GL00843

THE APPLICATION OF GEOPHYSICAL METHODS TO GEOTHERMAL EXPLORATION

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Prepared for

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Corazon de Maria, 2
28002-Madrid

January, 1985

CONTENTS

I. EXECUTIVE SUMMARY

An extensive literature search indicates that a wide variety of thermal, electrical, potential field and seismic geophysical methods are being employed in geothermal exploration. The published literature provides a fair indication of overall method applicability, although cost-effective exploration programs show a preference for a small number of well designed exploration surveys. In order for geophysical surveys to be successful, the type of survey and details of survey design must be consistent with the known geology and resource type.

Gravity, magnetic and electrical resistivity (VES) surveys are often employed in preliminary or reconnaissance studies. Thermal gradient and/or heat flow studies are employed in both the preliminary and the detailed reservoir evaluation stages of exploration. Electrical resistivity, magnetotelluric, and self-potential methods are commonly used in detailed, prospect-scale exploration programs. A less frequent use of the seismic methods, both active and passive, may reflect the relatively high cost of these surveys and the limited data base for some geologic environments (i.e. basins) where the methods could be expected to be effective.

The most important physical property associated with geothermal systems, apart from temperature, appears to be low electrical resistivity which results from the higher temperature of the fluids, the general higher dissolved ion content, and conductivity enhancement associated with wall rock alteration. The low electrical resistivity associated with many geothermal systems provides a favorable target for surface electrical methods.

Many of the geothermal resources already identified by the IGME are located in complex geologic environments with a variety of rock types. Most of the identified resources are associated with major faults or fracture systems and relatively few of these resources would appear to have a potential for high enthalpy and electric power development. Reservoir volume and/or permeability may limit the geophysical detectability and ultimate reservoir potential of several occurrences.

A generalized exploration strategy is presented for three major resource types: volcanic; igneous; and basins. The complexity of the local geology requires careful integration of geologic mapping, geochemical studies and selected geophysical surveys to arrive at the most cost-effective exploration strategy for a particular prospect area. It is imperative to note that exploration strategies that minimize cost while at the same time maximizing chances for success cannot be remotely designed, nor can a given strategy be applied blindly to many areas. Each exploration area is a separate case, and exploration techniques that work in one area may not work in another. An effective exploration strategy is best designed by the geologists, geochemists, geophysicists and hydrologists who are actively working in the exploration area. These people are in the best position to assess the potential contribution and the probable costs of applying any specific technique and of weighing the relative merits of the broad range of techniques available.

APPENDIX I - PHYSICAL PROPERTIES OF GEOTHERMAL SYSTEMS
(Selections from the literature)

A. Moskowitz and Norton, 1977
B. Ward and Sill, 1984
C. Sill, unpublished manuscript

ILLUSTRATIONS

ILLUSTRATIONS (cont.)

TABLES

II. INTRODUCTION

The Earth Science Laboratory, University of Utah Research Institute (ESL/UURI) and Law Engineering, Iberica, S.A. have been selected to undertake an evaluation of the application of geophysical methods in geothermal exploration. Important aspects of the study include:

- A bibliographic search to establish and document the use and relative cost effectiveness of a variety of geophysical methods on a worldwide basis.
- A statistical tabulation which presents the relative frequency use of various methods, available information on costs, and applicability for various geologic regions.
- An evaluation of the spatial resolution, ambiguity, limitations, and general effectiveness of the various geophysical methods supported by technical discussions or references to the published literature.
- A detailed analysis of the effects of temperature, pressure and fluid content on rock resistivity.
- A study and critique of selected geophysical surveys completed by the IGME, Spain.
- An exploration strategy for three different geothermal resource geologic occurrence models: sedimentary basins, volcanic areas, and granitic araes.

The details of the bibliographic search and an in-depth summary of the results, together with the bibliographic listing, is included in an accompanying report (West and Ross, 1985). All other aspects of the study are documented in this report. The present study was limited both by time and funding level but draws upon the extensive experience and publication record of the ESL/UURI and therefore achieves a state-of-the-art summary of the application of geophysical methods in geothermal exploration. A substantial effort was required in translating technical reports from Spanish to English to insure ESL/UURI scientists had a complete understanding of the IGME exploration program.

III. GEOLOGIC OCCURRENCE OF GEOTHERMAL RESOURCES

Geothermal energy is heat energy that originates within the earth. Under suitable circumstances a small portion of this energy can be extracted and used by man. So active is the earth as a thermal engine that many of the large-scale geological processes that have helped to form the earth's surface features are powered by redistribution of internal heat as it flows from inner regions of higher temperature to outer regions of lower temperature. Such seemingly diverse phenomena as motion of the earth's crustal plates, uplifting of mountain ranges, occurrence of earthquakes, eruption of volcanoes and spouting of geysers all owe their origin to the transport of internal thermal energy.

In the United States and in many other countries, geothermal energy is used both for generation of electrical power and for direct applications such as space heating and industrial process energy. Although the technical viability of geothermal energy for such uses has been known for many years, the total amount of application today is very small compared with the potential for application. Availability of inexpensive energy from fossil fuels has suppressed use of geothermal resources. At present geothermal application is economic only at a few of the highest-grade resources. Development of new techniques and equipment to decrease costs of exploration, drilling, reservoir evaluation and extraction of the energy is needed to make the vastly more numerous lower grade resources also economic.

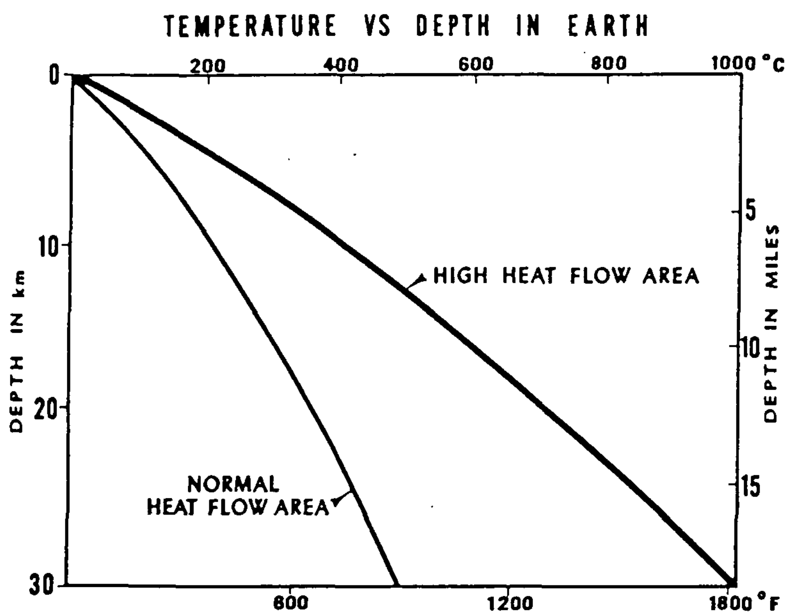
The objective of this chapter is to present an overview of the exploration for and exploitation of geothermal resources. The geological principles discussed have world-wide application. Geothermal resources of high temperature are found mainly in areas where a number of specific geologic processes are active today and resources of lower temperature are more widespread. A classification for observed resource types is presented and the geology of each type briefly described.

Overview of Geologic Processes

Although the distributions with depth in the earth of density, pressure and other related physical parameters are well known, the temperature distribution is extremely uncertain. We do know that temperature within the earth increases with increasing depth (Fig. 1), at least for the first few tens of kilometers, and we hypothesize a steadily increasing temperature to the earth's center. Plastic or partially molten rock at estimated temperatures between 700°C and 1200°C is postulated to exist everywhere beneath the earth's surface at depths of 100 km, and the temperature at the earth's center, nearly 6400 km deep, may be more than 4000°C.

Because the earth is hot inside, heat flows steadily outward over the entire surface, where it is permanently lost by radiation into space. The mean value of this surface heat flow for the world is about 60 X 10^{-3} watts/m^2 (White and Williams, 1975) and since the mean surface area of the earth is about 5.1 X 10^{14} m^2, the rate of heat loss is about 32 X 10^{12} watts (32 million megawatts) or about 2.4 X 10^{20} calories/year, a very large amount indeed. At present, only a small portion of this heat, namely that concentrated in what we call geothermal resources, can be captured for man's benefit. The mean surface heat flux of 60 milliwatts/m^2 is about 20,000 times smaller than the heat arriving from the sun when it is directly overhead, and the earth's surface temperature is thus controlled by the sun and not by heat from the interior (Goguel, 1976).

Two ultimate sources for the earth's internal heat appear to be most important among a number of contributing alternatives: 1) heat released throughout the earth's 4.5 billion-year history by radioactive decay of



TEMPERATURE VS DEPTH IN EARTH
200
400
600
800
1000 °C
0
10
20
30
DEPTH IN km
5
10
15
DEPTH IN MILES
HIGH HEAT FLOW AREA
NORMAL
HEAT FLOW AREA
600
1200
1800 °F

Figure 1.1.1

certain isotopes of uranium, thorium, potassium, and other elements; and 2) heat released during formation of the earth by gravitational accretion and during subsequent mass redistribution when much of the heavier material sank to form the earth's core (Fig. 2). The relative contribution to the observed surface heat flow of these two mechanisms is not yet resolved. Some theoretical models of the earth indicate that heat produced by radioactive decay can account for nearly all of the present heat flux (MacDonald, 1965). Other studies (Davis, 1980) indicate that, if the earth's core formed by sinking of the heavier metallic elements in an originally homogeneous earth, the gravitational heat released would have been sufficient to raise the temperature of the whole earth by about 2000°C. An appreciable fraction of today's observed heat flow could be accounted for by such a source. However, the distribution of radioactive elements within the earth is poorly known, as is the earth's early formational history some 4 billion years ago. We do know that the thermal conductivity of crustal rocks is low so that heat escapes from the surface slowly. The deep regions of the earth retain a substantial portion of their original heat, whatever its source, and billions of years will pass before the earth cools sufficiently to quiet the active geological processes we will discuss below. This fact helps lend order to exploration for geothermal resources once the geological processes are understood. At present our understanding of these processes is rather sketchy, but, with rapidly increasing need for use of geothermal resources as an alternative to fossil fuels, our learning rate is high.

Figure 3 shows the principal areas of known geothermal occurrences on a world map. Also indicated are areas of young volcanoes and a number of currently active fundamental geological structures. It is readily seen that many geothermal resource areas correspond with areas that now have or recently

INTERIOR OF THE EARTH

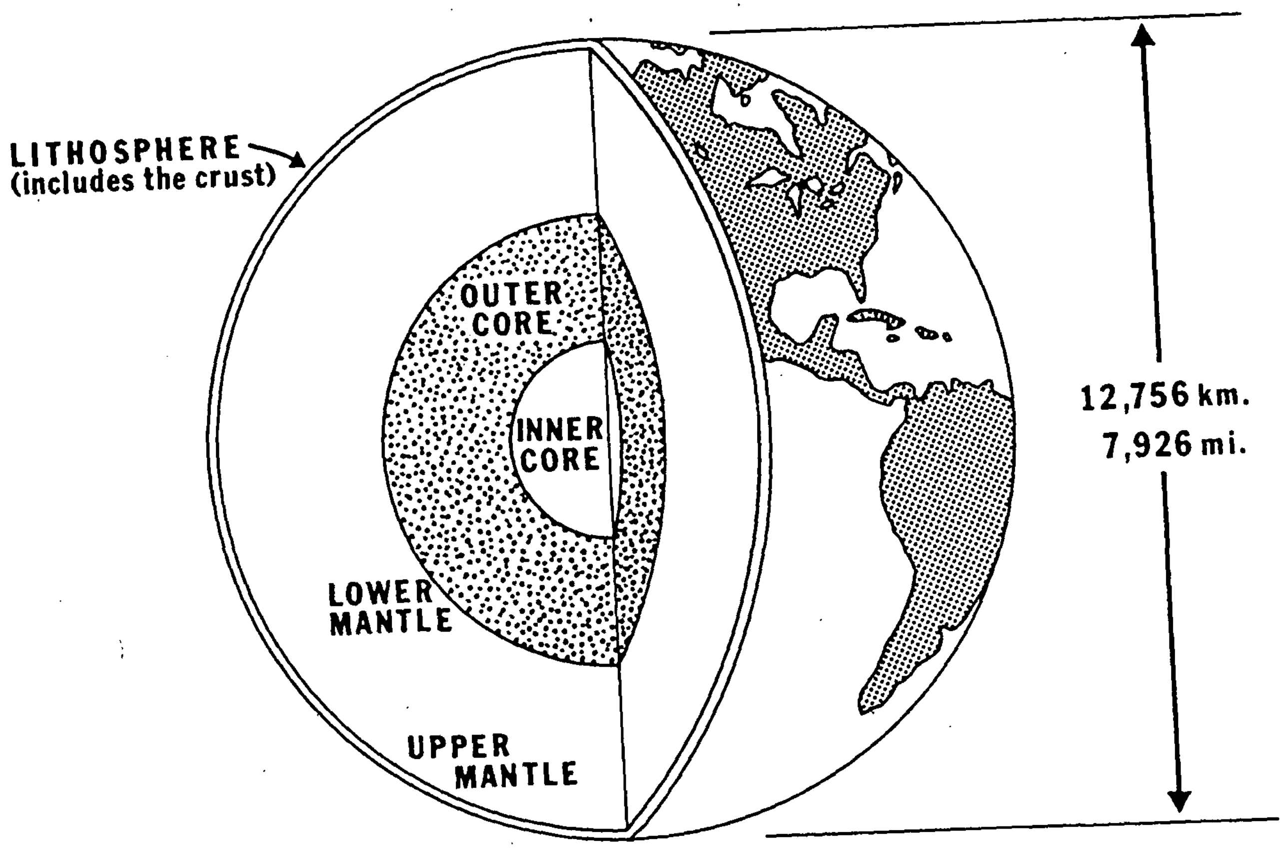
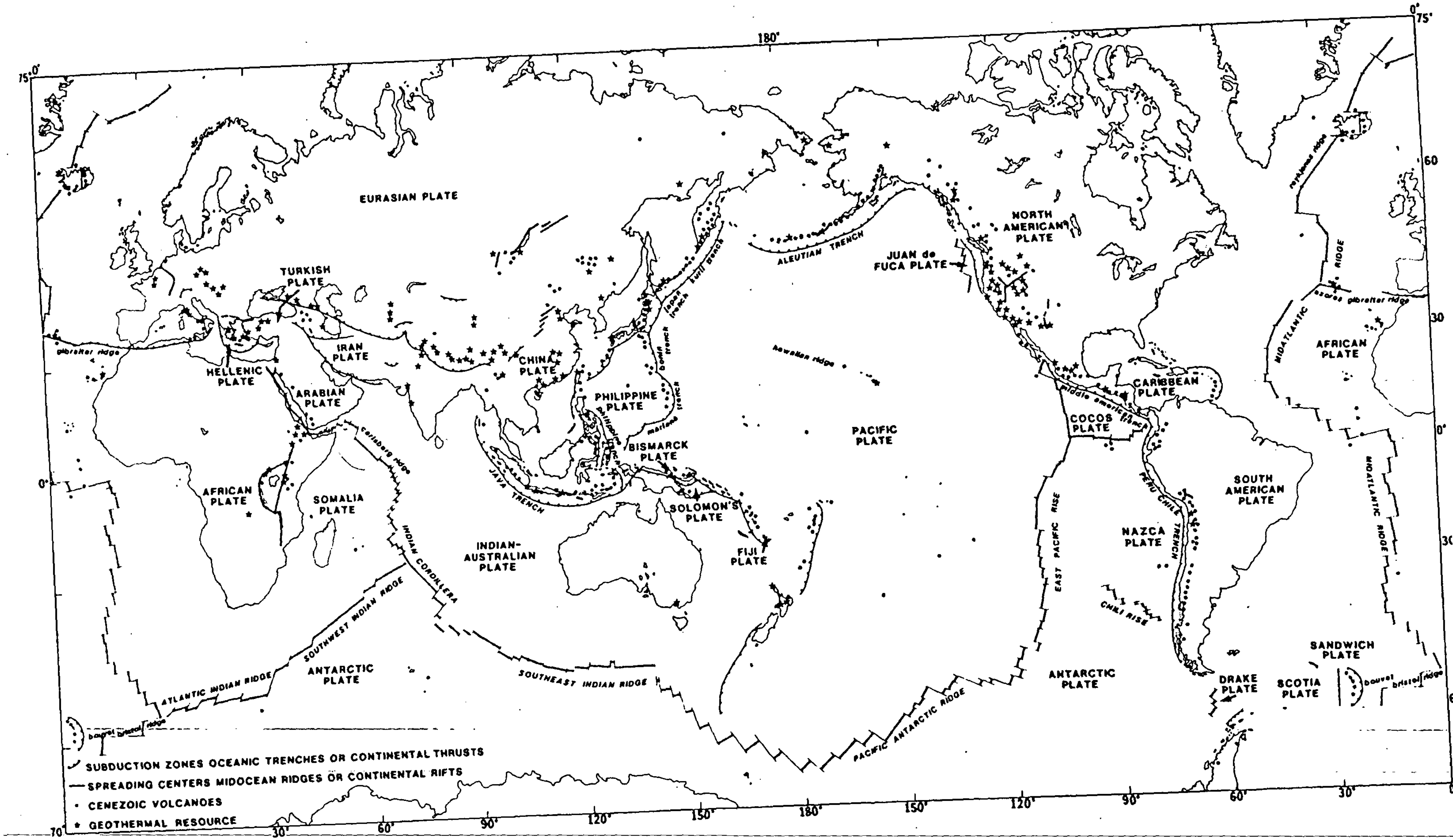


Figure 1.1.2



GEOTHERMAL RESOURCES AND PLATE TECTONIC FEATURES

Figure 1.1.3

have had volcanic and other geological activity. To understand why this is true we must consider some of the geologic processes going on in the earth's interior.

A schematic cross section of the earth is shown in Figure 2. A solid layer called the lithosphere extends from the surface to a depth of about 100 km. The lithosphere is composed of an uppermost layer called the crust and of the uppermost regions of the mantle, which lie below the crust. Mantle material below the lithosphere is less solid than the overlying lithosphere and is able to flow very slowly under sustained stress. The crust and the mantle are composed of minerals whose chief building block is silica (SiO_2). The outer core is a region where material is much denser than mantle material, and it is believed to be composed of a liquid iron-nickel-copper mixture. The inner core is believed to be a solid metallic mixture.

One very important group of geological processes that cause geothermal resources is known collectively as "plate tectonics" (Wyllie, 1971). It is illustrated in Figure 4. Outward flow of heat from the deep interior is hypothesized to cause formation of convection cells in the earth's mantle in which deeper, hotter mantle material slowly rises toward the surface, spreads out parallel to the surface under the solid lithosphere as it cools and, upon cooling, descends again. The lithosphere above the upwelling portions of these convection cells cracks and spreads apart along linear or arcuate zones called "spreading centers" that are typically thousands of kilometers long and coincide, for the most part, with the world's mid-oceanic ridge or mountain system (Figs. 3 and 4). The crustal plates on each side of the crack or rift move apart at rates of a few centimeters per year, and molten mantle material rises in the crack and solidifies to form new crust. The laterally moving oceanic lithospheric plates impinge against adjacent plates, some of which

CONCEPT OF PLATE TECTONICS

(NOT TO SCALE)

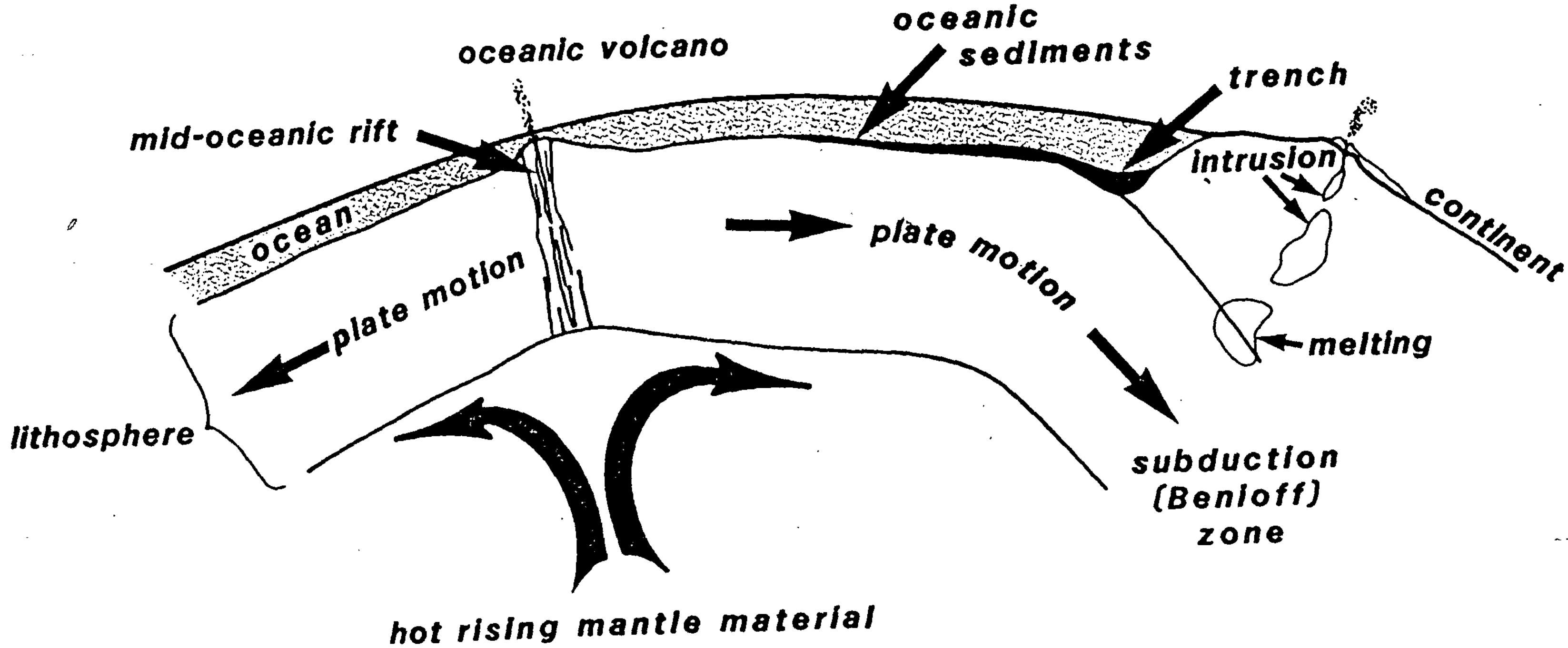


Figure 1.1.4

contain the imbedded continental land masses, and in most locations the oceanic plates are thrust beneath the continental plates. These zones of under-thrusting, called subduction zones, are marked by the world's deep oceanic trenches which result from the crust being dragged down by the descending oceanic plate. The oceanic plate descends into regions of warmer material in the mantle and is warmed both by the surrounding warmer material and by frictional heating as it is thrust downward. At the upper boundary of the descending plate, temperatures become high enough in places to cause partial melting. The degree of melting depends upon the amount of water contained in the rocks as well as upon temperature and pressure and the upper layers of the descending plate often contain oceanic sediments rich in water. The molten or partially molten rock bodies (magmas) that result then ascend buoyantly through the crust, probably along lines of structural weakness (Fig. 5) and carry their contained heat to within 1.5 to 15 km of the surface. They give rise to volcanoes if part of the molten material escapes to the surface through faults and fractures in the upper crust.

