, and it will usually be found to be dependent upon two factors:

1. The risk that the resource exists; and,

2. The risk that the exploration program will locate or tap the resource.

Both of these risk factors can be affected by human intervention. Although Mother Nature has placed geothermal resources where they are, the human factor enters by the developer either being astute enough or not to conduct the exploration program in the right place and in the right way. This applies both on the reconnaissance level, when one is trying to locate a geothermal system, and in the case where one is exploring for production within a geothermal system. The second risk factor, the risk of finding the resource, is also controlled by several other factors, including the amount of money available for exploration and the quality of the exploration team. If the project will not support much exploration, the tendency is to move directly to the drilling of a production well. In some projects, this may be justified and of relatively low risk. But in the majority of cases, moving directly to production drilling without first developing solid evidence that the resource is to be found underneath the drill site increases the exploration risk needlessly.

There has been an unfortunate history in the exploration for direct heat resources for the developer to do little or no exploration, but simply to drill at the spot where the thermal water is wanted. Weak geologic evidence may be available to support the drilling. In most cases, such an approach is doomed to failure. Geothermal resources, we reiterate, are where Mother Nature put them, not necessarily where we want them to be. The developer is cautioned to sure that there is solid earth science evidence for the occurrence of a resource in the area and that enough exploration data have been collected to point definitely to the drill site finally chosen as the best place to drill. There is only one way to ensure that the quality and quantity of exploration are sufficient to select drill sites, and this is by forming an experienced and educated exploration team and listening to their advice. Even with the best exploration team, there is no guarantee that the project will succeed, but one will be sure that the risk is minimized.

Costs of Exploration

The following discussion of the techniques most commonly applied to exploration for and within direct heat resource areas is accompanied by tables that give an indication of the costs and rates of progress to be expected for the various techniques. These costs and rates of progress are only approximate, and should be used as guidelines only. Because each resource area differs from all others, it is not possible to specify exact figures. It should also be noted that the tables show costs only for the field data gathering portion of exploration. As a general rule of thumb, an amount of money and time equal to those for field data gathering should also be allowed for data compilation and interpretation.

Geological Techniques

Geological mapping and drill chip or core logging are usually required on any project. Money paid to a good field geologist is well spent. Not every geologist is good in field work -- some are laboratory specialists. As with other personnel, it is wise to get recommendations on a geologist for the specific job at hand. The rate of progress for either mapping or logging depends mostly on the complexity of the geologic picture. A minimum time must always be spent on familiarization. Progress is more rapid once the geologist is familiar with the rock types and structure in the area.

For age dating, the sample collecting must be carefully done. In geothermal exploration, the aim usually is to locate, sample and date the volcanic rocks in the area that are less than 1 million years old. The geologist must be thoroughly familiar with the area in order to ensure that the proper rocks are sampled. Analytical costs are so high that it is usually not satisfactory to attempt to date all of the igneous rocks. Field relations among rock types can be used for relative age dating before samples for laboratory work are selected. The analytical work and interpretation will take some months. It is rare to get quicker turnaround from a dating laboratory.

Geochemical Techniques

The geochemical methods most often used in exploration for direct heat resources include sampling and analysis of waters from springs and wells and soil geochemical surveys. Special care is needed in sampling wells and springs, and special equipment is needed for sampling those wells that produce water above the boiling temperature. The geochemist should be well experienced in these sampling techniques if the data are to be relied upon.

Soil geochemical surveys can be used to help delimit the prospecting area and to help locate zones of upflow and faults. Mercury and radon are probably the most commonly used, but other elements such as arsenic or manganese are also used. The geochemist will the best judge of which element(s) to use in a specific case. In the case of mercury, the analytical equipment, an instrument known as a gold-film mercury detector, can be used in the field, in a trailer or a motel room. This allows one to analyze the samples in real time and to modify the survey as one goes along based on the results. Alternately the samples can be sent to a laboratory. Laboratory analysis is probably less expensive for small surveys (< 100 samples) but for larger surveys it may be judged best to take the analytical gear to the field.

Geophysical Techniques

The geophysical methods most commonly used in exploration for direct-heat resources include thermal methods, electrical methods, magnetic methods and gravity methods. Drilling of 300 to 500 feet holes for measuring temperature gradients and heat flow can be used to help locate zones of upflow or in order to gain confidence in a site prior to drilling a production-sized hole. Of course, the thermal gradient hole is usually smaller in diameter and much cheaper than production sized hole. Still gradient/heat flow drilling is expensive. The services of an experienced geothermal driller should be obtained, and the developer should check his credentials specifically in geothermal drilling.