Figure 3 shows where these processes of crustal spreading, formation of new oceanic crust from molten mantle material and subduction of oceanic plates beneath adjacent plates, are currently operating. Oceanic rises, where new crustal material is formed, occur in all of the major oceans. The East Pacific Rise, the Mid-Atlantic Ridge and the Indian ridges are examples. The ridge or rise crest is offset in places by large transform faults that result from variations in the rate of crustal spreading from place to place along the ridge. Oceanic crustal material is subducted or consumed in the trench areas. Almost all of the world's earthquakes result from these large-scale processes, and occur either at the spreading centers, the transform faults or in association with the subduction zone (Benioff zone), which dips underneath

CRUSTAL INTRUSION

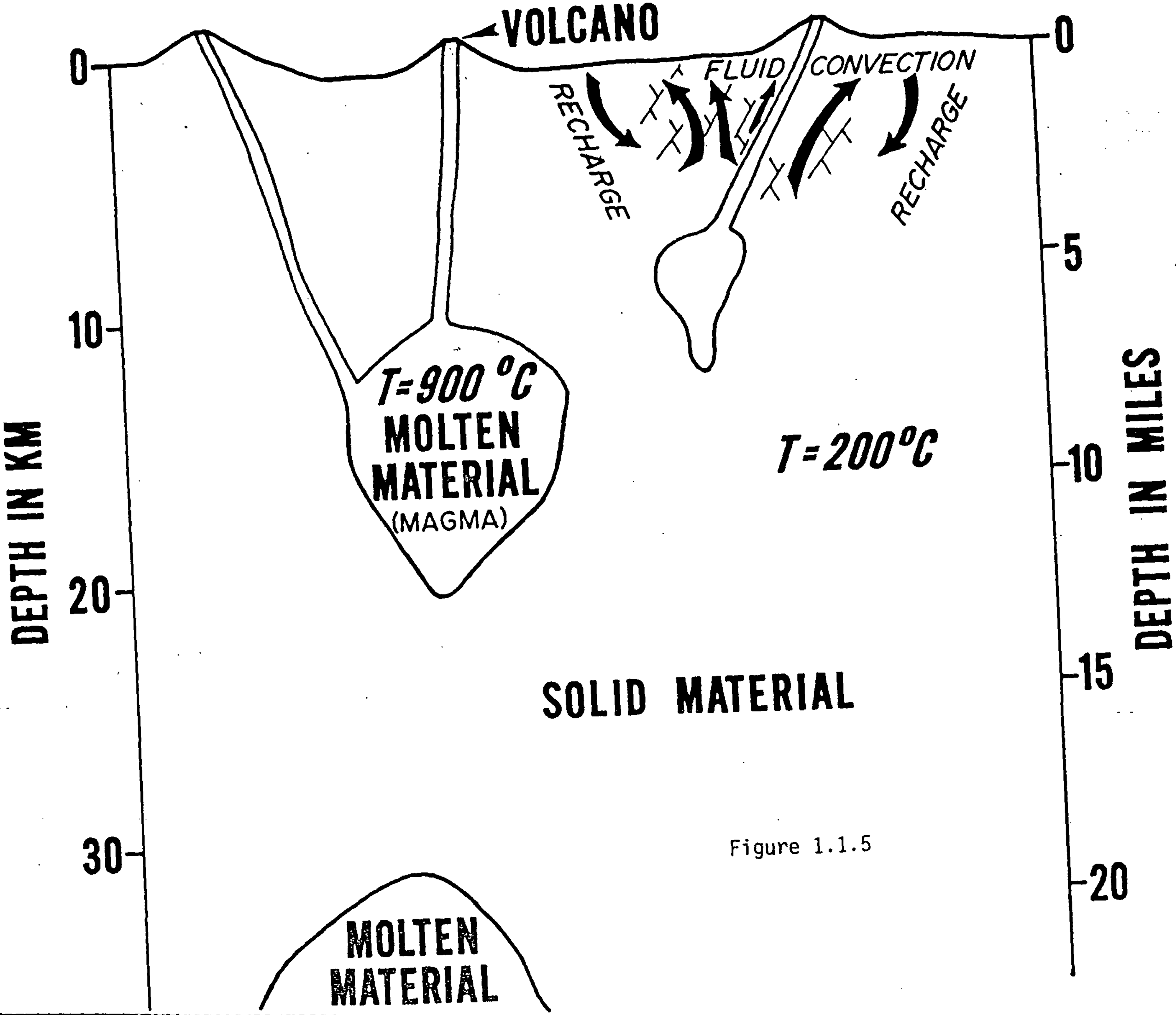


Figure 1.1.5

the continental land masses in many places. We thus see that these very active processes of plate tectonics give rise to diverse phenomena, among which is the generation of molten rock at shallow depths in the crust both at the spreading centers and above zones of subduction. These bodies of shallow molten rock provide the heat for many of the world's geothermal resources.

Before going on, let us discuss a bit more the processes of development of a crustal intrusion, illustrated in Figure 5. An ascending body of molten material may cease to rise at any level in the earth's crust and may or may not vent to the surface in volcanoes. Intrusion of molten magmas into the upper parts of the earth's crust has gone on throughout geological time. We see evidence for this in the occurrence of volcanic rocks of all ages and in the small to very large areas of crystalline, granitic rock that result when such a magma cools slowly at depth.

Volcanic rocks that have been extruded at the surface and crystalline rocks that have cooled at depth are known collectively as igneous rocks. They vary over a range of chemical and mineral composition. At one end of the range are rocks that are relatively poor in silica (SiO_2 about 50%) and relatively rich in iron (Fe_2O_3 + FeO about 8%) and magnesium (MgO about 7%). The volcanic variety of this rock is basalt and an example is the black rocks of the Hawaiian Islands. The crystalline, plutonic variety of this rock that has consolidated at depth is known as gabbro. At the other end of the range are rocks that are relatively rich in silica (SiO_2 about 64%) and poor in iron (Fe_2O_3 + FeO about 5%) and magnesium (MgO about 2%). The volcanic variety of this rock, rhyolite, is usually lighter in color than the black basalt and it occurs mainly on land. The plutonic variety of this rock is granite, although the term "granitic" is sometimes used for any crystalline igneous rock. Magmas that result in basalt or gabbro are termed "basic" whereas magmas that result in rhyolite or granite are termed "acidic"; however these terms are misleading because they have nothing to do with the pH of the magma.

The upper portions of the mantle are believed to be basaltic in composition. The great outpourings of basalt seen in places like the Hawaiian Islands and on the volcanic plateaus of the Columbia and Snake rivers in the northwestern United States seem to indicate a more or less direct pipeline from the upper mantle to the surface in places. The origin of granites is a subject of some controversy. It can be shown that granitic magmas could be derived by differential segregation from basaltic magmas. However, the chemical composition of granites is much like the average composition of the continental crust, and some granites probably result from melting of crustal rocks by upwelling basaltic magmas whereas others probably result from differentiation from a basaltic magma. In any case, basaltic magmas are molten at a higher temperature than are granitic magmas (see Fig. 6) and more importantly for our discussion basaltic magmas are less viscous (more fluid) than are granitic magmas. Occurrence of rhyolitic volcanic rocks of very young age (less than 1 million years and preferably less than 50,000 years) is generally taken as a sign of good geothermal potential in an area because presumably a large body of viscous magma may be indicated at depth to provide a geothermal heat source. On the other hand, occurrence of young basaltic magma is not as encouraging because the basalt, being fairly fluid, could simply ascend along narrow conduits from the mantle directly to the surface without need for a shallow magma chamber that would provide a geothermal heat source. In many areas both basaltic and rhyolitic volcanic rocks are present and often the younger eruptions are more rhyolitic, possibly indicating progressive differentiation of an underlying basaltic magma in a chamber like those illustrated in Figure 5.

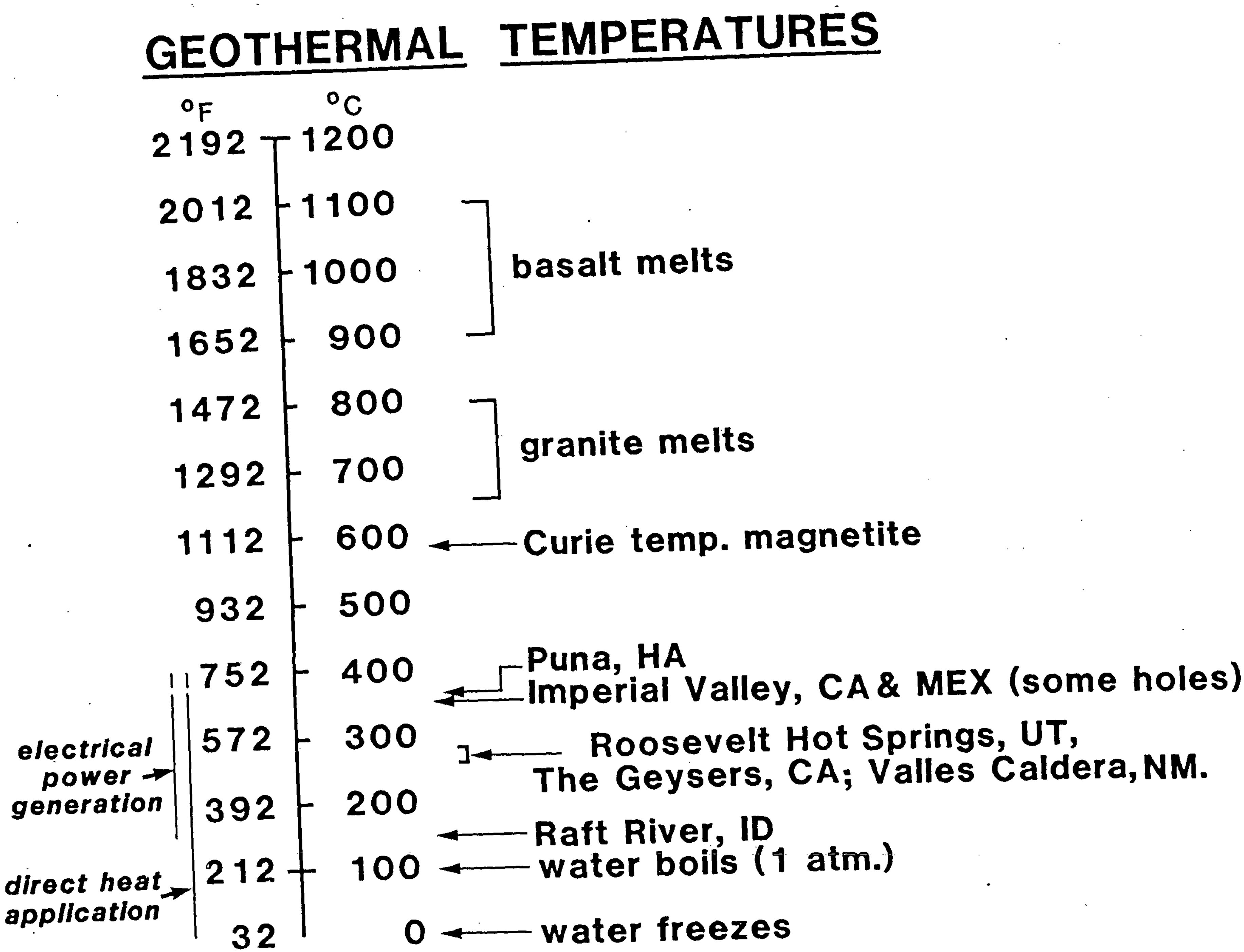

GEOTHERMAL TEMPERATURES
°F
°C
2192
1200
2012
1100
1832
1000
1652
900
1472
800
1292
700
1112
600
932
500
752
400
572
300
392
200
212
100
32
0
basalt melts
granite melts
Curie temp. magnetite
Puna, HA
Imperial Valley, CA & MEX (some holes)
Roosevelt Hot Springs, UT,
The Geysers, CA; Valles Caldera, NM.
Raft River, ID
water boils (1 atm.)
water freezes
electrical power generation
direct heat application

Figure 1.1.6

A second important source of volcanic rocks results from hypothesized point sources of heat in the mantle as contrasted with the rather large convection cells discussed above. It has been hypothesized that the upper mantle contains local areas of upwelling, hot material called plumes, although other origins for the hot spots have also been postulated. As crustal plates move over these local hot spots, a linear or arcuate sequence of volcanoes is developed. Young volcanic rocks occur at one end of the volcanic chain with older ones at the other end. The Hawaiian Island chain is an excellent example. Volcanic rocks on the island of Kauai at the northwest end of the chain have been dated through radioactive means at about 6 million years, whereas the volcanoes Mauna Loa and Mauna Kea on the island of Hawaii at the southeast end of the chain are in almost continual activity, at the present time having an interval between eruptions of only 11 months. In addition, geologists speculate that Yellowstone National Park, Wyoming, one of the largest geothermal areas in the world, sits over such a hot spot and that the older volcanic rocks of the eastern and western Snake River plains in Idaho are the surface trace of this mantle hot spot in the geologic past (see Fig. 16 and the discussion below).

Not all geothermal resources are caused by near-surface intrusion of molten rock bodies. Certain areas have a higher than average rate of increase in temperature with depth (high geothermal gradient) without shallow magma being present. Much of the western United States contains areas that have an anomalously high mean heat flow (100 mwatt/m^2) and an anomalously high geothermal gradient (50°C/km). Geophysical and geological data indicate that the earth's crust is thinner than normal and that the isotherms are upwarped beneath this area. Much of the western U.S. is geologically active, as manifested by earthquakes and active or recently active volcanoes. Faulting and fracturing during earthquakes help to keep fracture systems open, and this allows circulation of ground water to depths of 2 km to perhaps 5 km. Here the water is heated and rises buoyantly along other fractures to form geothermal resources near surface. Many of the hot springs and wells in the western United States and elsewhere owe their origin to such processes.

Geothermal Resource Types

All geothermal resources have three common components:

1) a heat source
2) permeability in the rock, and
3) a heat transfer fluid.

In the foregoing we have considered some of the possible heat sources, and we will discuss others presently. Let us now consider the second component, permeability.

Permeability is a measure of how easily fluids flow through rock as a result of pressure differences. Of course fluid does not flow through the rock matrix itself but rather it flows in open spaces between mineral grains and in fractures. Rocks in many, but not all, geothermal areas are very solid and tight, and have little or no interconnected pore space between mineral grains. In such rocks the only through-going pathways for fluid flow are cracks or fractures in the rock. A geothermal well must intersect one or more fractures if the well is to produce geothermal fluids in quantity, and it is generally the case that these fractures can not be located precisely by means of surface exploration. Fractures sufficient to make a well a good producer need only be a few millimeters in width, but must be connected to the general fracture network in the rock in order to carry large fluid volumes.

The purpose of the heat transfer fluid is to remove the heat from the rocks at depth and bring it to the surface. The heat transfer fluid is either water (sometimes saline) or steam. Water has a high heat capacity (amount of heat needed to raise the temperature by 1°C) and a high heat of vaporization (amount of heat needed to convert 1 gm to steam). Thus water, which naturally pervades fractures and other open spaces in rocks, is an ideal heat transfer fluid because a given quantity of water or steam can carry a large amount of heat to the surface where it is easily removed.

Geothermal resource temperatures range upward from the mean annual ambient temperature (usually 10-30°C) to well over 350°C. Figure 6 shows the span of temperatures of interest in geothermal work.

The classifications of geothermal resource types shown in Table I is modeled after one given by White and Williams (1975). Each type will be described briefly. In order to describe these resource types we resort to simplified geologic models. A given model is often not acceptable to all geologists, especially at our rather primitive state of knowledge of geothermal resources today.

Hydrothermal Resources

Hydrothermal convection resources are geothermal resources in which the earth's heat is actively carried upward by the convective circulation of naturally occurring hot water or its gaseous phase, steam. Underlying some of the higher temperature hydrothermal resources is presumably a body of still molten or recently solidified rock that is very hot (300°C-1100°C). Other hydrothermal resources result simply from circulation of water along faults and fractures or within a permeable aquifer to depths where the rock temperature is elevated, with heating of the water and subsequent buoyant transport to the surface or near surface. Whether or not steam actually exists in a hydrothermal reservoir depends, among other less important variables, on temperature and pressure conditions at depth.

TABLE 1

GEOTHERMAL RESOURCE CLASSIFICATION
(After White and Williams, 1975)

Resource Type	Temperature Characteristics
1. Hydrothermal convection resources (heat carried upward from depth by convection of water or steam)	
a) Vapor dominated	about 240°C
b) Hot-water dominated	
i) High Temperature	150°C to 350°C+
ii) Intermediate	90°C to 150°C
iii) Low Temperature	less than 90°C
2. Hot rock resources (rock intruded in molten form from depth)	
a) Part still molten	higher than 600°C
b) Not molten (hot dry rock)	90°C to 650°C
3. Other resources	
a) Sedimentary basins (hot fluid in sedimentary rocks)	30°C to about 150°C
b) Geopressured (hot fluid under high pressure)	150°C to about 200°C
c) Radiogenic (heat generated by radioactive decay)	30°C to about 150°C

Figure 7 (after White et al., 1971) shows a conceptual model of a hydrothermal system where steam is present, a so-called vapor-dominated hydrothermal system (1a of Table 1). Convection of deep saline water brings a large amount of heat upward from depth to a level where boiling can take place under the prevailing temperature and pressure conditions. Steam moves upward through fractures in the rock and is possibly superheated further by the hot surrounding rock. Heat is lost from the vapor to the cooler, near-surface rock and condensation results, with some of the condensed water moving downward to be vaporized again. Within the entire vapor-filled part of the reservoir, temperature is nearly uniform due to rapid fluid convection. This whole convection system can be closed, so that the fluid circulates without loss, but if an open fracture penetrates to the surface, steam may vent. In this case, water lost to the system would be replaced by recharge, which takes place mainly by cool ground water moving downward and into the convection system from the margins. The pressure within the steam-filled reservoir increases much more slowly with depth than would be the case if the reservoir were filled with water under hydrostatic pressure. Because the rocks surrounding the reservoir will generally contain ground water under hydrostatic pressure, there must exist a large horizontal pressure differential between the steam in the reservoir and the water in the adjacent rocks, and a significant question revolves around why the adjacent water does not move in and inundate the reservoir. It is postulated that the rock permeability at the edges of the reservoir and probably above also, is either naturally low or has been decreased by deposition of minerals from the hydrothermal fluid in the fractures and pores to form a self-sealed zone around the reservoir. Self-sealed zones are known to occur in both vapor-dominated and water-dominated resources.

VAPOR DOMINATED GEOTHERMAL RESERVOIR

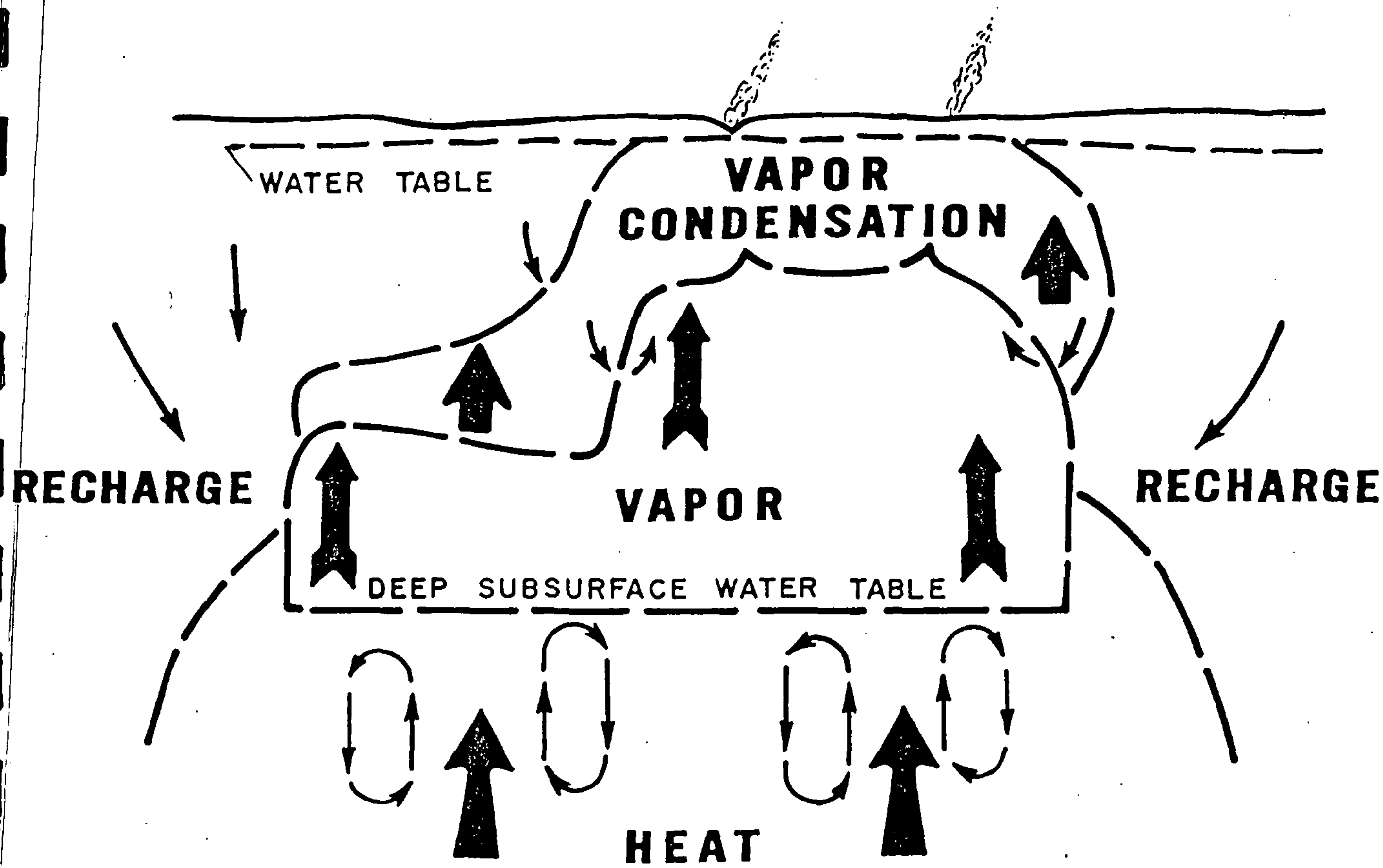


Figure 1.1.7

A well drilled into a vapor-dominated reservoir would produce superheated steam. The Geysers geothermal area in California is an example of this type of resource. Steam is produced from wells whose depths are 1.5 to 3 km, and this steam is fed to turbine generators that produce electricity. The current generating capacity at The Geysers is 1454 MWe (megawatts of electrical power, where 1 megawatt = 1 million watts). This compares to the world current total from all geothermal resource types of 3790 MWe.