Shallow-temperature surveys can be a less expensive way to map subsurface temperatures but are difficult to use due to the nature of the corrections that must be applied to yield meaningful data. Only a specialist should use them.

Of the several electrical arrays available for resistivity surveys, the dipole-dipole array is recommended. Dipole-dipole data can be reduced and plotted in the field, and the survey can be modified immediately if needed. Several of the EM techniques are applicable to direct heat exploration, and the geophysicist should make the selection. Both resistivity and EM are used to find areas of high subsurface conductivity that one hopes will correspond with high subsurface fluid temperatures. The SP method can be used to help locate zones of subsurface fluid movement. These data are often difficult to interpret, however. All of the electrical methods are subject to interference from man-made, so-called "cultural" features, which include electrically grounded objects such as railroads, fences, powerlines, pipelines and radio or TV stations.

Airborne magnetics would only be used in a large reconnaissance exploration program. Ground magnetic measurements may be of use in helping to pinpoint the location of faults. Ground surveys must be carefully done to eliminate geologic noise from the data caused by the nearness of the magnetometer to magnetic boulders in the shallow subsurface and by nearness to cultural features. Airborne surveying helps eliminate these noise sources by getting the magnetometer a few hundred feet off the ground.

Gravity surveys can be useful in locating faults. There must be a decision about whether to provide the elevation control needed for data connection using altimeters or by land surveys such as leveling. Altimeters are faster and less expensive, but elevations are only accurate to 2 to 10 feet, which may not be good enough for the precise work needed in some applications.

EXPLORATION AND RESERVOIR ASSESSMENT STRATEGIES

Successful developers of geothermal energy usually have a strategy or plan of attack before they begin. Exploration strategies should be developed to help guide exploration programs. Strategies have the purpose of minimizing risk of failure and of optimizing the cost-effectiveness of the exploration. One common feature of these strategies is that they provide one or more decision points where one can either elect to terminate the program or go on to the next stage. The less expensive techniques are usually used in the first stages of exploration, when the risk is highest, and each subsequent stage is usually more expensive than the previous stage. Each stage should reduce the risk of failure on the project. Exploration, of course, eventually leads to the drilling of a production- or injection-scale well, usually the most expensive component in subsurface development of geothermal resources. An optimum exploration strategy will depend on the size and purpose of the project, the amount of money and time available, the geologic environment, and the cumulative exploration experience in that environment, among other factors. In the remainder of this section, we will examine exploration strategies in more detail.

Limitations of Exploration Strategies

It is important to understand that because geothermal resources are so varied in detail, even within resources of the same general type, it is not possible to specify a certain sequence of exploration techniques that will be the most cost effective or that will even work in all circumstances. Stated differently, there is no exploration strategy that can be blindly applied with the expectation of success every time. The exploration strategy to be followed in any area must be designed specifically for application to that area by the geoscientists who are performing the work and interpreting the data. Nevertheless, we can present the components of exploration strategies in a generic way.

Generic Exploration Strategy

Figure 20 illustrates a generic exploration strategy. Before such a strategy can become useful on a specific project, specific detail must be added to each of the steps. Several aspects of Figure 20 merit discussion. First, exploration progresses from the consideration of large areas, perhaps 10,000 sq mi during the reconnaissance stage, to the development of a prioritized list of prospects within the reconnaissance area, and then to testing of each high-ranking prospect by detailed exploration and drilling. That is, exploration proceeds from the consideration of a large area, through elimination of most of this large area as being of little or no interest and on to detailed studies of a few small areas. During this process, it is prudent to use lower unit-cost exploration techniques during the earlier sages of the program and reserve higher-cost techniques for use later when the area of interest has been reduced. There are decision points at the end of each stage, when one may elect to terminate the project. By assessing odds for success at each decision point and comparing the project to others or other uses of the money and manpower, an optimum exploration program will result and the risks and costs of exploration will be minimized.

We assume, as reconnaissance exploration progresses in a region, that several favorable individual prospects will be identified. The relative priorities among these prospects for further exploration must always be considered. In what follows, we discuss exploration strategies as applied mainly to a single project, but we must always bear in mind that various prospects may be in various stages of exploration, and that we must always prioritize work among the prospects so that money and human resources are deployed in the optimum way.