Other vapor-dominated resources that are currently being exploited occur at Lardarello and Monte Amiata, Italy, and at Matsukawa, Japan. The famous Yellowstone National Park in Wyoming contains many geysers, fumaroles, hot pools and thermal springs, and the Mud Volcanoes area is believed to be underlain by a dry steam field.

There are relatively few known vapor-dominated resources in the world because special geological conditions are required for their formation (White et al., 1971). However, they are eagerly sought because they are generally easier and less expensive to develop than the more common water-dominated system discussed below.

Figure 8 schematically illustrates a high-temperature, hot-water-dominated hydrothermal system (1b(i) of Table 1). The source of heat beneath many such systems is probably molten rock or rock that has solidified only in the last few tens of thousands of years, lying at a depth of perhaps 3 to 10 km. Normal ground water circulates in open fractures and removes heat from these deep, hot rocks by convection. Fluid temperatures are uniform over large volumes of the reservoir because convection is rapid. Recharge of cooler ground water takes place at the margins of the system through circulation down fractures. Escape of hot fluids at the surface is often minimized by a near-surface sealed zone or cap-rock formed by precipitation from the geother-

WATER DOMINATED GEOTHERMAL SYSTEM

FLOW CONTROLLED BY FRACTURES

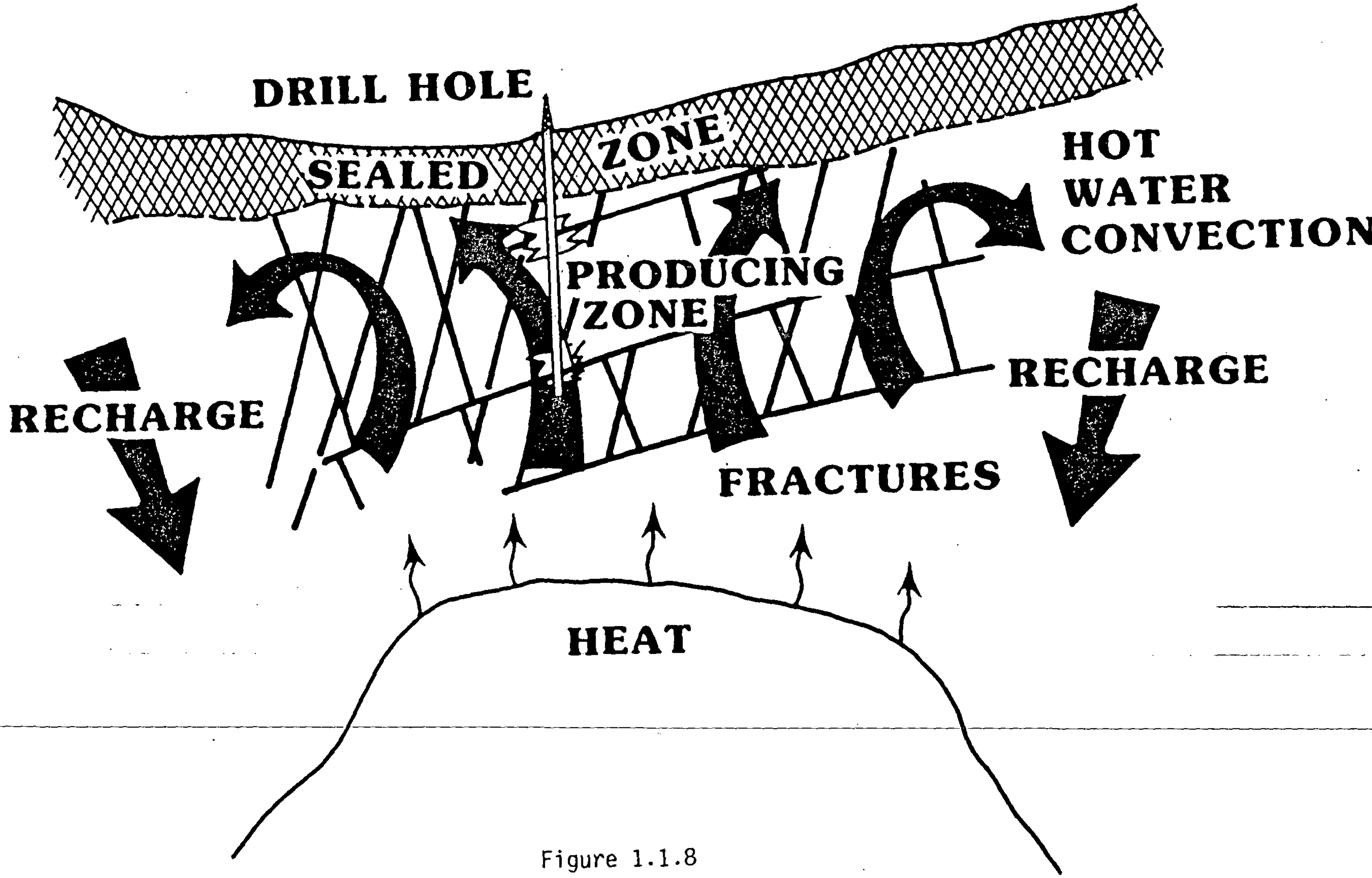


Figure 1.1.8

mal fluids of minerals in fractures and pore spaces. Surface manifestations of such a geothermal system might include hot springs, fumaroles, geysers, thermal spring deposits, chemically altered rocks, or alternatively, no surface manifestation may occur at all. If there are no surface manifestations, discovery is much more difficult and requires sophisticated geology, geophysics, geochemistry and hydrology. A well drilled into a water-dominated geothermal system would likely encounter tight, hot rocks with hot water inflow from the rock into the well bore mainly along open fractures. Areas where different fracture sets intersect may be especially favorable for production of large volumes of hot water. For generation of electrical power a portion of the hot water produced from the well is allowed to flash to steam within the well bore or within surface equipment as pressure is reduced, and the steam is used to drive a turbine generator.

A second type of hot-water dominated system is shown in Figure 9. Here the reservoir rocks are sedimentary rocks that have intergranular permeability as well as fracture permeability. Geothermal fluids can sometimes be produced from such a reservoir without the need to intersect open fractures by a drill hole. Examples of this resource type occur in the Imperial Valley of California and Mexico. In this region the East Pacific Rise, a crustal spreading center, comes onto the North American continent. Figure 3 shows that the rise is observed to trend northward up the Gulf of California in small segments that are repeatedly offset northward by transform faults. Although its location under the continent cannot be traced very far with certainty, it is believed to occur under and be responsible for the Imperial Valley geothermal resources. The source of the heat is upwelling, very hot molten or plastic material from the earth's mantle. This hot rock heats overlying sedimentary rocks and their contained fluids and has spawned volcanoes. The locations of

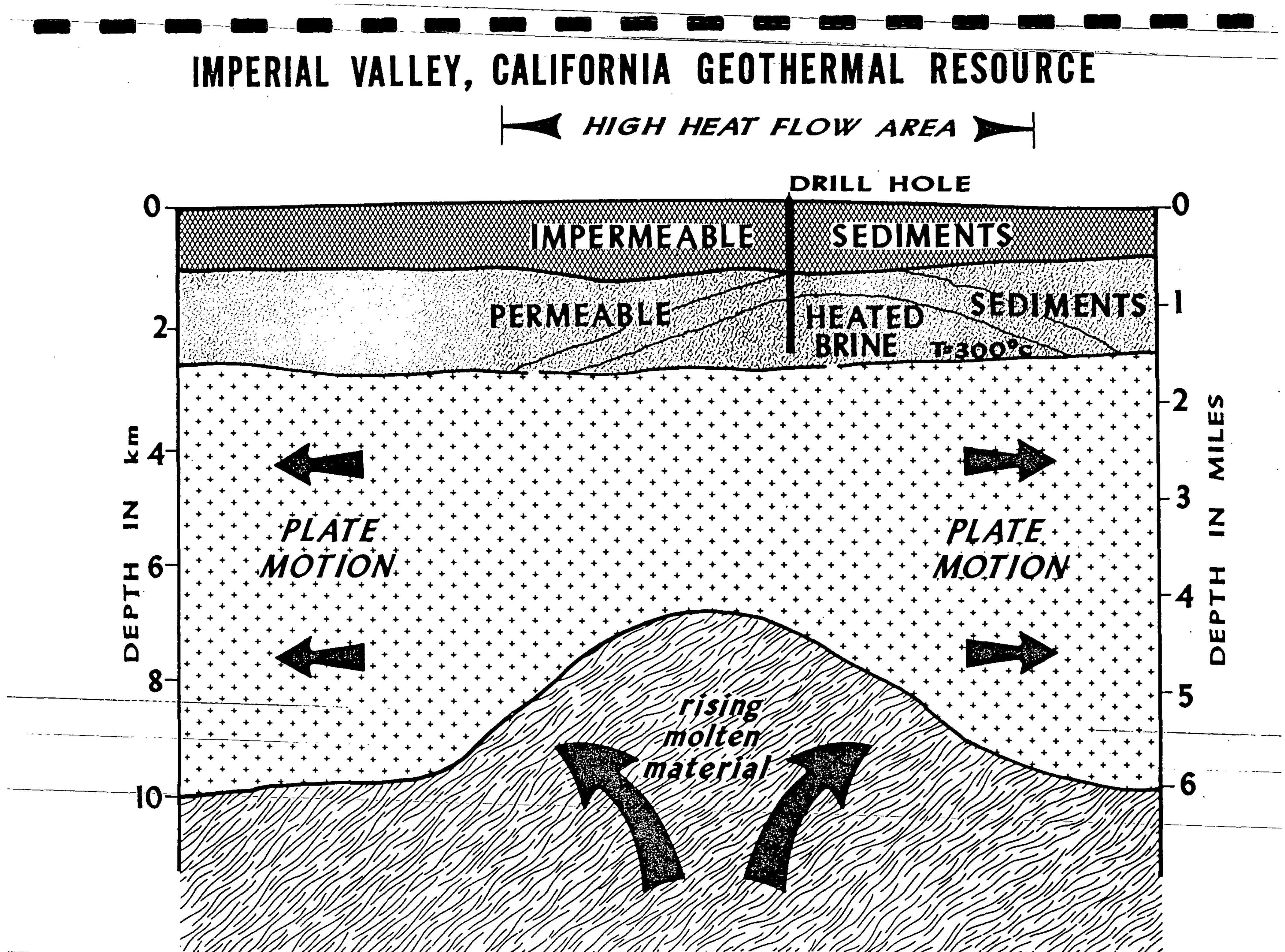

IMPERIAL VALLEY, CALIFORNIA GEOTHERMAL RESOURCE
HIGH HEAT FLOW AREA
DRILL HOLE
IMPERMEABLE
SEDIMENTS
PERMEABLE
HEATED
BRINE
SEDIMENTS
T=300°C
PLATE
MOTION
PLATE
MOTION
rising
molten
material
DEPTH IN km
0
2
4
6
8
10
DEPTH IN MILES
0
1
2
3
4
5
6

Figure 1.1.9

specific resource areas appear to be controlled by faults that presumably allow deep fluid circulation to carry the heat upward to reservoir depths.

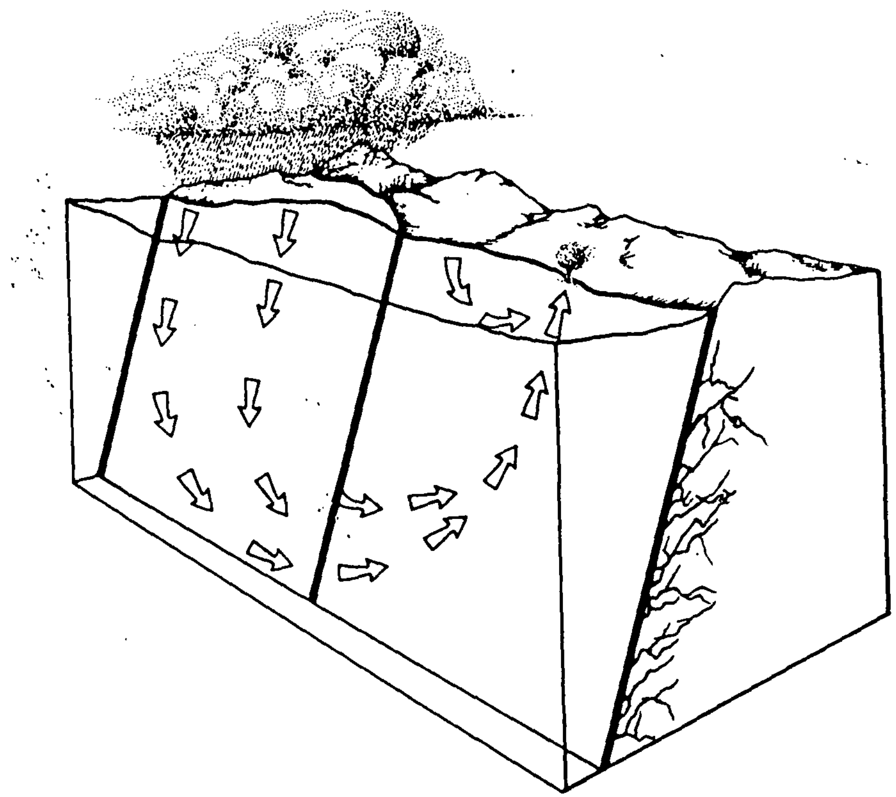
The fringe areas of high-temperature vapor- and water-dominated hydrothermal systems often produce water of low and intermediate temperature (1b(ii) and 1b(iii) of Table 1). These lower temperature fluids are suitable for direct heat applications but not for electrical power production. Low- and intermediate-temperature waters can also result from deep water circulation in areas where heat conduction and the geothermal gradient are merely average, as previously discussed. Waters circulated to depths of 1 to 5 km are warmed in the normal geothermal gradient and they return to the surface or near surface along open fractures because of their buoyancy (Fig. 10). There need be no enhanced gradient or magmatic heat source under such an area. Warm springs occur where these waters reach the surface, but if the warm waters do not reach the surface they are generally difficult to find.

Sedimentary Basins

Some basins are filled to depths of 3 km or more with sedimentary rocks that have intergranular and open-space permeability. In some of these sedimentary units, circulation of ground water can be very deep. Water may be heated in a normal or enhanced geothermal gradient and may then either return to the near-surface environment or remain trapped at depth (3a of Table 1). Figures 11a and 11b illustrate these resources. Substantial benefit is being realized in France from use of this type of resource for space heating by production of warm water contained in the Paris basin. Many other areas of occurrence of this resource type are known worldwide.

Geopressured Resources

Geopressured resources (3b of Table 1) consist of deeply buried fluids



MODEL OF DEEP CIRCULATION HYDROTHERMAL RESOURCE

Figure 1.1.10

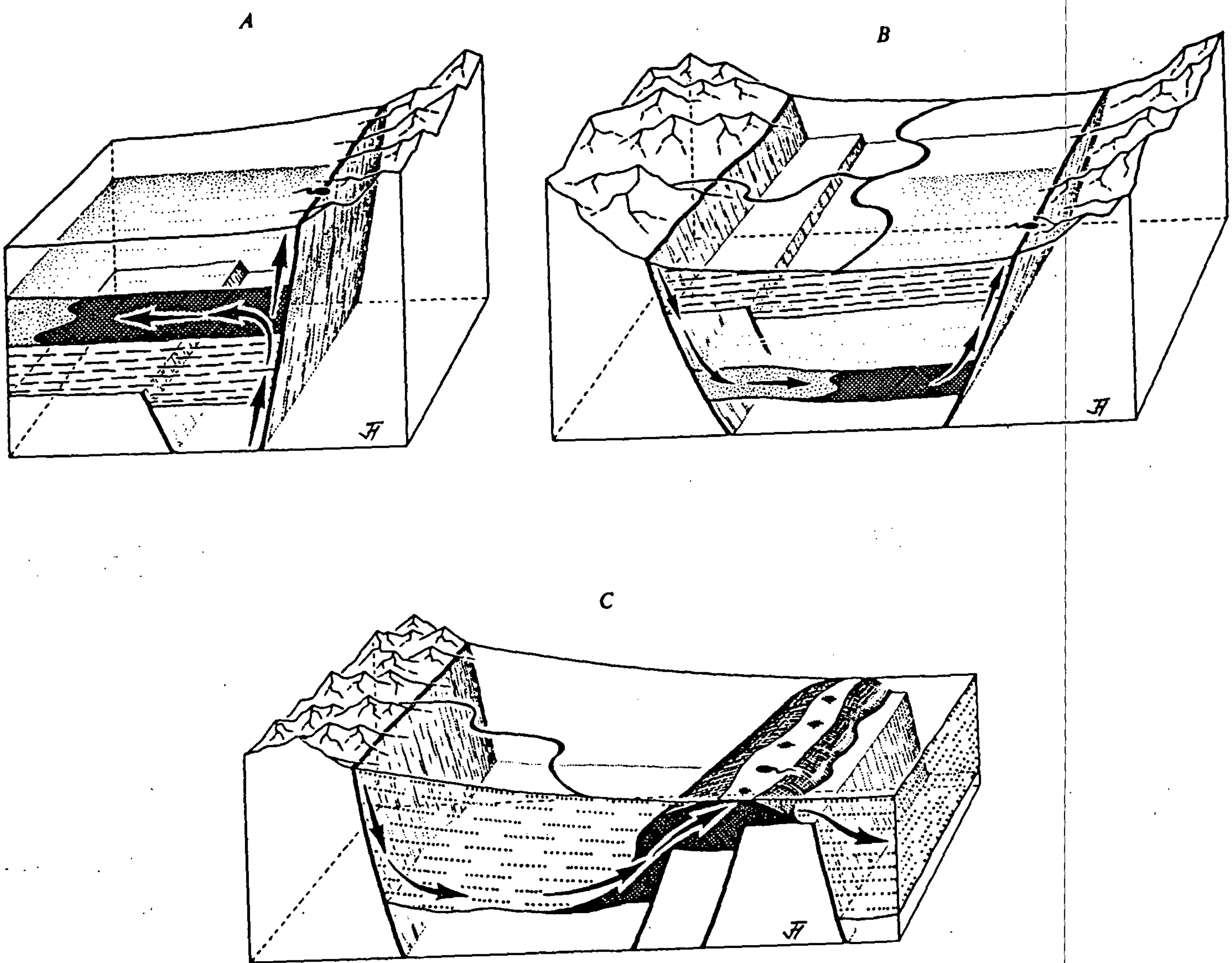


Figure 1.1.11a

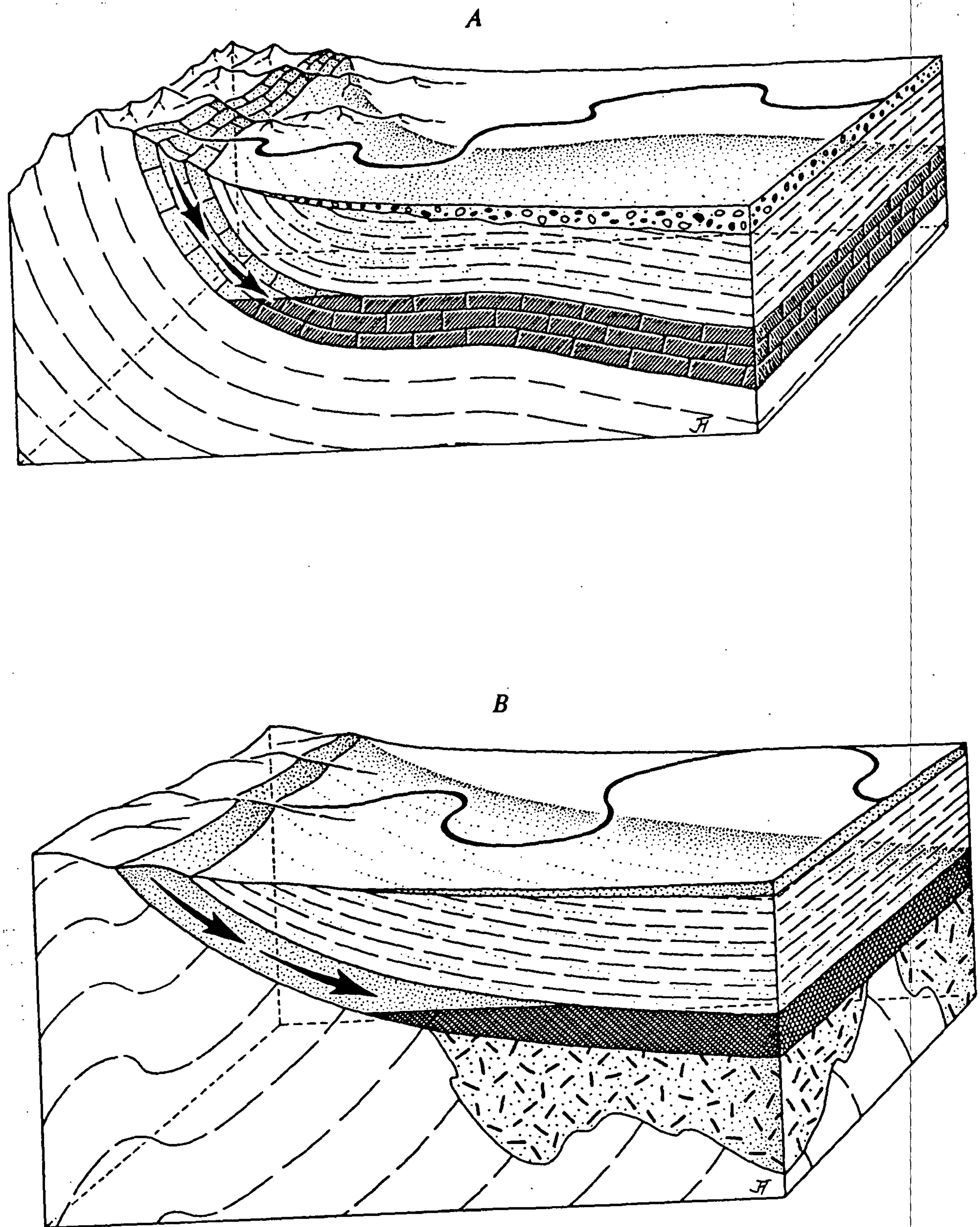


Figure 1.1.11b

contained in permeable sedimentary rocks warmed in a normal or anomalous geothermal gradient by their great burial depth. These fluids are tightly confined by surrounding impermeable rock and thus bear pressure that is much greater than hydrostatic, that is, the fluid pressure supports a portion of the weight of the overlying rock column as well as the weight of the water column. Figure 12 (from Papadopulos, 1975) gives a few typical parameters for geopressured reservoirs and illustrates the origin of the above-normal fluid pressure. These geopressured fluids may contain dissolved methane. Therefore, three sources of energy are actually available from such resources: 1) heat, 2) mechanical energy due to the great pressure with which these waters exit the borehole, and 3) the recoverable methane.

Radiogenic Geothermal Resource

Radiogenic geothermal resources are found in places such as the eastern U.S. (3c of Table 1). The coastal plain is blanketed by a layer of thermally insulating sediments. In places beneath these sediments are intrusions having enhanced heat production due to higher content of radioactive U, Th, and K are believed to occur. Geophysical and geological methods for locating such radiogenic rocks beneath the sedimentary cover are being developed, and drill testing of the entire geothermal target concept (Fig. 13) are being completed. Success would most likely come in the form of low- to intermediate-temperature geothermal waters suitable for space heating and industrial processing.

Hot Dry Rock Resource

Hot dry rock resources (2b of Table 1) are defined as heat stored in rocks within about 10 km of the surface from which the energy cannot be economically extracted by natural hot water or steam. These hot rocks have

GEOPRESSURED GEOTHERMAL RESOURCE

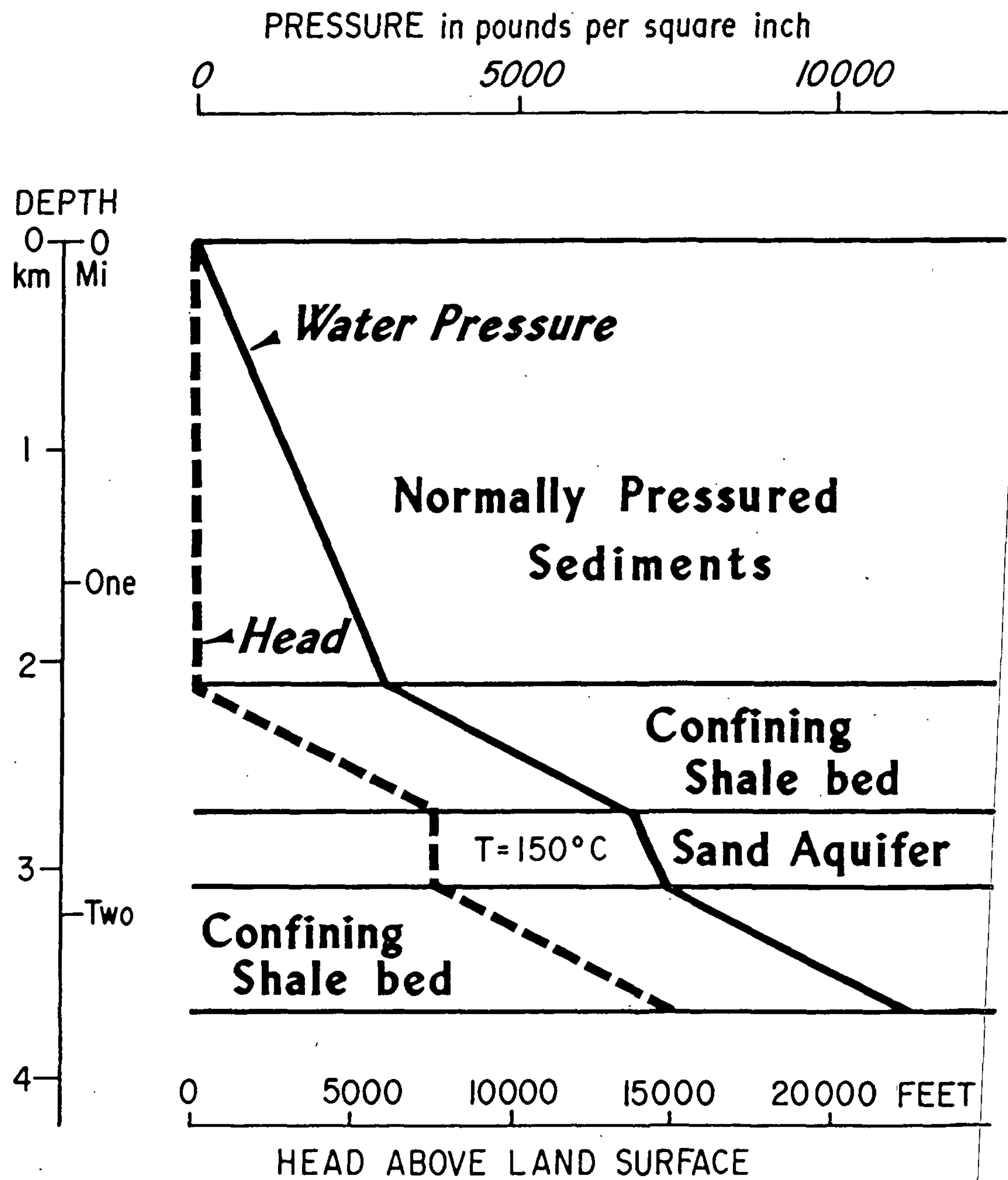


Figure 1.1.12

RADIOGENIC GEOTHERMAL RESOURCE

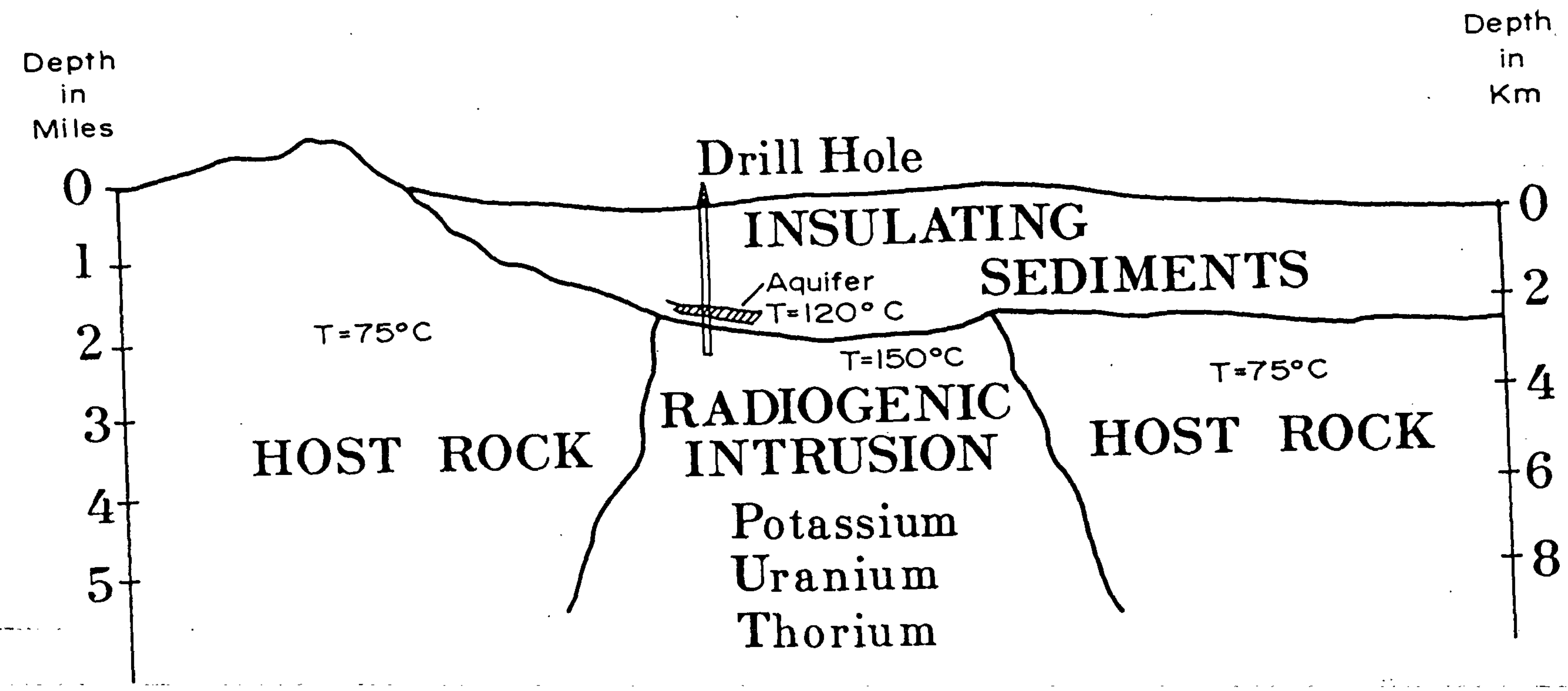
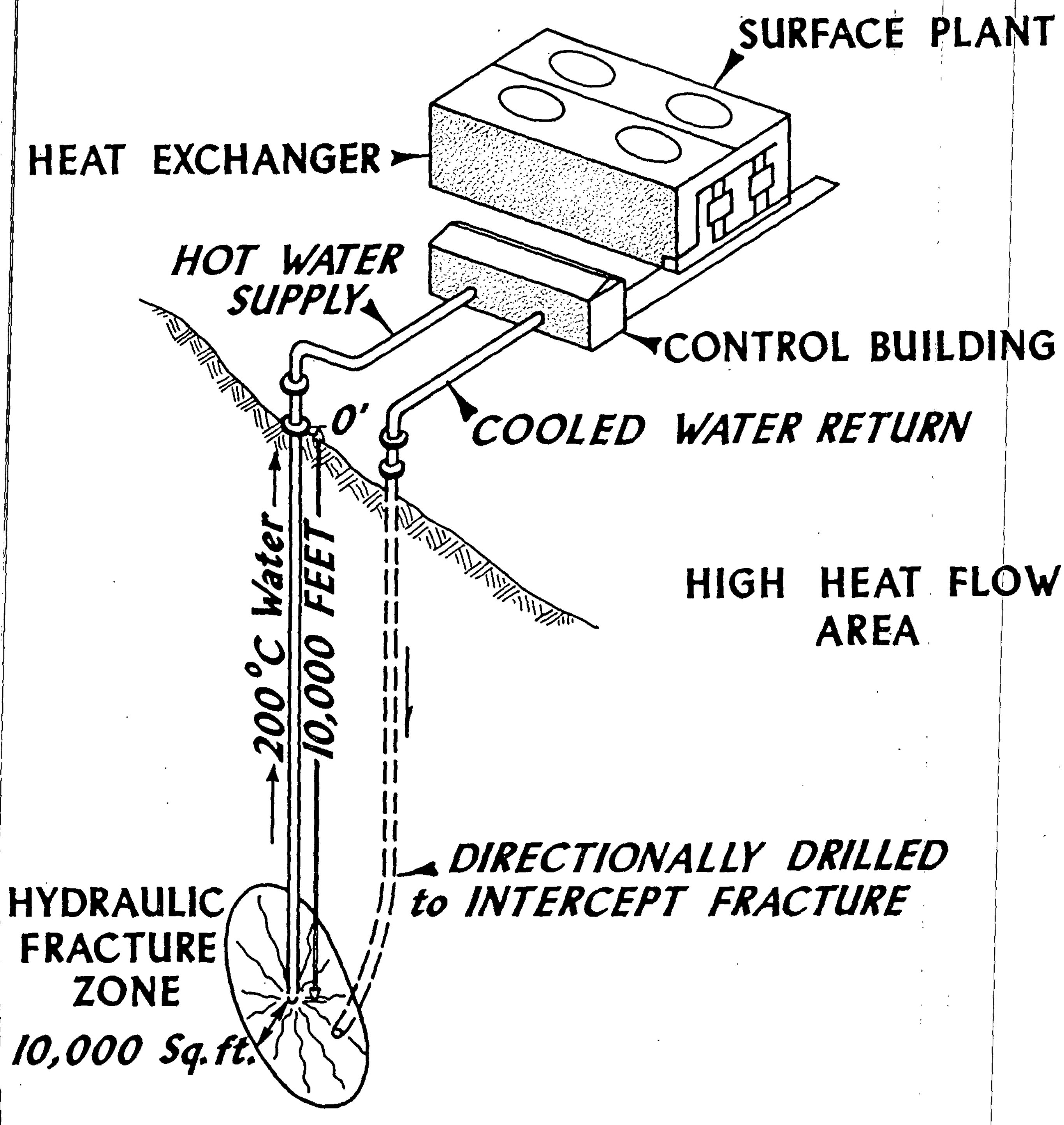


Figure 1.1.13

few pore spaces or fractures, and therefore contain little water. The feasibility and economics of extraction of heat for electrical power generation and direct uses from hot dry rocks is presently the subject of intensive research at the U.S. Department of Energy's Los Alamos National Laboratory in New Mexico (Smith et al., 1976; Tester and Albright, 1979). Their work indicates that it is technologically feasible to induce an artificial fracture system in hot, tight crystalline rocks at depths of about 3 km through hydraulic fracturing from a deep well. Water is pumped into a borehole under high pressure and is allowed access to the surrounding rock through a packed-off interval near the bottom. When the water pressure is raised sufficiently, the rock cracks to form a fracture system that usually consists of one or more vertical, planar fractures. After the fracture system is formed, its orientation and extent are mapped using geophysical techniques. A second borehole is sited and drilled in such a way that it intersects the fracture system. Water can then be circulated down the deeper hole, through the fracture system where it is heated, and up the shallower hole (Fig. 14). Fluids at temperatures of 150°C to 200°C have been produced in this way from boreholes at the Fenton Hill experimental site near the Valles Caldera, New Mexico. Much technology development remains to be done before this technique will be economically feasible.

Molten Rock Resource

Experiments are underway at the U.S. Department of Energy's Sandia National Laboratory in Albuquerque, New Mexico to learn how to extract heat energy directly from molten rock (2a of Table 1). These experiments have not indicated economic feasibility for this scheme in the near future. Techniques for drilling into molten rock and implanting heat exchangers or direct electrical converters remain to be developed.



HOT DRY ROCK GEOTHERMAL RESOURCE

Figure 1.1.14

Hydrothermal Fluids

The processes causing many of today's high temperature geothermal resources consist of convection of aqueous solutions around a cooling intrusion. These same basic processes have operated in the past to form many of the base and precious metal ore bodies being currently exploited, although ore forming processes differ in some aspects from hydrothermal convection processes as we understand them at present. The fluids involved in geothermal resources are complex chemically and often contain elements that cause scaling and corrosion of equipment or that can be environmentally damaging if released.

Geothermal fluids contain a wide variety and concentration of dissolved constituents. Simple chemical parameters often quoted to characterize geothermal fluids are total dissolved solids (tds) in parts per million (ppm) or milligrams per liter (mg/l) and pH. Values for tds range from a few hundred to more than 300,000 mg/l. Many resources in Utah, Nevada, and New Mexico contain about 6,000 mg/l tds, whereas a portion of the Imperial Valley, California resources are toward the high end of the range. Typical pH values range from moderately alkaline (8.5) to moderately acid (5.5). A pH of 7.0 is neutral at normal ground water temperature--neither acid nor alkaline. The dissolved solids are usually composed mainly of Na, Ca, K, Cl, SiO_2, SO_4, and HCO_3. Minor constituents include a wide range of elements with Hg, F, B, and a few others of environmental concern. Dissolved gases usually include CO_2, NH_4 and H_2S, the latter being a safety hazard. Effective means have been and are still being developed to handle the scaling, corrosion and environmental problems caused by dissolved constituents in geothermal fluids.

Conclusions

Although many types of geothermal resources exist, only some of these are presently economic. Vapor- and water-dominated resources (type 1) and sedimentary basin resources (type 3a) are presently most attractive for exploitation, while hot dry rock, magma, geopressured, and radiogenic resources are further from commercial development.

Geothermal resources are found in a wide variety of geologic environments and tectonic terrains. Those characteristics relevant to Spanish resources are discussed in later chapters.

IV. GEOTHERMAL RESOURCES OF SPAIN

The geology of the Iberian Peninsula is extremely complex, and reflects all the tectonic and lithologic variations associated with continental plate collision, extensive overthrusting and oceanic rifting. Tectonic activity continues to the present day and Quaternary volcanism is recorded both on the Iberian Peninsula and the Canary Islands. This complex geologic setting is favorable for the occurrence of geothermal resources of several types, and these have been identified and are currently being explored by the IGME.

The IGME has identified three basic resource occurrence types and seeks to identify a systematic exploration strategy for these resource types/occurrences:

I. Sedimentary Basins

II. Igneous Areas

III. Volcanic Areas

Our reading of the "Inventario General De Manifestaciones Geotermicas En El Territorio Nacional" has given us a basic understanding of the resource types. It is difficult to simultaneously categorize the identified resources by reservoir type, geology and location. Many resources have the characteristics of deep circulation along structures, irrespective of the host rock type.

Table 2 identifies the key elements for the various geothermal resources identified and described by the IGME. The multiplicity of lithologies and tectonic styles in some resource areas makes a simple classification difficult, and complicates the exploration of the resource.

Sedimentary Basins

Geothermal reservoirs may be present in the basal units of 1000-3000 m of detrital materials or in underlying dolomites and limestones. Permeability of the reservoirs is generally better where enhanced by fractures, i.e. along basin border faults. Thermal gradients of 3 to 5 °C/100 m have been documented for several of the favorable basins, indicating probable temperatures of 60°-100°C. Quaternary volcanism is associated with faulting near the margins of two basin areas, near Ciudad Real and the region of Olot-Gerona. Large reservoir volumes are possible in the basins.

Igneous Areas

Permeability and reservoir volume are almost exclusively limited to fractures in the igneous (and metamorphic) rock complexes of Galicia, and the Central, Extremadura, and Pirineos Cordillera. In some areas the fracturing is relatively minor and may not extend to great depth, thus only low enthalpy systems may exist. Deep circulation along faults may occur in other areas.

Volcanic Areas

The principal volcanic region of interest is the Canary Islands. Large reservoirs may occur in basalt flows, although major portions of some islands consist of diorite, gabbro or peridotite. Temperatures exceeding 100°C may be present at moderate depth. Quaternary volcanism has also occurred at Ciudad Real and the region of Olot-Gerona. In these areas, the volcanism suggests the possibility of a thermal source at shallow depths, but the reservoirs are most likely to occur in basin fill sediments.

The complex setting of many of the resource areas requires careful consideration in developing an exploration strategy. A generalized exploration strategy for the three basic resource types is presented and discussed in Chapter VIII.

TABLE 2

GEOLOGICAL CHARACTERISTICS OF GEOTHERMAL
RESOURCE AREAS IDENTIFIED BY IGME

RESOURCE AREA	GEOLOGIC CHARACTERISTICS							
	Sedimentary Basins	Igneous Areas	Volcanic Areas	Quaternary Volcanism	Precambrian/Paleozoic Intrusives	Paleozoic-Triassic-Jurassic Carbonates-reservoir	Cretaceous-Tertiary-Quat. Basin or Rift Fill	Overthrust Terrain
I. REGION DEL MACIZO CENTRAL O HESPERICO								
Galicia		X						
Ciudad Real	X		0	X			X	
Region Astur Leonesa	0	0				X		X
Cordillera Central y Extremadura		X			X			
II. CORDILLERAS CIRCUNDANTES Y CUENCAS ANEXAS								
Cordilleras Cantabrica y Vascocantabrica	X					X	X	X
Cordillera Iberica	X					X	X	X
Cuencas del Duero y del Tajo	X					0	X	
III. DEPRESSIONES EXTERNAS								
del Ebro y del Guadalquiver	X					X	X	X
IV. CORDILLERAS PERIFERICAS								
Pirineos	0	X			X	X	0	X
Cordillera Costera Catalona y region de Olot-Gerona	X	0	0	X	X		X	
Cordilleras Beticas y Baleares						X		X
V. ISLAS CANARIAS		0	X	X				

X = Primary Importance; 0 = Secondary Importance

TABLE 2 (cont.)

GEOLOGICAL CHARACTERISTICS OF GEOTHERMAL RESOURCE AREAS IDENTIFIED BY IGME

	Deep Circulation On Faults	Rift Tectonics	Fracture Permeability Dominant	Large Reservoir Volume	High Enthalpy	Deep Reservoir > 2000 m	Shallow Reservoir < 2000 m	Metamorphic Rocks
I. REGION DEL MACIZO CENTRAL O HESPERICO								
Galicia	0		X				X	X
Ciudad Real			X				X	X
Region Astur Leonesa	0		X				X	X
Cordillera Central y Extremadura	X		X				X	
II. CORDILLERAS CIRCUNDANTES Y CUENCAS ANEXAS								
Cordilleras Cantabrica y Vascocantabrica	X		X	0			X	
Cordillera Iberica	X	X		X	0	X		
Cuencas del Duero y del Tajo	X			X	0	X		
III. DEPRESSIONES EXTERNAS								
del Ebro y del Guadalquiver	X	X	0	X	0	X	X	
IV. CORDILLERAS PERIFERICAS								
Pirineos	X	X	X	?			X	
Cordillera Costera Catalona y region de Olot-Gerona	X	X	X	?	X	X	X	
Cordilleras Beticas y Baleares			X	X	0	X	X	
V. ISLAS CANARIAS			0	X	X	0	X	

X = Primary importance; 0 = Secondary importance

V. PHYSICAL PROPERTIES ASSOCIATED WITH GEOTHERMAL SYSTEMS

Geophysical exploration methods measure physical properties, or changes in physical properties, of the subsurface. Taken as a whole, the subsurface includes the rocks and their contained fluids. We are particularly interested in this section in the changes in physical properties that result from the presence of thermal fluids in a rock. These changes result mainly from the heat itself and also from chemical changes to the rocks associated with the thermal fluids.