Available Data Base (1). All available regional and local geological, geochemical, geophysical and hydrological data should be assembled for the prospective exploration area and its surroundings. Once assembled, specialists in each of the earth science disciplines should assess the data in a preliminary fashion to determine its quality and to identify any obvious gaps (2). Often basic geologic data will be missing. It should be obtained at this point by geologic mapping. It is very important to have a sound geologic data base at the outset of an exploration project because interpretation of all of the other data sets will depend upon it and must be in agreement with it.

Integrated Interpretation (3). When the data base is judged to be sufficient, it should be interpreted by specialists. By "integrated interpretation" we mean to convey the necessity for the various specialists to work closely together in the data interpretation process. The objective of this integrated interpretation is to formulate a conceptual geologic model of the subsurface (4) in the exploration area that agrees with all of the available data. Of course, the model should concentrate on those features that are pertinent to the potential for occurrence of a geothermal resource in the area.

In order to perform this interpretation step, a number of ingredients must be available (5). These include (a) knowledge of geologic models of geothermal resources in other areas as a basis for conceptualization about the study area; (b) data interpretation aids such as computer modeling programs and type curves for geophysical and geochemical data; and, (c) experience in geothermal exploration for the general type of resources being sought.

Conceptual Model (4). Once a model has been formulated, it is used to answer a number of questions. The first question is "does the model reveal anything to indicate that a resource may not be present", i.e. is there negative information? (6) If so, its quality and impact must be assessed, and one may decide at that point not to pursue exploration in the area any further. If the decision is made to proceed, then the model is useful in formulating the subsequent exploration.

Exploration Techniques and Survey Design (7). There are several important aspects to selection of exploration techniques. If geophysical surveys are being considered, there must be some reason to believe that the geothermal system or some its features will cause a change in one or more of the basic physical or chemical properties that geophysical surveys measure, e.g. density, magnetic susceptibility, electrical resistivity, sonic velocity. Such assurance results by deductive reasoning from the preliminary conceptual model of the system. The model encompasses what is known about the exploration area and a best estimate of the configuration of the subsurface. One then asks the question, "if a geothermal system exists in this area, what effect will it most likely have on physical properties of the area that can be measured by geophysical surveys?" Once expected physical-property changes have been identified, then an estimate should be made of the geometry of the region over which the physical property is believed to vary. One might postulate, for example, that if a geothermal system large enough to be of interest for development exists in a given area, then it should cause the electrical resistivity to be reduced by a factor of 10 over a volume 1,000 ft by 5,000 ft by 2,000 ft thick buried 500 ft to the top. The geologic model of the area helps place an expected size on the anomalous area while consideration of the effect of geothermal fluids on physical properties allows an estimate of its effect on resistivity.

Given that a resistivity low is expected, one then uses forward computer modeling programs or type curves (8) to help decide (a) whether or not the anomalous body should be detectable by a surface resistivity survey, (b) what electrode array to use for the survey, (c) what electrode spacing to use for the survey, (d) what configurations of survey lines would be optimum, and other pertinent questions. Notice that the same modeling aids that are used in interpreting the final survey data are used at this stage also to do predictive modeling during the survey selection process. This helps to ensure that the survey will indeed measure a detectable response from a geothermal system if it exists. Also, if no such resistivity response is detected, then the model of the subsurface must be changed accordingly.

Our resistivity example is given simply to illustrate the thought processes. Similar considerations would be used if a geochemical or any other survey were proposed.

Integrated Interpretation (9). After the survey(s) have been successfully completed, there again needs to an integrated interpretation of the entire data base, with emphasis on incorporation of the newly acquired data. In order to perform this integrated interpretation, the explorationist must have access to interpretation aids such as computer programs, type curves, etc. (10).

Updated Model (11). The result of the integrated interpretation will be an updated, upgraded geologic model of the subsurface. The model should represent the actual subsurface with a greater degree of accuracy because of the survey(s).

With an updated model, one is in a position to determine the next step is. Were the survey results negative? Does this establish with reasonable certainty that no resource exists? If so, the prospect should be abandoned. Is there another survey that should be run? Or perhaps the survey results were positive, were reasonably quantitative and encouraging. In this case one many want to drill test the area.

Drilling (13). Drilling could be in shallow (< 500 ft) holes to measure thermal gradient and heat flow, or one could decide to drill to intercept the target. Drill hole parameters, including diameter, casing plan and the need for blow-out prevention equipment must be carefully considered.