It is generally true that the higher the temperature of the thermal fluids, the greater the changes in physical properties that may result. Higher-temperature fluids heat the rocks to a greater extent and, more importantly, are generally more reactive chemically. At the lower-temperature end of the scale, thermal fluids are much the same as normal groundwater, and may produce only small changes in physical properties of the subsurface. For this reason, low-temperature geothermal fluids (<100°C) may be very hard to detect at depth using geophysical techniques.

As we have said, the residence in or passage through a rock matrix of geothermal fluids may result in changes in physical or chemical properties of the bulk rock either as a result of properties the geothermal fluids may themselves possess or as a result of fluid-rock interaction. The chemical interaction process is often called "wall rock alteration" or "hydrothermal alteration", and may result in a substantial modification of the initial rock properties. High-enthalpy fluids, a reactive rock matrix, and a long period of fluid-rock interaction are generally required to effect changes extensive enough to affect surface geophysical measurements.

Density

Rock density depends upon mineral composition, degree of induration, porosity, and compressibility. Tables of typical rock densities can be found in any geophysics text, such as Dobrin (1976). Shales display marked variations of density with depth because of their relatively high compressibility. As a general rule, older sedimentary rocks are higher in density than younger sedimentary rocks. Most plutonic and metamorphic rocks display smaller ranges in density than do sedimentary and volcanic rocks. Acid igneous rocks are less dense than basic igneous rocks. Volcanic rocks often display rapid density variations due to porosity changes from place to place. Density variations greater than 25 percent of the average crustal density, 2.67 gm/cm^3, are rare in near-surface rocks. This observation is in sharp contrast to electrical and magnetic properties of rocks, which can vary over several orders of magnitude.

The precipitation of silica and carbonate minerals in sediments above moderate-temperature and high-temperature hydrothermal systems has been documented by several authors. A density increase of 0.2 to 0.4 g/cm^3 may result from partial deposition in a sediment with an initial porosity of 30 percent. Biehler (1971) has defined positive gravity anomalies in the Imperial Valley of California which are due to silica and carbonate deposition and to metamorphism of the native minerals to denser forms above and within hydrothermal systems. In igneous environments, there is generally less possibility for a major increase or decrease of bulk rock density due to fluid-rock interaction and available porosity. In addition, the presence of complex faulting or lithologic changes may result in a complex gravity field which would dominate or obscure the anomaly resulting from a density change due to secondary mineral deposition. A difference in density between an intrusion or an intrusive complex at depth, which may form a source of heat, and its host rock can sometimes make it possible to map the intrusion using gravity surveying.

Magnetic Susceptibility

The origin of magnetization in rock materials involves considerations on the atomic and molecular level, and is beyond the scope of this report. Rock magnetism is a complex topic whose details are still being studied. Strangway (1967a and b; 1970) and Doell and Cox (1967) give good summaries of this and related topics. Rock magnetism has also been treated in detail by Nagata (1961).

For our purposes there are three main points to note. First, magnetic minerals and rocks have a component of magnetization, often the chief component, due to induction in the earth's magnetic field. This induced component is the response of magnetic minerals to the earth's field, is proportional in intensity to the earth's field strength, and is in a direction parallel to the earth's field. The constant of proportionality is termed the <u>magnetic susceptibility</u>. Second, another form of magnetization called <u>remanent</u> or <u>permanent</u> magnetization often exists and is superimposed on induced magnetization. Remanent magnetization can form as a result of cooling of an igneous rock from a molten state, as a result of metamorphism, as a result of chemical changes, or from other causes. The remanent component of magnetization can be either weaker or stronger than the induced component, and it is often not in the same direction as the induced component. Remanent magnetism complicates interpretation. Rocks having small mineral grains commonly have a larger remanent component than those having larger mineral grains because the stability of remanent magnetization is related to grain size. Third, above a temperature known as the Curie temperature, magnetization changes and, for exploration purposes, rocks cease to be magnetic. The Curie temperature of pure magnetite is 580°C, but impurities can alter this value. This temperature is attained in the earth's crust at a nominal depth of 25 km, although the Curie point isotherm is believed to be much shallower in some areas such as areas of high heat flow and extensive geothermal activity. The majority of the anomalies seen on magnetic maps result from sources in the earth's crust because deeper rocks are above the Curie temperature and therefore do not contribute.

Only a few minerals are sufficiently magnetic to cause measurable changes in the earth's magnetic field. These are listed together with their magnetic susceptibility and ranges for the susceptibility of common rocks in Table 3. Magnetite is usually the magnetic mineral under consideration in exploration. It is both highly magnetic and widely distributed, principally as an accessory mineral. Empirical relations have been established between magnetite content and magnetic susceptibility of rocks (for example, see Mooney and Bleifuss, 1953). One commonly used rule of thumb is that 1 volume percent magnetite results in a magnetic susceptibility of about 3000×10^{-6} cgs, but this can be highly variable. If remanent magnetization is present and unrecognized, the magnetic susceptibility, and therefore magnetite content, interpreted from the anomaly can be too large or too small.

TABLE 3

MAGNETIC SUSCEPTIBILITY FOR COMMON MINERALS AND ROCKS

ROCK OR MINERAL	MAGNETIC SUSCEPTIBILITY $\times 10^6$ (cgs)	
	Approx. Range	Typical Range
Sedimentary Rocks	0-2,000	200
Acidic Igneous Rocks	600-6,000	2,500
Basic Igneous Rocks	1,000-20,000	5,000
Magnetite	300,000-800,000	500,000
Pyrrhotite	---	125,000

Most magnetic maps show lateral variations of magnetic susceptibility in rocks of the crust. Geologists who understand the meaning of magnetite distribution in particular areas can materially assist the geophysicist in interpretation.

As we have noted, the magnetization of most rocks results from the magnetic susceptibility of the mineral magnetite (Fe_3O_4), although remanent magnetization and susceptibility of other minerals may occassionally be more important for certain volcanic and sedimentary rocks. Hydrothermal alteration associated with geothermal fluids, particularly those fluids carrying large amounts of H_2S, can replace the magnetite and other iron minerals with a new assemblage dominated by pyrite which is only weakly magnetic. Thus, the original magnetism of the affected rock volume can be destroyed by interaction with hydrothermal fluids.

In sedimentary rocks of low initial magnetic susceptibility, i.e. 0-50 $X10^{-6}$ cgs, the effects of magnetite alteration, even if complete, would probably not be detectable by ground or airborne magnetic surveys.

Igneous (intrusive and volcanic) and metamorphic rocks often have magnetic susceptibilities in the range 1000-5000 $X10^{-6}$ cgs, and the destruction of magnetization (induced and permanent) by hydrothermal alteration can be complete. When the reacting fluids move along a single fracture in an otherwise "tight" igneous rock such as a granite, the effects of alteration and magnetite destruction may be limited to a zone less than a meter wide, and this may be recorded as a single "sharp" magnetic low by a ground magnetometer traverse. If the geothermal reservoir area is a large zone of fracturing near the intersection of major faults, extensive alteration of several cubic kilometers of rock volume may result, as has been observed in several porphyry copper deposits in the southwestern United States. The Coso, California geothermal system is typified by a large magnetic low on low level aeromagnetic data (Fox, 1978) and this corresponds to extensive wall rock alteration and complex facies changes in the granodiorite host rock (Hulen, 1978).

The production of extensive rock alteration and accompanying magnetite destruction is probably restricted to high enthalpy geothermal systems active for thousands of years, and may not be present in most Spanish geothermal areas.

Electrical Properties

Electrical Resistivity. Perhaps the most important physical property change due to the presence of a geothermal system, other than temperature and heat flow itself, is the change in electrical resistivity of the rock-fluid volume. Crustal rocks conduct electricity primarily via the movement of ions through pore water, although semiconduction in minerals such as sulfides and graphite sometimes contributes significantly. Ionic conduction in rocks increases with increasing porosity, increasing salinity, or increasing amounts of minerals exhibiting cation exchange. Higher temperature increases ionic mobility up to a certain point, and hence increases conductivity. Various geophysical surveys which respond to the electrical resistivity of the earth are used routinely and successfully in geothermal exploration. These techniques map regions of thermal brines and/or wall rock alteration resulting from the interaction of the thermal fluids with the reservoir rock.

At depths exceeding 5 to 15 km, mineral semiconduction dominates aqueous electrolytic conduction (Ward and Sill, 1984) and partial melts and magma become very conductive as compared to host rock. The magnetotelluric method offers one possibility for detection of high-level partial melts at these depths and may thereby lead to the discovery of areas of anomalous thermal gradient and blind geothermal systems.

Several publications discuss the details of fluid and rock resistivity in geothermal areas. Moskowitz and Norton (1977) provide an excellent physio-chemical discussion of the topic and numerical model results. A recent review paper by Ward and Sill (1984) provides an excellent summary of the topic. Excerpts of these papers are provided as Appendix I for those seeking an in-depth discussion of the topic.

The effects of temperature and dissolved ion content (related to the content of total dissolved solids, TDS) on fluid resistivity are very evident from a standard Schlumberger (1960) well log interpretation chart, Figure 15. From Figure 15 we may abstract the following fluid resistivity values of Table 4. With an assumed porosity, one can calculate expected earth apparent resistivities from Archie's law,

$$F = \frac{\rho_r}{\rho_w} = \phi^{-m}$$

where F is the formation factor, ρ_r is the resistivity of the rock, ρ_w is the resistivity of the saturating electrolyte, ϕ is porosity and m is the cementation factor which usually varies between 1.5 and 3.

TABLE 4

VARIATION OF BULK ROCK RESISTIVITY FOR AN ASSUMED 20% POROSITY

T (°C)	TDS NaCl (ppm)	ρ_w (ohm-m)	ϕ (%)	ρ_r* (ohm-m)
20°	500	13	20	325
20°	2000	2.9	20	72
60°	1000	2.7	20	68
60°	3000	1.1	20	28
100°	1000	1.8	20	45
100°	5000	0.48	20	12
100°	10,000	0.20	20	5

* m = 2

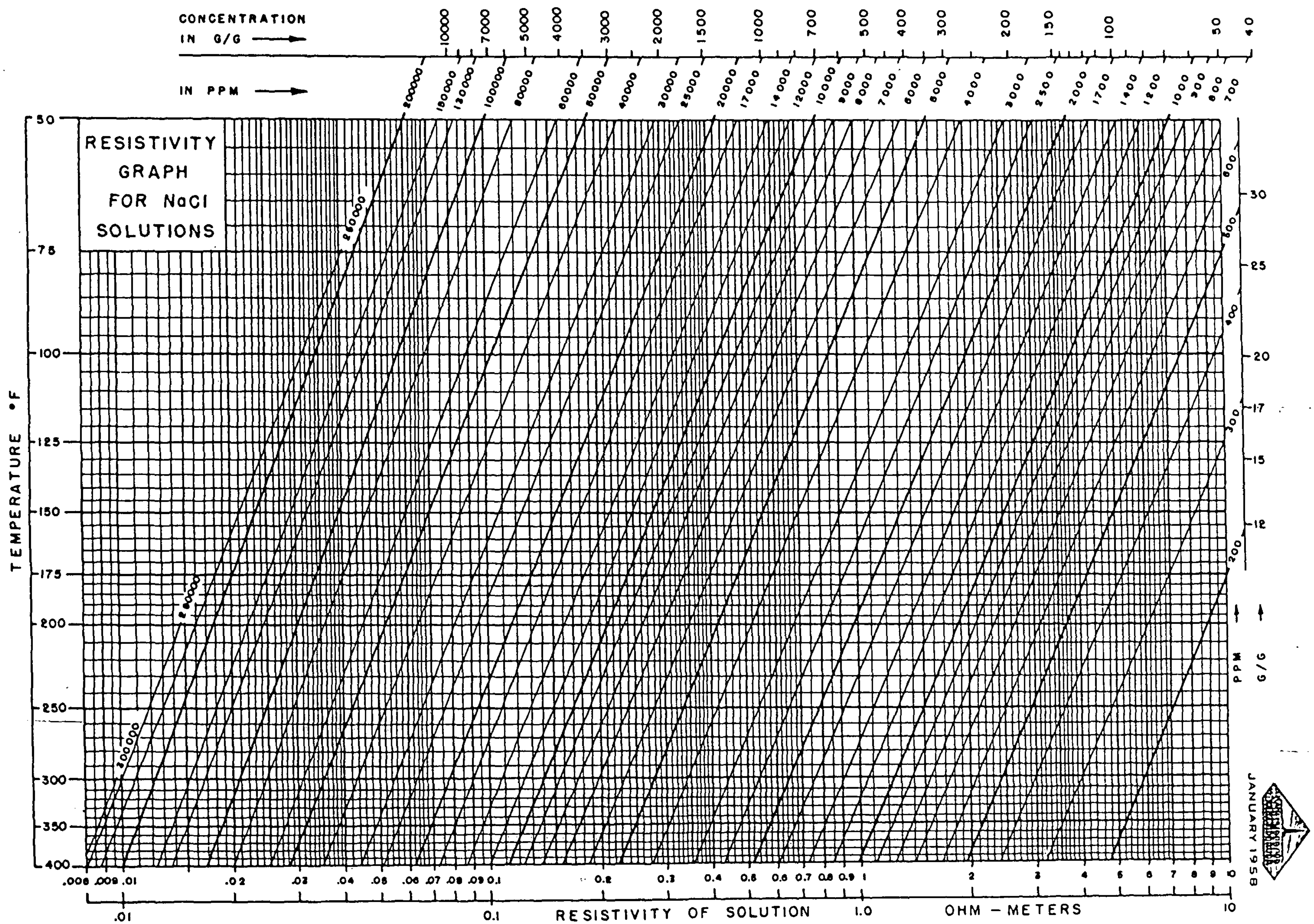

CONCENTRATION
IN G/G
IN PPM
RESISTIVITY
GRAPH
FOR NaCl
SOLUTIONS
TEMPERATURE °F
RESISTIVITY OF SOLUTION
OHM - METERS
PPM
G/G
JANUARY 1958

Figure 15.

Table 4 illustrates the variation of bulk rock resistivity, ρ_r, for an assumed 20% porosity and cementation factor, m = 2. The first case, T = 20°, TDS = 500 ppm corresponds to good water quality at background (ambient) temperatures and predicts a rock resistivity of 325 ohm-m. The 100°C, 10,000 ppm case predicts a ρ_r of 5 ohm-m. This is close to the in situ resistivity for more than 2 km^3 of earth observed at Cove Fort-Sulphurdale, Utah (Ross et al., 1982) and noted in other high temperature areas in the western United States. Table 2 suggests a likely range of reservoir resistivities of 70-10 ohm-m in Spain, 5 to 30 times lower than a likely background resistivity. The net volume of rock that would average 20% porosity is, of course, a function of the local geology.

In areas where extensive hydrothermal alteration has taken place, clay and zeolite minerals may line fractures along which fluids flow. These minerals have a tendency to increase the conductivity of the rocks, and in such areas Archie's Law does not apply. The bulk rock resistivity will be lower and may be much lower than Archie's Law would predict.

Temperature and Thermal Conductivity

Temperature is the fundamental property exploited from a geothermal resource and its measurement is relatively straightforward. The interpretation of temperature measurements and the evaluation of resource potential from these measurements can be complex. The interpretation of temperature, thermal gradient, and heat flow data is discussed in detail by several authors to which the reader is referred (Lachenbruch, 1978; Sass et al., 1971; Chapman and Pollack, 1977; Sass et al., 1981; Ryback and Muffler, 1981). Cathles (1977) presents an analysis of the cooling of intrusives by ground water convection. His model suggests that elevated rock temperatures (> 100°C) may be present within a few km of medium sized intrusive bodies as much as 200,000 years after emplacement.

In general, the thermal conductivity of rocks spans little more than one order of magnitude. Rocks are classified broadly in terms of their mineralogical and structural characteristics. Within a particular rock type there may be a sufficient variation of these characteristics to give very different thermal properties; for example, in granite the thermal conductivity value depends quite strongly upon the quartz content but, nevertheless, other factors may have a strong influence, producing much scatter about a linear trend when quartz content is plotted against conductivity.

For porous rocks, the thermal conductivity may be strongly dependent upon the conductivity of the pore fluid; it may sometimes be necessary to calculate thermal conductivity from a knowledge of the rock matrix, porosity, and nature of the pore fluid (Roy et al., 1981).

Among the various parameters influencing the thermal properties of rocks, porosity is perhaps the most important. Because of this, much of the original literature data has to be rejected because the data on conductivity values are given with no reference to the porosity of the rocks. Roy et al. (1981) give data to show that thermal conductivity decreases as much as a factor of 2.5 as porosity increases from 6 to 32 percent for a selected rock sample. Typical thermal conductivities are 2.0 to 3.4 W/m°K for diorites, 2.0 to 4.6 W/m°K for granites (centering around 3.0), have roughly the same range for limestone as for granite, and extend up to 7.5 W/m°K for quartzites. The effects of hydrothermal alteration on thermal conductivity are only poorly documented, but probably follow inversely changes in porosity by this process. Densification of rocks through deposition of minerals would be expected to increase thermal conductivity.

Seismic Velocity

The effect of pressure on velocity has been studied extensively and is well understood. In low-porosity rocks, cracks that are open at low confining pressure, close with pressure to yield large increases in both V_p and V_s, increases that are anomalous with respect to increases expected from mineral constituents alone.

In porous rocks, pore collapse and crack closure have the same effect although generally the effect of pore collapse takes place over a wider pressure range (Wyllie et al., 1958; King, 1966) since pores are in general stronger than cracks (Walsh, 1965). Other studies have been made specifically on the effect of porosity on velocity (Wyllie et al., 1958; Warren, 1969).

The effect of temperature on velocity in rocks is much more variable and less well understood. In general, velocity decreases with increasing temperature, probably mostly because of the expansion of existing cracks and the propagation of new cracks because of thermal stress. Few measurements have been made under controlled conditions, but thermal cycling and thermal gradients have been identified as producing cracks in rocks (Richter and Simmons, 1974).

Credible seismic attenuation data are not abundant since measurements are inherently difficult. Values depend not only on the environmental parameters (pressure, temperature, etc.) but also in general on the parameters of the measurement such as amplitude and frequency of excitation. Gregory (1977) and Toksöz et al. (1979), Winkler (1979), Winkler and Nur (1979; 1982), and Tittmann et al. (1979) presented the most recent relevant data.

The seismic velocity of melts is appreciably lower than for solid rock, and, of course, melts do not propagate shear waves. These facts have been used by the U. S. Geological Survey to detect the presence of magma beneath the Geysers, California.

Other Properties

Other properties which indicate the presence of a geothermal system include natural seismicity (seismic noise and seismic emissions, microearthquakes), changes in seismic wave propagation, and fluid flow. These properties are less well defined and do not generally play a major role in the exploration for geothermal resources, and will not be discussed in detail here. A limited discussion relating to these geophysical techniques follows in Chapter VI.

VI. GEOPHYSICAL METHODS APPLIED TO GEOTHERMAL EXPLORATION

Introduction

The role of geophysics in geothermal exploration, as in petroleum and mineral exploration, is to help select from a large region several much smaller areas that have the highest potential for occurrence of a resource. Detailed geophysical surveys within these smaller areas then attempt to define the optimum target for drill testing, which is the most costly but also the most certain method of evaluation of resource potential. The application of most geophysical methods has evolved with relatively little change from the petroleum and mining industries, but also includes some new methods (thermal, passive seismic) and perhaps a new emphasis on other methods (magnetotelluric; electrical resistivity, self-potential). In addition, a high level of integration with geologic and geochemical studies is required for the successful exploration program.

The geophysical methods may be categorized as regional or detailed and may have the role of subsurface geologic mapping or direct (or indirect) detection of the geothermal resource. Other basic considerations to the utility of a given technique are the geologic setting and/or resource type, and the depth of occurrence of the intended resource. These considerations are inherent to our discussion and critique of methods in this section.

Interpretation of Geophysical Data

Interpretation of any type of geophysical data is essentially a two-step process. The first step is accomplished by estimating the parameters of simplified models of the earth, i.e. the use of the observed data to form a picture of the vertical and lateral variations in the subsurface of the physical property being measured. In this step, the geophysicist uses various interpretation aids such as curve matching or computer modeling to help construct a model of the physical property variations. The second step in the interpretation of geophysical data is the process of interpreting the subsurface geophysical model of step one in terms of the local geothermal geology and hydrology -- rock type variations, positions of faults, depth to water, location of thermal water, etc. In this step the geophysicist and geologist must work closely together to assure that the most accurate picture of the subsurface evolves.

Geophysical Modeling. Two ingredients are essential for success in this task: (1) a geophysicist experienced in interpretation of the types of geophysical data being collected, and (2) availability of interpretation aids for the method being used. Interpretation aids include computed catalogs of various subsurface models and computer programs for computing any particular model desired.

The earth is far too complex for its geophysical responses to be evaluated exactly, but the simplified model, if close enough to reality, can be of considerable help in interpreting geophysical data. In discussing the possible kinds of physical structure in the subsurface, the geophysicist usually speaks in terms of 1-dimensional (1-D), 2-dimensional (2-D) or 3-dimensional (3-D) models or a combination. This is true of gravity, magnetic, seismic, heat flow, electrical, and other models of the subsurface. The concepts embodied in the various models of the subsurface are illustrated in Figure 16. When the resistivity varies only with depth, z, and is horizontally uniform at a given depth, the earth is said to be "layered" or "1-dimensional". We speak of layered-earth or 1-D models for this type of structure. Examples of areas where such a model might be appropriate include large sedimentary basins where petroleum and geothermal resources are sometimes found. In these areas, rock units are often horizontal, continuous

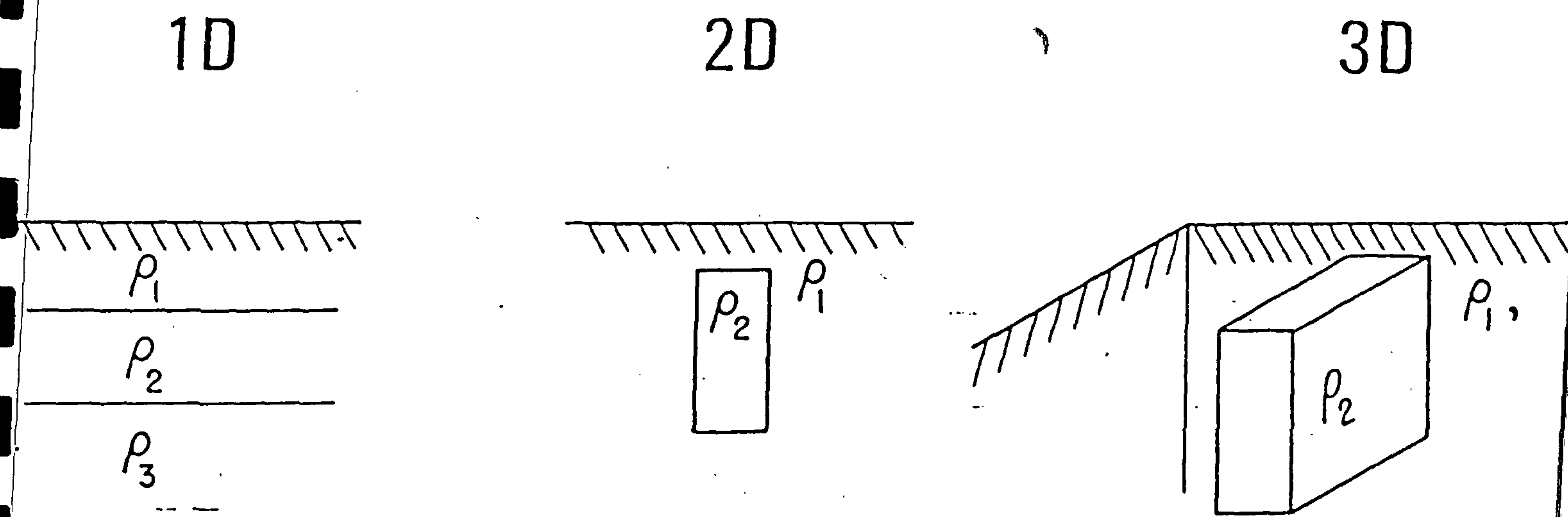


Figure 16. Various Physical Models of the Subsurface. The parameters, ρ, give values for physical properties of interest in exploration such as density (gravity surveying), magnetic susceptibility (magnetic surveying), seismic velocity (seismic surveying) or electrical resistivity (MT/AMT, EM and galvanic resistivity surveying).

and uniform over many miles, but rock type (and therefore resistivity) varies with depth.

In a 2-dimensional structural setting, the physical property varies downward and also in one horizontal direction, but is uniform in the other horizontal direction. Such structure is sometimes seen, for example, in the Basin and Range geologic province of western North America, where long continuous valleys are separated by long north-trending mountain ranges. For purposes of geophysical interpretation, a 2-D model of the subsurface may be adequate if certain criteria are met. For example, if the length of a body is more than about 8 to 10 times its burial depth to the top, then use of a 2-D model to calculate the anomaly is adequate for interpreting a surface resistivity survey. Each type of geophysical survey has similar relationships.

In a 3-dimensional earth, the physical property being studied varies in depth as well as in both horizontal directions. Such models are the most general in terms of applicability to geothermal exploration and reservoir definition because geothermal systems themselves are usually three-dimensional.

These illustrations of 1-D, 2-D and 3-D geophysical models are only for understanding of the concept. In actual application, a 2-D or 3-D model is usually built up of a number of 2-D or 3-D blocks that approximate the size and shape of the model whose response is being calculated. Because most geothermal areas are geologically complex, the 3-D model is generally the most applicable. However, 3-D models are usually more difficult to implement on a computer, are more difficult to use, and therefore are approximated by 2-D models where possible.

Geophysical interpretation methods can be divided into four classes: 1) rule-of-thumb, 2) characteristic curve matching, 3) forward modeling, and 4) inverse modeling. Progress in development of techniques in each class has led to better interpretation, especially since the advent of the digital computer. Rules of thumb can be used to get a preliminary overview of location and depth of anomalous bodies before more sophisticated techniques are applied. Many curve matching techniques are available, generally for interpretation in terms of specific bodies or models (Grant and West, 1965). These techniques are pursued if no computer modeling capability is available or if only a few profiles or anomalies are to be interpreted.

In more complex situations, forward computer modeling is beneficial. In forward modeling, a preliminary estimate (i.e., a model of the subsurface configuration of anomalous physical property) is formed, perhaps by application of rules of thumb. Then, the anomalies to be expected are calculated from the model. The calculated results are compared with the observed anomalies, and the model is modified to start the cycle again. This iterative process is continued until a satisfactory match between computed and observed results is obtained. Any geologic control available can be used to constrain the model so that the results, while not unambiguous (see section below on ambiguity) are geologically sound. Computer graphics and user-interactive programs facilitate this approach greatly. At the present time, comprehensive 2-D and 3-D computer programs are available for many of the common geophysical techniques.

In the inverse approach, sophisticated mathematical techniques are used to calculate a model directly from the data. Inversion does not yield a unique model either, however. The promise that inversion offers is for rapid and inexpensive interpretation of large amounts of data by letting the computer do most of the work. The challenge is to assure appropriate model constraints and to allow input of geologic knowledge so that the final result is geologically sound. Techniques for 2-D inversion of magnetic data have been developed and successfully applied by Hartman et al. (1971) and by O'Brien (1971, 1972). Such modeling is currently at the forefront of development. These techniques are more reliably applied to rather simple geologic situations such as basement studies for petroleum exploration. Interpretation in more complex geologic environments still relies heavily on experience in spite of increases in the level of sophistication of interpretational aids.

Ambiguity in Geophysical Investigation

No interpretation of geophysical data alone is unique. Generally, it can be said that many different subsurface models could be devised to explain a certain set of data equally well. This is illustrated for the case of gravity data by Figure 17. This figure shows a gravity profile that is presumably caused by topography on the contact of less dense rocks above with denser rocks below. It can be shown that each of the basement profiles, numbered 1 through 7, explains the observed anomaly equally well. There is nothing in the geophysical data alone that would lead one to choose one basement profile over the others. Of course, if other subsurface information is known, the choices of basement profile may be limited, with the ones not geologically plausible being eliminated. In fact, **the key to reducing ambiguity in geophysical interpretation is integrated interpretation of all data in the area --** geological data, geochemical data, other geophysical data and hydrological data. If an interpretation can be found that agrees with all available data sets in an optimum way, then this is usually considered to be the best interpretation.

It is important to note that the ambiguity that arises in geophysical

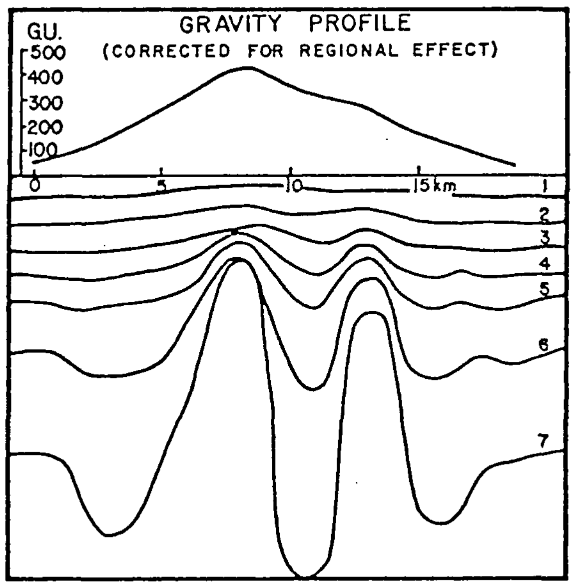


Figure 17. Ambiguity in gravity data interpretation.

interpretation is not the result of problems with surveys or interpretation techniques. Even if a perfect set of data were available, there would be ambiguity in its interpretation. No amount of research into new techniques, etc., will eliminate the problem, since it arises basically in the fact that there are many more unknown variables in the way physical properties vary in the earth than can be determined from a single set of geophysical data alone, i.e. there are more unknowns that there are relations among the unknowns. This is inherent in the situation and cannot be eliminated.

Noise in Geophysical Surveys

All geophysical surveys are subject to a variety of noise sources, each of which tends to degrade the quality of the data and of the resulting interpretation to a greater or lesser extent. It is important for the geophysicist to understand the various noise sources operating in the particular area and for the particular technique in use and to attempt to minimize these noise sources. Noise sources can be broadly classified as: (1) geologic noise, (2) instrumental noise, (3) cultural noise, (4) environmental noise, (5) natural field noise, and (6) topographic noise.

Geologic Noise. This term refers to geophysical responses from bodies or zones that are not of interest and that interfere with the target response. For example, an upper layer that is highly variable in resistivity will introduce a larger variation in data values, even for deep resistivity soundings, than if the variable layer were not present. A second example might be a low-resistivity zone at depth that is due to shale and that masks a low-resistivity response coming from a nearby geothermal system.

Noise in the Survey System. No measurement is completely precise -- all have a certain precision that is set by the measuring system. For example, although modern gravity meters can easily detect changes in gravity as small as 0.001 mgals, survey results are rarely accurate to this figure because of variations in instrument operation, haste in reading, failure to make adequate base ties to properly account for drift and tidal variations, inaccuracy in elevation determination, and other causes. The geophysicist should consider separately each component of the measuring system in attempting to minimize this type of noise.

Cultural Noise. This term is used when man-made causes interfere in survey precision. For example, it would be unwise to do a resistivity survey in the middle of a town because of the electrical noise introduced into grounded objects by the electrical utility power system and because the presence of many grounded conductors such as phone lines, fences, and water pipes redistributes current flow in the ground and causes false resistivity readings. A second example would be the seismic noise generated by a road or town that may preclude using passive or even active seismic techniques nearby.

Environmental Noise. This refers to such effects as wind or the sea surf on seismic geophones or on MT coils. Rain may cause noise in self-potential surveys.

Natural Field Noise. Variations in natural earth fields may disturb survey results. For example, magnetic surveying is difficult or impossible during a magnetic storm, and should probably be discontinued. Also, the natural electromagnetic fields that are used signal sources in MT and AMT surveying become noise sources for resistivity and SP surveying.

Topographic Noise. Topography can introduce unwanted effects into survey results. In some cases, topographic corrections can quite easily be made (gravity surveys, for example) whereas in other cases correction may be difficult or impossible (MT surveys, for example). One should be at least aware of the potential adverse effects of topography and design the survey to minimize them to the extent possible.

Thermal Methods

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

Thermal Gradient and Heat Flow. Thermal gradient and heat flow surveys provide basic data about subsurface temperatures. Sass et al. (1981) and Wilson and Chapman (1980) present detailed discussions of the method. Drill holes should be deep enough to penetrate the near-surface hydrologic regime dominated by meteoric recharge and cold water overflow. In high rainfall areas, this zone may exceed 700 m in thickness. In some sedimentary basins and crystalline rock environments depths of 30 to 100 m may be adequate. Some program of several shallow and a few deep thermal gradient holes is applied in most of the systematic geothermal exploration programs throughout the world.

The (vertical) heat flow is given as:

$$q = K(z) \frac{dT}{dz} \text{ (milliwatts/M}^2 \text{ or } \mu \text{ cal/cm}^2\text{-sec)},$$

where k = thermal conductivity (W/M-°C) or (mcal/cm-sec-°C)

T = temperature (°C)

and z = the vertical coordinate in meters.

The quantity dT/dz is, of course, the geothermal gradient, and in practice it is approximated by measuring temperature down a borehole and forming ratios $\Delta T/\Delta Z = \frac{T_{z2} - T_{z1}}{z_2 - z_1}$ for various depth intervals. A typical value for the geothermal gradient is 30°C/km, or 0.03°C/m. Apparatus to measure the geothermal gradient usually consists of a thermometer probe capable of measuring temperature differences of about 0.01 C° and several hundred to several thousand meters of cable. Small units for shallow holes can be highly portable whereas more sophisticated, deep-hole units must be truck mounted. Temperature logging is quick and relatively inexpensive.

The thermal conductivity, K, must be measured on rock samples in the laboratory as there is no suitable down-hole probe. It can be shown that

K = kpc, where

k = thermal diffusivity,

p = density,

c = specific heat.

Although down-hole probes have been constructed to measure thermal diffusivity, k, there is significant variation in both p and c as a function of rock composition, and samples are required in any case upon which to measure the latter quantities. This need for samples of subsurface rocks exists in application of many geological, geochemical, geophysical and engineering techniques.

There are two obvious uses of data such as these. First, a measure of temperature can provide directly an indication of anomalous heat and of existence, therefore, of a geothermal resource. Second, if the heat flow is anomalous, this provides an indication of possible geothermal activity.

Extrapolation of Temperature Profiles. One is always tempted to extend the temperature profile obtained in a drill hole beyond the bottom of the hole by straight-line extrapolation. This is obviously dangerous. In the first place, the heat flow is the quantity that will remain constant with depth, assuming that there are not sources or sinks for heat in the rocks, which is usually equivalent to assuming that groundwater flow does not disrupt the heat flow pattern. Therefore, the thermal gradient will vary inversely with thermal conductivity. If representative values of thermal conductivity can be determined for all of the rock types in the stratigraphic section below the well, then these may be used to perform a more meaningful extrapolation of temperature with depth. Of course, this process can still be highly unreliable.

Temperature-depth profiles that show a maximum temperature and then a negative or reversed gradient with cooler temperatures below are quite common in geothermal areas, especially on the outer fringes of an area where thermal waters may flow long distances laterally along at certain depths. It should be noted, too, that temperatures above the boiling point versus depth curve will usually not be observed, and this effect will cause the temperature gradient to diminish with depth in high-temperature resource areas.

Reconnaissance Data. In addition to basic geologic and geochemical data, regional heat flow values may provide an indication of resource potential and grade. 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).

Snow melt photography and thermal infrared imagery are two other temperature sensitive methods which, although not generally considered to be geophysics, per se, have been used in reconnaissance geothermal exploration. Snow melt photography has been used at Coso Hot Springs, California and Yellowstone National Park, to indicate surface areas of even slightly elevated temperatures at low survey costs. Color aerial photographs of these areas were made hours to days after light to moderate snowfall. The thermally anomalous areas were visible because the snow melted faster over these areas than over non-thermal areas. The successful use of snow melt photography requires extreme flexibility in survey scheduling and a temperate climate. One must take the pictures at the optimum time after the snowfall.

Airborne thermal infrared surveys have been used to map the occurrence of warm ground and hot springs on land (Kenya) and hot springs along the coastline of volcanic islands (Hawaii). Large areas can be mapped at reasonable unit costs but mobilization charges for the survey crew may be substantial.

This method finds little application except in areas of very poor access or thick ground cover. It is not generally used except experimentally in most exploration programs because it lacks the sensitivity that would be necessary to locate any but the most obvious surface thermal features. The surface temperature is affected to a much greater degree by such variables as exposure to sun, slope angles and directions, nature of surface rocks and soils, amount and nature of surface vegetation, and hydrology than it is by subsurface heat flow.

Detailed Surveys. 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 low cost method to determine near-surface temperatures is a shallow-temperature survey. With a hand-held or truck-mounted power auger drill 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 air temperature 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 and Lewis, 1983) 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 the United States industry of this technique (Ward et al., 1981).

Limitations and Costs. Although thermal methods are the only direct methods of detecting geothermal resources, they have some limitations. Among these are:

1. Cost Per Data Point. Because drilling is expensive, application of any of the thermal methods requiring a hole becomes expensive. It is not possible to give specific costs for thermal gradient or heat flow studies due to the great variability in drilling cost. In the U.S., temperature gradient holes of 10 cm to 20 cm diameter are often rotary drilled to depths of 100 m to 300 m for costs ranging between U.S. $30

and U.S. $80 per meter. Temperature logging can be done in such shallow holes using a backpack portable system with sensitivity of 0.01°C costing of the order of U.S. $5000. If deeper thermal gradient holes are needed to get below zones of active groundwater circulation, costs can increase quickly.

2. Hydrologic Problems. Perhaps the biggest problem with application of the thermal methods is lateral movement of ground water in shallow aquifers. In some areas, shallow aquifers tens to hundreds of meters deep may carry large quantities of meteoric water which sweep away any anomalous amount of heat coming from depth and completely obliterate a high heat flow or thermal gradient pattern over the resource. It is imperative that one understands the hydrology of the exploration area in order for thermal methods to be used reliably.

3. Lack of Thermal Equilibrium. The drilling process disturbs the thermal equilibrium around a borehole. One must wait a period of days to months in order for the hole to recover thermal equilibrium. Considerations of this kind have been discussed by Lachenbruch (1978) and by Jaeger (1965), among many others.

Electrical Methods

Most electrical geophysical methods are based on measurement of the electrical conductivity (or its reciprocal, the resistivity) of the earth. Measurements made at the surface can be interpreted in terms of lateral and vertical variations of resistivity within the earth, and under appropriate conditions geothermal resources and/or the structures with which they are associated can be detected.

Thermal waters become increasingly conductive with increasing salinity (dissolved solids) and with increasing temperature up to 300°C, above which conductivity decreases. In addition, 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.

All electrical geophysical methods involve the measurement of an <u>impedance</u>, with subsequent interpretation in terms of the subsurface electrical properties and, in turn, the subsurface geology. Basically an impedance is the ratio of the response (output) to the excitation (input). In resistivity the input is a current injected into the ground between two electrodes, while the output is a voltage measured between two other electrodes. In electromagnetics (EM) the input might be a current through a coil of wire and the output is the voltage induced in another coil of wire.

In <u>frequency domain</u> impedance measurements, the input current is a sine wave at a particular frequency. The output also is a sine wave whose amplitude (A) and phase (ϕ) depend upon electrical properties of the earth. The frequency (f) of the sine wave is the inverse of the period (T). Often it is convenient to decompose the output wave into <u>in-phase</u> (real) and <u>quadrature</u> (imaginary) components.

Impedance also can be measured in the <u>time domain</u>, in which case the current is periodically turned on and off. The output is the voltage measured at various times when the transmitter current is off. Note that the input again is periodic, because measurements must be made for each of several periods and then added together, or stacked, to eliminate noise. Time and frequency domain measurements are directly related through the Fourier transform, and in that sense, are equivalent. However, in practice, each system has advantages and disadvantages.

There are three basic modes of operation for any electrical method: (1) sounding, (2) profiling, and (3) sounding-profiling. In sounding, the transmitter-receiver separation is changed, or the frequency is changed, and the results are interpreted in terms of a layered earth, i.e. the depth to the top and the resistivity of multiple layers may be determined. If the earth is truly layered, this method may be applicable but because the earth may not be layered in geothermal prospecting, sounding must be used with caution. For example, crossing a contact between rocks of differing resistivities as the electrodes are expanded can affect the data and there is generally no good way to connect the data so that a viable interpretation can be made. In profiling, the transmitter or receiver, or both, are moved along the earth to detect lateral anomalies. However, in pure profiling no depth information to anomalous bodies is generated, and variation in thickness of layers can be interpreted incorrectly as lateral boundaries.

The most useful method is a combination of sounding and profiling, which delineates structures with both lateral and vertical variations. As examples of the above, the Schlumberger array is often used to make vertical electrical soundings (VES). This method should be used only in areas where one knows that the resistivity structure is layered. A better resistivity array for general use is the dipole-dipole array, because in its usual method of field deployment, one obtains <u>both</u> soundings and profiles, and the two-dimensional variations of resistivity are therefore observed in such a fashion as to be easily interpreted.

Electrical methods have become more useful in recent years through advances in both interpretation and instrumentation. Modern field instruments are based on micro-computers. Processing the signals digitally greatly increases the accuracy and, in fact, makes possible new types of measurements. Further, data reduction in the field results in more reliable results and more cost effective surveys.

Hohmann and Ward (1981) have recently reviewed the applications of electrical methods in mining exploration, and many of the points made in this important article are also applicable to geothermal exploration. Other authors are cited in the bibliography to the report.

It is not possible in a report of this nature to discuss the many electrical methods individually because of their great number. In what follows, we will group our discussions into galvanic techniques, electromagnetic techniques and other techniques.

<u>Galvanic Electrical Resistivity Methods</u>. Galvanic methods use grounded electrodes to introduce electrical currents directly into the ground and to measure the resulting voltage. Electrical resistivity data are routinely acquired in geothermal exploration on the detailed, site-specific scale and, less frequently, in regional or reconnaissance exploration.

The resistivity and induced polarization methods (discussed below) are based on the response of earth materials to the flow of current at low frequencies. Strictly speaking the resistivity method is based on <u>potential theory</u> which requires direct current, i.e. zero frequency, but noise and measurement problems quickly lead to the use of alternating currents of low frequency. The induced polarization method, on the other hand, requires the use of alternating current, because it is based on changes in resistivity as a function of frequency. As the frequency increases to some critical frequency, f_c, determined by the resistivity (ρ) of the materials and the scale size L of the measurement, electromagnetic coupling between transmitting and receiving circuits violates potential theory so that electromagnetic theory is required.

For low frequencies where potential theory is applicable the voltage (V) produced by a point source of current (I) on a homogenous half-space of resistivity ρ is

$$V = \frac{\rho I}{2\pi r},$$

where r is the distance from the point current source. For a given voltage and current measurement, this equation can be solved for the resistivity. In actual practice, current is introduced through a pair of electrodes, and the voltage difference (ΔV) is measured between another pair. For a homogeneous earth the resistivity is given by

$$\rho = K \frac{\Delta V}{I},$$

where K is a geometric factor, which depends on the electrode configuration. When the ground is not homogeneous, the voltage and current data are still reduced using the above equation, but the resistivity is called the apparent resistivity ρ_a. It is the resistivity a homogeneous earth would have to produce the same measurement.

When polarizable materials are present, the voltage will have a component in quadrature with the transmitter current. The apparent resistivity is then complex and can be represented by its real, or in phase, and imaginary, or quadrature, components or by its magnitude and phase angle.