Collect Subsurface Data (14). Because drilling is expensive, the best possible use must be made of drill data and results. Drill cuttings should be collected from rotary holes. These are used to help define lithology, petrography and hydrothermal alteration and for measurement of physical properties. Conventional geophysical well logs may be run in the hole, with a typical logging suite probably being temperature, caliper, resistivity, gamma ray and acoustic logs. If the well is flowed or if there is a drill- stem formation test, samples of the fluids from the well must be carefully collected and preserved for analysis. Often a hydrothermal component of such fluid samples can be detected through chemical analyses, lending encouragement for further exploration. Chemical geothermometer calculations can be made from the analyses of pristine samples to help determine potential resource temperatures.

Integrated Interpretation (15). Again the new data are interpreted in light of existing data and existing models of the resource area, and the conceptual geologic model is upgraded (16). The question of what to do next is then answered in light of the resulting model (17). One may elect to perform further surface exploration (7), drill a second test well (13), drill a production well (18) or abandon the project (19).

Hypothetical Examples of Exploration

We might best illustrate the application of the above exploration strategy by discussing several hypothetical exploration programs for various projects. We will follow up in the next section by considering actual case studies.

Small Project. We assume in this illustration a small-scale user interested in drilling one well into a fault-controlled resource near a known hot spring. The main questions to be answered in this case are the exact location and drilling depth of the well.

The first step is to acquire all available geologic data in the area. These data should be posted on a map at a scale of 1:24,000 (1 in = 1/2 mile) or more detailed. If the intended use of the geothermal fluid is small in scale, the production well would need to be drilled near the use site in order to minimize piping expense. The main problem in siting the well becomes one of mapping in detail the location and dip angle of the fault in the near vicinity of the surface use site. A geologist should perform this job. Although the location of the fault might be quite well known, the angle of dip of the fault may not be easily established. Determination of dip angle is critical because presumably one would want to drill in such a way as to intersect the fault at a depth of, say, a few hundred feet. The angle and direction of dip must be known in order to determine how far away from the surface trace of the fault and on which side of the surface trace to drill in order to hit the fault at a specified depth in the well. The hope would be to produce water of higher temperature than the temperature of the hot spring, which may be diluted by cold surface water.

The exploration program might be outlined as follows:

1. Specify the temperature and flow rate needed from a production well in order for the intended application to be economically successful.

2. Take a water sample from the hot spring for chemical analysis. Interpret the geothermometer calculations from the chemical analysis to indicate the maximum expected water temperature.

3. If the maximum expected temperature is lower than the minimum temperature needed for economic utilization, decide whether or not to proceed. The geothermometer data may be incorrect due to breakdown of one of the assumptions made in the geothermometer calculations (see Chemical Geothermometers above).

4. If the decision is to proceed, have a geologist produce a detailed map of the surface trace of the fault and determine its angle and direction of dip.

5. Based on (4), select the spot on the ground where the well will be drilled, and specify a drilling depth. Adequate margin for error in predicted drilling depth should be allowed.

6. Select a drilling contractor and drill the well. During drilling, the geologist should log the drill chips in real time as the hole deepens in order to look for indications of geothermal activity at depth and to recognize the fault when it is intersected. One would hope, of course, that intercept of the fault would be accompanied by a large flow of thermal water into the well. If the fault is sealed where it is intersected, however, one needs to recognize this in order to avoid wasting money by drilling far beyond the fault.

7. If the first well is successful, one could proceed to well testing to determine production characteristics. If the well is not successful, there would be a decision about whether to select a site for a second test well or to abandon the project.

Mid-Sized Project. Suppose a geologic situation similar to the small project discussed above, but in this case there is a requirement for several production wells. This can add considerably to the complexity of exploration because the production wells must be placed far enough apart to avoid interference during production.

In this case, one would probably again begin by seeking encouragement from chemical geothermometry. The next problem is to map the fault and the upwelling thermal water over a much larger area, away from the known hot spring. Detailed geologic mapping over and surrounding this larger area would be undertaken to trace the fault, which, of course, can not be assumed to be on a straight-line projection from the known hot spring area. A geophysical resistivity survey may be selected after the detailed geologic mapping in order to determine the total extent of the area of upwelling thermal water. This area should correspond to a region of low measured resistivity.

Once the resistivity data are obtained and interpreted, they would outline the prospective geothermal area and give some indication of the depth beneath the surface at which thermal water may be found. The area within which wells may be successfully drilled would now be known, but the resistivity may or may not provide enough information to select actual drill sites. If the trace of the fault is not apparent from geologic mapping, as it may not be because of alluvium covering it, the explorationist may elect to do a soil mercury survey or a soil-gas radon survey to help pin-point the fault within the resistivity low.

Data are collected until the explorationist feels comfortable in being able to spot well locations that have a suitably high chance of intersecting the resource.