Schlumberger soundings (vertical electrical soundings (VES)) are often measured at many scattered sites within a large region, and depth to a given conductive horizon is contoured from these data. Although the method is efficient for regional data acquisition, its interpretation usually assumes a layered earth model, which may or may not be true. This assumption of onedimensional environments must be evaluated, particularly as current and potential electrodes expand across structures or other lateral resistivity contrasts in complex geologic environments. The results are generally valid for basin exploration but may be misleading for a regional assessment in complex, non-layered geologic terrains. The basic geometry of the Schlumberger array is compared with other popular arrays in Figure 18.

Another reconnaissance resistivity technique uses the bipole-dipole array, which permits the most flexibility in deployment of the transmitter dipole (and hence electrodes) and the selection of receiver sites. The bipole-dipole method permits a rapid mapping of the areal distribution at the expense of resolution. It has been widely used in geothermal exploration (i.e. Keller et al., 1975; Stanley et al., 1976) even though the contoured apparent resistivity patterns are complex and difficult to interpret. Keller et al. (1977) used this method effectively in the reconnaissance exploration for geothermal resources on the East Rift Zone of Kilauea Volcano, Hawaii Island.

Figure 19a illustrates the bipole-dipole array geometry and parameters as used in surveys conducted by UURI. A transmitter dipole length of 610 meters is usually chosen to provide adequate current penetration to depths of 600 to 1200 meters for receiver sites located from 600 to 3000 meters from this dipole. The resultant voltages were measured with two orthogonal 152 m dipoles. The total-field apparent resistivity is computed from the expression

$$\rho_a = [(V_1^2 + V_2^2)]^{1/2} \frac{Q}{I}$$

where V_1 and V_2 are the observed (orthogonal) voltages, I is the transmitted current, and Q is the geometric factor for the standardized dipole lengths and variable transmitter-receiver positions (Hohmann and Jiracek, 1979; Frangos

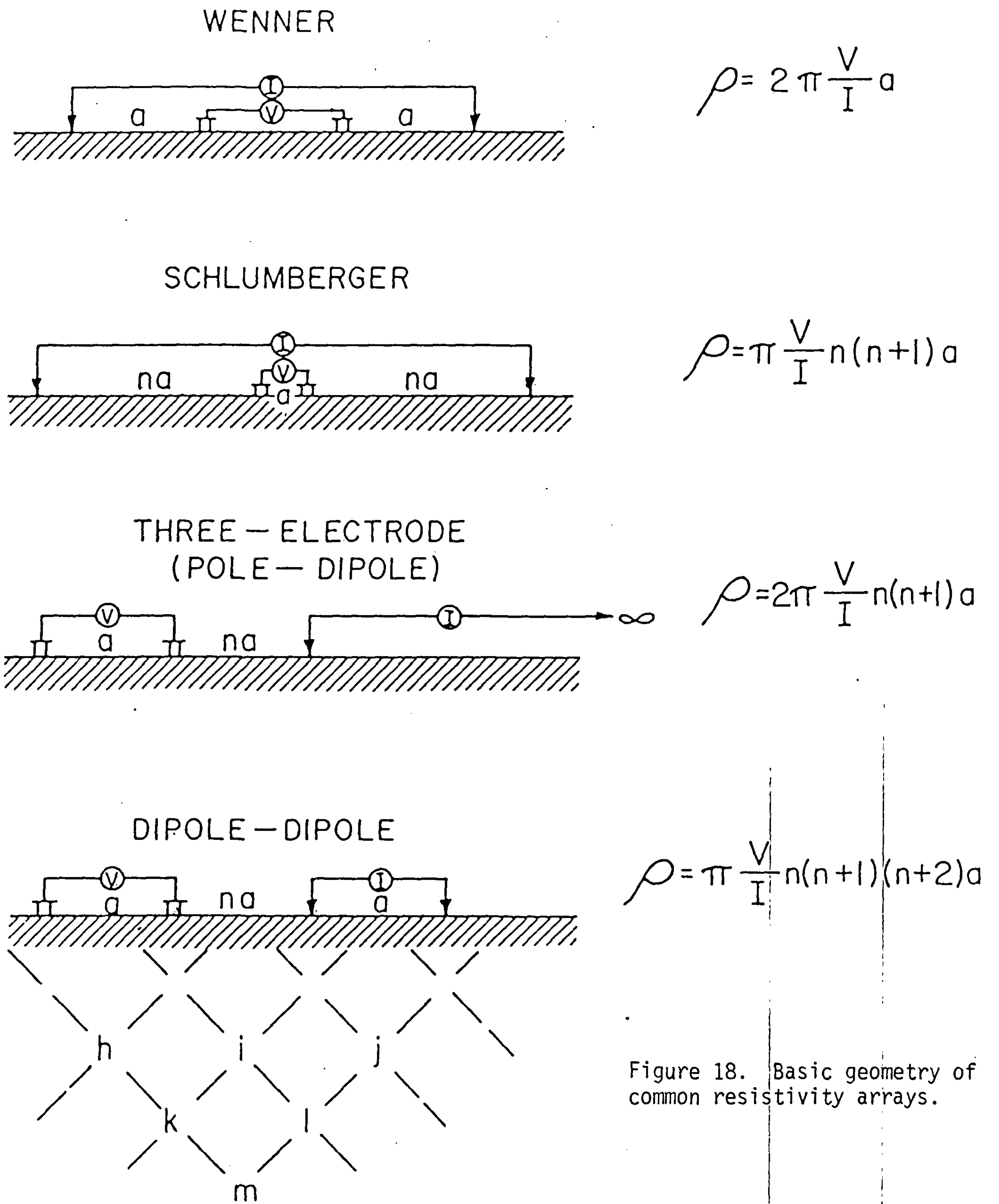


Figure 18. Basic geometry of common resistivity arrays.

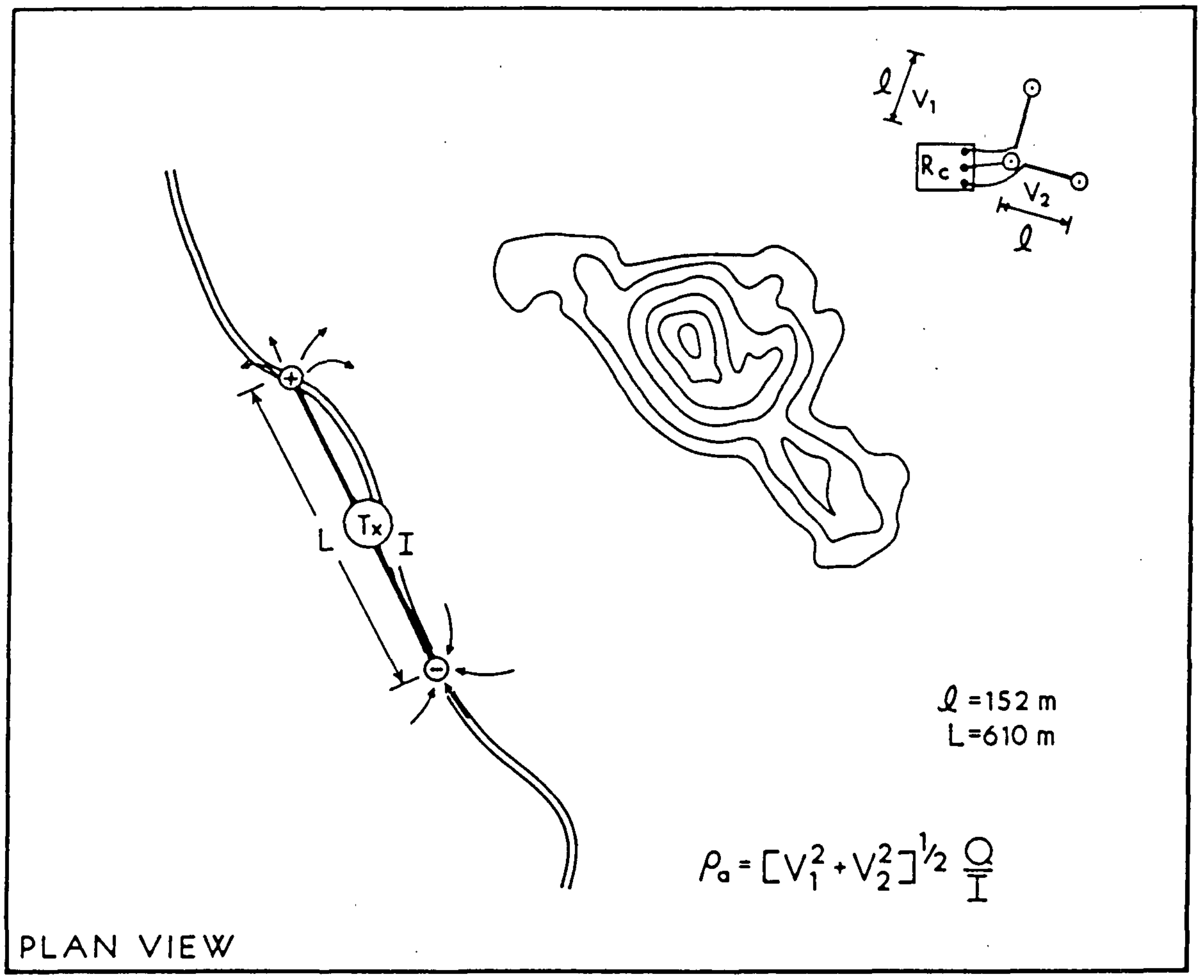


Figure 19a. Bipole - dipole array geometry as used in the reconnaissance resistivity survey.

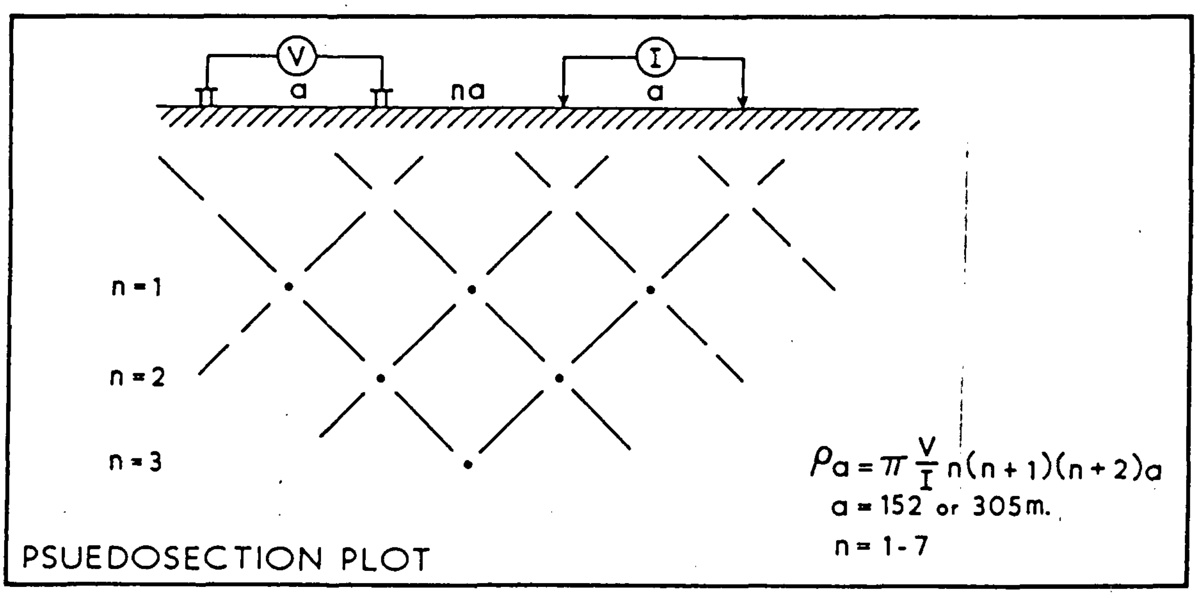


Figure 19b. Dipole - dipole array geometry used for detailed surveys.

and Ward, 1980). For regional reconnaissance surveys, the current is introduced through a longer (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 may be difficult to interpret accurately and are, in general, more appropriate for regional-scale interpretation.

Dipole-dipole surveys. The dipole-dipole array has evolved as the most popular resistivity array for detailed geothermal exploration. It has been widely used in the United States, Kenya, Ethiopia, the Philippines and numerous other countries, and often follows the use of Schlumberger or total-field surveys to provide more detail. The geometry and plotting scheme for this array are shown in Figure 19b. All electrodes are placed in a line, a uniform distance (separation) apart. The dipole-dipole array is widely used in geothermal, mineral and petroleum exploration because it is an efficient means of collecting a large number of data points which are influenced by both the lateral position and depth characteristics of the resistivity distribution. Numerical modeling programs can be used in a forward modeling or iterative manner to determine the resistivity distribution and the intrinsic resistivity values.

A variety of other resistivity arrays is possible of course, but seem to offer no advantage over the appropriate application of the Schlumberger, bipole-dipole, or dipole-dipole arrays for the purpose described.

Induced polarization (IP). Induced electrical polarization (IP) is a phenomenon that is much used in mining exploration because of the large polarization effects of sulfide minerals. The origin of the phenomenon arises mainly in chemical concentration gradients created by the flow ofcurrent in the earth. IP anomalies may arise from pyrite and clay distributions found as alteration products in geothermal areas. The measurement can be made with the dipole-dipole or bipole-dipole array at small additional cost of the resistivity measurement. Ward and Sill (1984) recently reviewed the principals and measurement techniques for this method as applied to geothermal exploration. In practice, few induced polarization measurements are reported for geothermal areas, and those we have examined show low amplitude anomalies and no definite relationship to the geothermal system.

Magnetotelluric and Audiomagnetotelluric Methods. 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.

The MT/AMT method utilizes the earth's natural electric and magnetic fields to infer the electrical resistivity of the subsurface. Figure 20 is a generalized natural magnetic field amplitude spectrum taken from Campbell (1967). There is, of course, a corresponding electric field spectrum, related through Maxwell's equations.

In general, the fields above 1 Hz are due to worldwide thunderstorms, the principal storm centers being in South America, Africa, and the Southwest Pacific. Because the ionosphere (a layer in the earth's atmosphere that extends from 60 km to 300+ km above the surface) is an electrical conductor (a plasma) the energy from lightning discharges propagates in a wave guide mode in the earth-ionosphere cavity. The resonances shown in Figure 20 are due to constructive interference.

Below 1 Hz the fields, called micropulsations, are mainly due to the complex interaction of charged particles from the sun with the earth's magnetic field and ionosphere. As Figure 20 shows, the amplitude of the electromagnetic (EM) field increases with decreasing frequency below 0.1 Hz. Important references on natural EM fields are: Jacobs (1970), Matsushita and Campbell (1967), and Bleil (1964).

These natural electromagnetic fields represent noise for controlled-source EM (CSEM) methods, but they are the source fields for MT/AMT. Since low frequencies are needed for deep penetration, it is easy to see from Figure 20 why MT has been used so extensively for crustal studies and deep exploration: The source fields increase at low frequencies for MT while the noise increases at low frequencies for CSEM.

Figure 21 shows typical MT/AMT signals as inscribed on chart recorders in the field. This information would be recorded on magnetic tape in digital format simultaneously.

In the magnetotelluric methods, the apparent resistivity, ρ_a, is calculated from the ratio of electric and magnetic field magnitudes for a given frequency according to the relationship,

$$\rho_a = \frac{1}{2\pi f \mu_o} \left(\frac{E_x}{H_y}\right)^2 ,$$

where E_x is the horizontal component of the electric field, H_y is the perpendicular magnetic field component, and μ_o is the magnetic permeability (henrys/meter) of free space. The present state-of-the-art generally requires

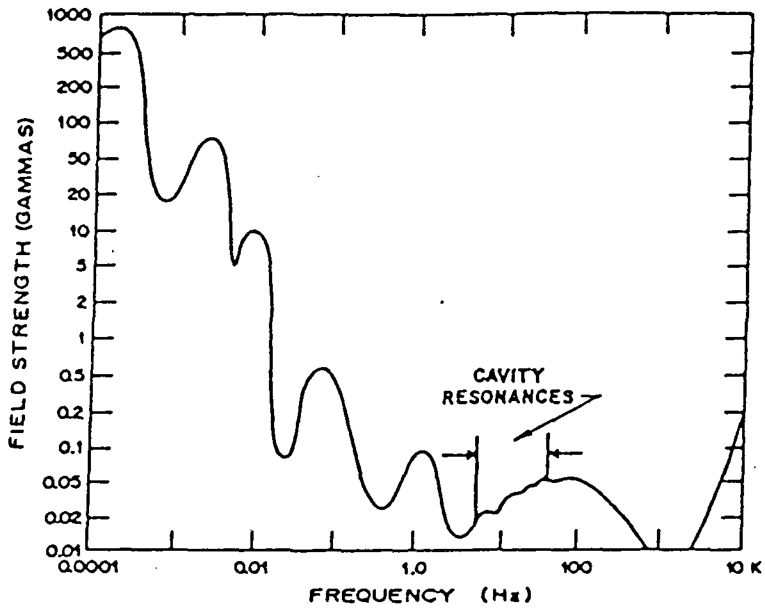


Figure 20. The Natural Electromagnetic Field Spectrum. The portion above 1 Hz is due mainly to worldwide thunderstorm activity whereas that below 1 Hz is due mainly to the complex interaction of the solar wind with the earth's magnetic field and ionosphere.

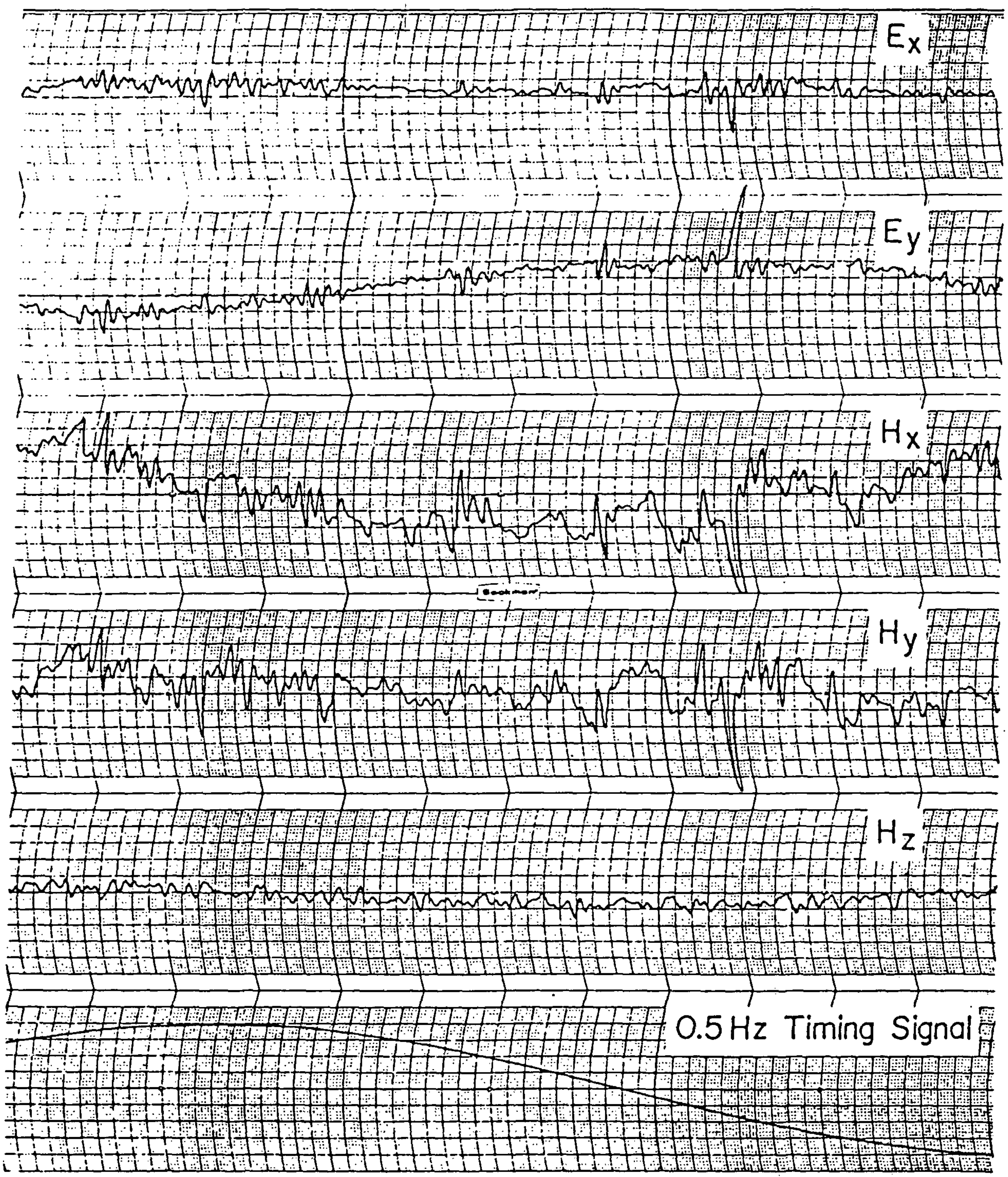


Figure 21. Chart Recordings of Typical MT/AMT Signals (after Vozoff, 1972). Two horizontal components of electric field (E_x and E_y) and three components of magnetic field (H_x, H_y and H_z) are shown along with a 0.5 Hz timing signal. E_z is always taken as zero because essentially no electric current crosses the earth-air interface at the surface. High frequency signals (the jagged peaks) are superimposed on lower frequency variations.

full tensor measurements, superconducting magnetometers, remote reference magnetometers and continuous magnetic tape recording of all parameters and in field data recording. A typical frequency range for exploration surveys may be 0.0005 to 200 Hz which could correspond to depths of 0.2 to 20 Km. Vozoff (1972) and Wannamaker et al. (1980) present excellent descriptions and examples of the MT survey methods. Vozoff's paper is especially applicable to sedimentary basins.

We have noted that MT has been used in most of the high-temperature resource exploration programs in the western United States. We attribute this to its advertised great depth of exploration and 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 volcanoes 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).

Drawbacks of the MT method are the logistics, expensive equipment and long recording times, which may result in costs of $2000 to $4000 per station. Sophisticated data processing and numerical model interpretation will be additive to these costs. The interpretation in areas of complex 2-D or 3-D geometries can also be misleading and a poor representation of the true earth situation.