From here, the exploration proceeds as before. The geologist makes a geologic log of each well from the drill chips and the work proceeds until the specified amount of geothermal water is found or until the decision is made to stop.

Reconnaissance Exploration. In this program, we assume that the developer is looking for a resource suitable for powering a specific type of plant, say a vegetable-drying plant. The plant can be built near a suitable resource. In this case, the developer has specific criteria for what constitutes a suitable resource in terms of minimum production temperature and flow rate, but the resource could be located anywhere near a highway or railhead within a multi-state area.

Designing and executing an exploration program for such a resource presents different problems from those illustrated in previous examples. The first step is to compile selected available data on known geothermal occurrences in the multi-state area. Maps and reports on known thermal features are available at the offices of many state geological survey agencies or at universities as noted below in Table x. The data compilation should include the name and location of each thermal spring in the area along with its measured temperature, flow rate and any available chemical data which could be used to calculate geothermometer temperatures. If geothermometer calculations have been made by others, the results of the calculations, the water chemistry data and the qualifying assumptions should be noted. If there are no chemical data for some thermal springs, these springs should be added to a list of candidate springs for sampling. Also noted in the first data compilation are geothermal features other than springs. These include known fumaroles and deposits of siliceous sinter (SiO_2) or travertine ($CaCO_3$) known to have come from formerly active springs.

The total list of geothermal features at this point would include all known thermal springs and other geothermal features. The task is to cull this list into geothermal features of further interest by discarding features of no further interest. Several criteria can be applied. Springs showing actual measured temperatures in the range of interest and areas of active fumaroles would top the list of features of interest. Also on the list would be springs whose potential temperatures, as indicated by chemical geothermometers, are above the specified minimum temperature as well as areas of sinter deposits. Sinter is believed to precipitate in quantity only from waters whose subsurface temperature is 180 deg C (360 deg F) or more, and thus sinters are a positive indication of high subsurface temperatures, at least in the past. Travertine, on the other hand, can deposit from springs of essentially any temperature, even non-thermal springs. Also listed as lower priority at this time are thermal springs for which there are no chemical geothermometer data but which lie close enough to a highway or railhead that an economic installation could be made given reasonable assumptions of geothermal production temperature and flow rate.

Note in all of these considerations so far that the chemistry of the produced fluids is primarily important in terms of calculating geothermometers and classifying geothermal waters. We assume that variation in chemistry of the produced fluids would pose no environmental, scaling or corrosion problems that could not be handled by proper engineering when the resource is discovered and tapped.

Continuing our example, the next step is to visit each site for which there are no chemical geothermometer data and obtain a suitable sample for analysis. The uninitiated would go to the spring, wade in with both feet, fill a beer bottle with water, plug it with a piece of whittled sage brush and send it to the cheapest lab around for analysis. But the enlightened reader of this handbook would obtain the services of a competent geothermal explorationist, who would obtain suitable bottles for unpreserved and acid-preserved samples, carefully collect the samples while measuring the spring temperature, pH and conductivity, and send the samples to a laboratory whose reputation for analysis of geothermal samples is proven. Which sampling style would you bet a $250,000 production well on?

When all gaps in geothermometer data are filled, a list of candidate sites for further exploration is generated. Highest on the list will be sites whose known, measured temperature exceed the temperature requirements for the application. Perhaps the developer has already examined the land situation at these sites and determined that land is not available. Next will be sites whose chemical geothermometer temperatures are in the range of interest and which have high flow rates. These sites will be followed by sites whose indicated temperatures are high but the flow rates are lower. The list will probably be too long for more detailed exploration to be conducted at all sites. Therefore, the next step is to cull the list of candidate sites further and to determine which ones will be explored further. At this stage, the decision may be taken to do resistivity geophysical surveying over the five top-ranked areas and to rank these five areas on the basis of the areal size of the resistivity anomalies under the assumption that the larger in size the resistivity anomaly is, the larger the capacity of the reservoir for delivery of thermal water to the plant. After the surveys are done and the areas are ranked, drill testing would be the next step, beginning with the area ranked highest in priority. There may be need at this point to obtain additional data in some of the areas to assist in siting test wells. Mercury geochemistry may be a choice. On the other and, if the anticipated production interval is fairly deep, say 2,000 feet or more, one may want to drill several shallow (500 ft), small-diameter holes for measurement of temperature gradient and heat flow. The area of highest indicated heat flow would then be selected for a deep production test. Such an approach may be effective in reducing the risk of failure of the expensive production well in producing adequate thermal waters.