The depth of penetration of the electromagnetic fields into the earth is generally related to the skin depth δ, at which depth the fields have fallen to $(e)^{-1}$ of the value at the surface. Thus,

$$\delta = 503 \sqrt{\rho/f} \quad \text{(meters)}$$

when ρ is the resistivity of a homogeneous earth (in ohm-m) and f is the frequency of the signal.

The audiomagnetotelluric (AMT) method, in which field measurements are restricted to a higher range of frequencies (10-2000 Hz) has been used in many geothermal areas at a much reduced cost and higher rate of aerial coverage. The use of the method in geothermal areas has been reported by Hoover et al. (1978) and by other authors, and appears to be cost effective in many geologic settings where the depth of exploration is less than 1000 meters.

Two problems have been frequently encountered in AMT surveys: 1) low or erratic natural field strengths (near the minimum of the earth's natural field strengths as a function of frequency) and 2) poor depth penetration due to these weak field strengths. To overcome these problems a controlled source such as a long transmitter dipole can be used (Goldstein and Strangway, 1975; Sandberg and Hohmann, 1982). Contract CSAMT surveys are being conducted in the United States, Mexico, and Australia for the frequency range 0.5-2048 Hz

(K. Zonge, personal communication) at costs of approximately $1000 per station.

The effective depth penetration for the CSAMT method is given by

$$d = \delta/\sqrt{2} = 503\ \sqrt{2\rho/f} = 356\ \sqrt{\rho/f}\ \text{(meters)}.$$

Thus CSAMT surveys have the potential depth exploration exceeding 1000 m in a uniform 10 ohm-m earth, and have been used in geothermal exploration in the western United States.

The <u>MT. - 5 - E.x.</u> is one particular field magnetotelluric survey method which seems to have been used for many geothermal surveys in Europe and Africa. It utilizes a harmonic analysis of the recorded data and an exponential solution of Maxwell's equations to arrive at values of longitudinal conductance and apparent vertical resistivity (Musé, 1973). Although the contractor offering these surveys was located in the United States there is little documentation of its use in the United States, and no publication of the method in the principal geophysical literature. The computed parameters may be poor estimates of the true earth characteristics in complex 2-D and 3-D environments.

<u>Controlled Source Electromagnetic Methods (CSEM)</u>. These methods have been used as an alternative to resistivity methods in some geothermal environments. Time domain and pulse e.m. methods (TDEM) can be used in volcanic areas of high surface impedance such as Hawaii (Kauahikaua, 1981) where grounded resistivity surveys are slow and costly. Ward (1983) and Keller (1970) have reviewed the application of these methods to geothermal exploration. Wilt et al. (1981) describe a high power system developed primarily for geothermal exploration.

The primary limitation to these methods is that interpretation techniques have been well worked out only for the layered-earth, 1-D case. If the actual subsurface has a resistivity distribution that is 2-D or 3-D in nature, interpretations using 1-D theory and techniques can be very misleading.

It seems to us that the CSEM methods offer little or no advantage over conventional galvanic resistivity surveys and that their interpretation is much less satisfactory. However, these techniques should be matched because advances in instrumentation and interpretation are being made, and a tool suitable for geothermal use may evolve.

Some Problems with the Resistivity and IP Methods. Conductive overburden, generally in the form of porous alluvium or weathered bedrock, sometimes prevents current from effectively penetrating to deeper levels where resistivity data are needed. Hence, the deep resistivity structure influences the measurements less than they would if the overburden was absent. For surface electrode arrays, conductive overburden represents a fundamental limitation. However, one way of combatting it is to force current into the bedrock by placing an electrode in a drill hole.

Electromagnetic (EM) coupling presents a serious problem for IP and resistivity surveys, particularly when large electrode separations are used in areas of low resistivity. The EM eddy currents in the ground caused by current in the transmitting circuit vary with frequency, causing resistivity and IP values to be incorrect.

The first step in combatting EM coupling is to use an appropriate electrode array. Arrays such as the Schlumberger and Wenner, where measurements are made between widely spaced current electrodes, generate large EM coupling and should be used with caution. If a long current line is necessary to increase the signal, measurements can be made perpendicular to the current wire near one of the electrodes, as in the three-array or the perpendicular pole-dipole array. If the earth is homogeneous, there is no EM coupling for a perpendicular array. But lateral or vertical resistivity changes can produce large, and sometimes negative, EM coupling. The commonly used in-line dipole-dipole array offers both high earth resolution and lower EM coupling, at the expense of low receiver voltage levels.

EM coupling is generally a much more serious problem in IP surveys (which are only infrequently applied to geothermal exploration) than it is in resistivity surveys.

Techniques for removing EM coupling over a broad frequency range and retaining the spectral character of IP have been proposed by Wynn and Zonge (1975) and Pelton et al. (1978). However, their validity remains to be demonstrated.

Self-potential (SP)

The self-potential (SP) method is based on the measurement of naturally occurring potential differences generated mainly by electrochemical, electrokinetic and thermoelectric sources. The multiplicity of sources can be either an advantage or a disadvantage. On the one hand, a number of phenomena can be studied with the techniques and, on the other hand, the possibility of a number of different sources can sometimes be confusing.

There has been a mild resurgence in the use of the SP method in geothermal exploration (Corwin and Hoover, 1979), in the study of earthquake related phenomena (Fitterman, 1978, 1979; Corwin and Morrison, 1977), and in engineering applications (Ogilvy et al., 1969; Bogoslovsky and Ogilvy, 1973). Self-potential (SP) measurements in geothermal areas have shown anomalous regions associated with the near surface thermal zones and faults thought to be fluid conduits (Zohdy et al., 1973; Corwin, 1975; Anderson and Johnson, 1976; Zablocki, 1976; Combs and Wilt, 1976; Mabey et al., 1978;

Corwin and Hoover, 1979). The signs of these anomalies have been both positive and negative.

Possible sources for these self potentials are electrokinetic effects (streaming potentials), thermoelectric effects and chemical potential differences. Streaming potential effects tend to be slightly favored since the combination of the typical streaming potential coefficient and the force (gradient of the pressure) generally combine to produce effects of larger magnitude from buried spherical or point sources (Corwin and Hoover, 1979). However, strong near surface thermal gradients are capable of producing sizeable self-potential anomalies.

By definition fluids are moving in a convective hydrothermal system and the system usually has zones of clay alteration. Clay minerals, through their larger cation exchange capacity, aid in the generation of electrokinetic and thermoelectric potentials. However, large percentages of clays and the higher conductivity of hydrothermal fluids would tend to reduce streaming potential effects through the reduction of permeability and the electrical resistivity. SP surveys have been conducted on many high-temperature geothermal areas in the western United States (Corwin and Hoover, 1979) and Hawaii (Zablocki, 1977). The association with steaming fissures and molten lava on Hawaii is most impressive (Zablocki, 1976). Pronounced SP anomalies, often dipolar in shape, have been documented for several geothermal systems which occur along basin and range faults in the western United States. Sill (1981; 1982a,b,c) has developed the mathematics for modeling these anomalies as cross-coupled flows due to hydrothermal convection, principally arising from electrokinetic and thermoelectric effects. The SP surveys are generally conducted in a detailed exploration mode, as a series of traverses perpendicular to structures believed to be carrying the thermal fluids. The surveys are more likely to be cost effective when the moving fluids are within 500 m of the surface.

Measurement errors, i.e. noise, arise from the electrodes, small scale variations in the ground potential, and on a larger scale, telluric currents. Electrode generated errors, sometimes called pot noise, can arise from temperature changes, electrolyte concentration changes in the porous pots and in the porous ceramic. These errors usually occur as slow drifts in relative potential over a period of hours. They can be partially compensated by checking the potential differences between the electrodes, at the same location, several times during the course of the survey and using this data to make linear drift correction. Watering of the electrode stations, to reduce pot resistance should be avoided as it can cause potential transients of 5 to 10 mv, lasting as long as an hour.

Small scale (cm to m) potential differences exist in the ground due to changes in the soil and soil moisture and the biological activity of plants. These potential differences are typically in the range from 1 to 10 mv and can be partially compensated by making a number of readings over a small area and averaging the results.

Telluric currents produce potential gradients in the range from 1 to 10 mv/km. On long lines these potentials can be a source of error. Relatively rapid fluctuations (from 1 to 10 sec), when observable on the meter, can be averaged but this is not practical for longer period fluctuations. These could be partially compensated by monitoring the low frequency variations on a fixed dipole but this is not usually done.

Cultural effects due to DC power systems, pipes, cased drill holes, roads (disturbed soil) and cultivated fields (fertilizer) have been observed. Topographic effects, possibly due to the motion of groundwater, are also present.

In considering the possible noise sources we see that the leapfrog method has the logistical advantage of using a short line but since both electrodes are moved it is subject to more pot noise. This technique combined with the finite precision of the measurement acts like a high pass spatial filter and attenuates the long wavelength, low amplitude fluctuations. This is effective in reducing the effects of telluric currents, but it will also attenuate the long wave length anomalies due to other sources. The long line method has the advantage of pot noise from only a single electrode but at large distances the telluric current variations can cause problems.

Data quality can be assessed by repeated measurements and by closure errors on closed loop surveys. With reasonable care, repeated surveys show a typical scatter of ±5 mv to ±10 mv and closure errors as small as a few tens of millivolts.

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.

Several types of passive seismic surveys have been conducted on a local or prospect scale for geothermal exploration. Seismic noise and seismic emission surveys attempt to record and locate very low amplitude seismic activity that has been noted in several high-temperature thermal areas. The emissions appear to arise from the movement of hot fluids and gases and rather continuous minor rock deformation.

Our review of the technical literature suggests that much of the interest in the seismic emission study was developed by academic and government scientists in the United States where seismic noise anomalies correlated well with several venting or near surface high-temperature geothermal resources (Roosevelt Hot Springs, Utah; Yellowstone National Park; The Geysers etc.). Sample ground noise surveys have yielded high levels of noise over Taupo, New Zealand (Clacy, 1968), The Geysers (Lange and Westphal, 1969), and in the Imperial Valley (Douze and Sorrells, 1972). However, it has been pointed out that these high noise levels can have other causes (Liaw and McEvilly, 1979). Private companies engaged in geothermal exploration have tried to expand the technique to the exploration for deeper, blind resources with little, if any, documented success (Ward et al., 1981). We do not consider seismic emissions or noise studies to be an integral part of the exploration program.

The passive seismic technique in use today is the microearthquake method (MEQ). In this technique a tight array of dectors is deployed to map microearthquake hypocenters, and numerous surveys have been conducted with variable degrees of success at many geothermal fields. Microearthquake surveys have thus evolved into a very systematic integral part of geothermal exploration programs. The technique seems to be particularly well suited to the exploration for fault controlled resources (Basin and Range, Western United States) and volcanic resources (Hawaii, Cascade Province).

Microearthquake surveys have been completed in several geothermal areas including East Mesa (Combs and Hadly, 1977); Coso (Combs and Rotstein, 1976); Wairakei (Hunt and Latter, 1982) to name but a few. 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.

The typical exploration survey would use a network of four to ten micro-earthquake recording stations systematically deployed over an area of perhaps 100 to 1000 sq km. If the survey area has already been restricted by other geothermal indicators or economic considerations the stations may be occupied continuously for a period of 14 to 100 days, depending on the level of seismicity, judgement of episodic behavior and funding committed to the survey. In a reconnaissance mode half or more of the stations are "leap frogged" to new locations every three-to-ten days resulting in a less complete coverage of a much larger area.

The most certain results of a microearthquake survey are: the determination of relative seismicity of the area (but only for the time period of the survey), and the location of hypocenters. A linear alignment of hypocenters may define the location of active structures most likely to carry geothermal fluids, as at Roosevelt Hot Springs, Utah (Nielson and Zandt, 1984). The occurrence of earthquake swarm activity would also add to the priority of a target area. In favorable cases zones of fractured reservoir rock may be indicated by P-wave delay and S-wave attenuation.

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 in the United States. These surveys may also be important in locating major structures within, or bounding, sedimentary basin resources.

Active Seismic Methods

Seismic refraction profiles have been recorded at The Geysers, Yellowstone National Park, Roosevelt Hot Springs, and other geothermal areas in the western United States. 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. (1982) 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.

Reflection seismic surveys have been used in deep sedimentary basins, basin and range (fault block), and volcanic geothermal provinces. The experience history is analagous to the much broader petroleum exploration experience. When coherent reflections from the depth range of interest are recorded the method offers the best means of mapping buried structure and lithology. Thus the method is appropriate, but expensive for most sedimentary basin and some basin and range provinces. High seismic signal attenuation by poorly consolidated sediments, and scattering by near surface volcanic flows may preclude obtaining useful data in some areas, even with an intensive effort in data acquisition and digital processing. The reflection seismic method has rarely been cost effective in recent volcanic terrains and in the flood basalt filled basins of the western United States even with carefully designed survey parameters and determined processing efforts.

Of all the surface geophysical techniques in use, high resolution seismic reflection with modern 2-D and 3-D imaging techniques is receiving the greatest amount of attention. While most of the work is not necessarily directed toward fault and fracture mapping, there are many reported cases where both flat- and steeply-dipping faults were detected and properly imaged. How well that technology can be extended to geothermal environments remains a question largely unanswered. Because of the volcanic-plutonic rock assemblages and the hydrothermal alteration effects present in typical geothermal environments, it is an arguable point whether seismic reflection will have broad application to geothermal reservoir mapping problems. However, the few published results to date from geothermal areas have been encouraging. Denlinger and Kovach (1981) showed that seismic-reflection techniques applied to the steam system at Castle Rock Springs (The Geysers area) was potentially useful for detecting fracture systems within the steam reservoir, as well as for obtaining other structural-stratigraphic information. Beyer et al. (1976) reported on the value of seismic-reflection profiling for mapping concealed normal faults associated with the Leach Hot Springs geothermal system, Grass Valley, Nevada. Blakeslee (1984) processed seismic-reflection data obtained by the Comision Federal de Electricidad over the Cerro Prieto geothermal field, was able to define subtle fault features and other important velocity features related to hydrothermal effects.

Magnetic Methods

The earth's magnetic field is believed to originate at great depth, although time-varying perturbations to this field originate outside the earth, principally in the ionosphere. Although many theories have been advanced to explain the earth's magnetism, the favored one is that fluid motions in the electrically conducting iron-nickel core of the earth cause a self-perpetuating dynamo effect that generates and sustains the field. The detailed fluid motions and mechanisms have never been formalized, but the basic concept seems sound.

To a good approximation, the field at the earth's surface is dipolar and thus resembles the field that would occur if a powerful bar magnet were placed at the earth's center. The dipolar axis does not correspond with the earth's rotational axis but is displaced slightly. Thus the north and south magnetic poles, where the field becomes vertical, do not correspond with the geographic poles.

The earth's field varies in intensity from about 25,000 grams (1 gamma = 1 nanotelsa = 10^5 oersted) at the magnetic equator to about 70,000 gammas at the poles. In direction, the field is horizontal at the equator and vertical at the poles. Over most of the United States the field dips about 60 degrees northward, as it does also in Spain.

Magnetometers, in common use, measure variations in the intensity of the earth's field to about 1 gamma, although instruments that detect changes as small as 0.001 gammas are available. Spatial variations in the earth's magnetic field of interset in exploration are due to lateral variations in the magnetization of rocks near surface. Vertical, that is, layered changes in rock magnetization are not detected in magnetic surveying.

Interpretation of magnetic data is considerably more complicated than is interpretation of gravity data although both represent applications of potential field theory. One complicating factor in magnetic interpretation is that the inclination of the earth's magnetic field varies from horizontal at the magnetic equator to vertical at the magnetic poles. Therefore, the direction of induced magnetization in rock bodies varies in the same way. By contrast, the gravity field is always vertical. The result is that the gravity anomaly due to a certain body is the same no matter what its latitude or longitude on the earth, but a given magnetic body has an anomaly that is much different at the poles than at the equator. In low magnetic latitudes the body does not lie directly beneath the magnetic high. There is generally an accompanying magnetic low that is as much a part of the anomaly as is the high; it too needs to be defined in order to interpret the anomaly. This anomaly characteristic is a result of the presence of both positive and negative magnetic poles. Hence, most magnetic bodies have an anomaly that has both positive and negative components. By contrast, bodies with a positive density contrast yield only positive gravity anomalies.

Yet another complicating factor in magnetic interpretation is the possibility of remanent magnetization, which can be in any direction. The remanent component can be stronger or weaker than the induced component. Reliable location of magnetic bodies and determination of susceptibility are difficult in the presence of remanent magnetization. We can conclude that thorough knowledge must be gained of the effects of varying body shape, depth, and physical property contrasts for gravity interpretation, and to that must be added knowledge of the effects of body dip and strike, and relative magnetic field inclination. In addition, the total field, the vertical and the horizontal magnetic field components can be measured in magnetic surveying. Techniques for interpreting anomalies in each of these cases must be understood by the interpreter. Anyone lacking such knowledge should not attempt interpretation.

A wide variety of numerical techniques can be applied to gravity and magnetic data prior to interpretation in terms of subsurface physical property contrasts. Many of these techniques can be classified as filtering techniques in the sense that the data are operated upon, usually by computer, by a numerical operator whose characteristics can be tailored to specific purposes (Fuller, 1966; Battacharyya, 1965, 1978). For example, the data can be numerically filtered so that anomalies of certain spatial wavelengths are retained while others of different wavelengths are rejected. Filtering is accomplished by Fourier transforming the data, in map or profile form, to the frequency domain where frequencies are retained or rejected by simple mathematical operations. The filtered data are then transformed back to the space domain. In this way, magnetic noise due to near-surface volcanic cover can sometimes be partly removed in order to enhance anomalies below the cover.

Operators can be designed to perform other tasks. Gravity and magnetic data can be continued both upward and downward to determine the map or profile as it would be observed at a higher or lower level. Upward continuation is straightforward and reliable, but care must be taken with downward continuation because small errors in the data are amplified. Potential field data can be continued downward only to the top of the uppermost anomaly-producing body. Continuation operations can be of assistance in matching aeromagnetic surveys at different elevations (Bhattacharyya et al., 1979). Sometimes magnetic data are reduced to the pole; i.e., an operator is applied to transform the data to appear as they would if the survey had been performed at the magnetic pole where the inducing field direction is vertical (Baranov, 1957).

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 have many 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 government magnetic survey programs. These data, as at the Baltazor Hot Springs 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 et al. (1978) 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. (1982a) 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. Magnetics are routinely used in Iceland to delineate dikes, some of which are bordered by zones of high permeability (Flovenz and Georgeson, 1982; Palmasson, 1976).

We are familiar with several low level (less than 200 m above terrain) closely spaced (1 km) surveys conducted by industry and ESL/UURI which have been useful for mapping geologic structures and lithology in volcanic and igneous rock environments. The general utility of the model, the applicability to numerical modeling, the low unit costs, all argue strongly for considering the use of aeromagnetic studies in the regional and detailed assessment of geothermal resources.

Gravity Methods

The gravitational force between two bodies masses M_1 and M_2 is given by Newton's law to be $F = G M_1 M_2/r^2$, where G is the universal gravitational constant and r is the distance of separation. The force is one of attraction and is directed along the line connecting the bodies. In gravity prospecting we often speak about the acceleration of gravity, which is the acceleration that a freely falling body would experience in the earth's gravitational field. This acceleration is given by $G M_e/r_e^2$, where M_e and r_e are the mass and radius of the earth, respectively. It is found by measurement that the earth's gravitational acceleration is about 983 gals (cm/sec^2) at the poles and about 978 gals at the equator. The gal and the milligal are common units, named after Galileo, used in gravity prospecting. Gravity is less at the equator than at the poles because the equatorial radius is greater than the polar radius and because of the variation with latitude of centrifugal force due to the earth's rotation.

Modern gravity meters routinely measure spatial variations in the earth's gravity field to 0.01 milligals (1 part in 10^8) or better in field application, and the newest generation of instruments is capable of ± 0.002 milligals under ideal field conditions. These spatial variations in gravity are caused by lateral variations in rock density when measurements are restricted to the earth's surface. The average density of the earth is 5.5 gm/cm^3 and the average density of crustal rocks is about 2.67 gm/cm^3. We conclude that density must increase with depth in the earth. Such vertical density changes are not detected in surface surveys; only lateral density changes are detected. Because near-surface density variations affect the gravimeter more than do deep variations, in accordance with the inverse square nature of

Newton's law, most gravity variations of interest in geothermal exploration result from lateral changes in density within shallow crustal rocks.

The gravity technique can facilitate solution to a wide variety of geological problems. As with other geophysical techniques, successful application depends critically upon trained and experienced geophysicists and technicians who pay attention to detail and who work closely with the geologist during survey design, data reduction, and interpretation.

Because the gravimeter detects lateral variations in rock density, a density contrast must exist between the rock body under investigation and its country rock. If the body under investigation has a smaller density than the country rock, we say that there is a negative density contrast, and we expect the body to show a relative gravity low. Because the range of density in rocks in small, density contrasts of interest in exploration are small compared with the physical property contrasts in magnetic and electrical surveys. Survey variations due to latitude and elevation changes will often be much greater than the anomaly sought. Meticulous care must be taken in survey procedure and data reduction.

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 university or government 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 faults, thickness of sedimentary and 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.

Detailed gravity data have delineated major faults that probably control the geothermal fluid flow at Cove Fort-Sulphurdale, Utah (Ross et al., 1982) at Alamosa, Colorado (Mackelprang, 1983) 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. (1982b) report similar interpretations. Williams and Finn (1982) have described complexities in reduction of gravity data especially important to the Cascade Province. They report that the large silicic volcanoes, 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 volcanoes produce gravity highs as a result of higher-density subvolcanic intrusive complexes.

Plouff and Pakiser (1972) show a good example of the use of gravity data to model the geometry of a rather large intrusive complex in southwest Colorado. A large gravity low is postulated to be due to a concealed batholith that underlies a caldera complex in the San Juan Mountains.

It would appear that gravity data may contribute to both regional and detailed exploration programs in most geothermal environments.

Geophysical Well Logging

Well logging is the measurement within a borehole of physical and chemical properties of the borehole environment itself and of the rocks closely surrounding the borehole by probes which are lowered into the borehole. Although well logging is routinely applied in oil and gas exploration and development, and is also useful in the search for mineral resources it is still somewhat experimental as directed toward geothermal application. Much research remains to be done in order to fully understand the responses of various well logs in geothermal reservoirs and their typically fractured, altered, commonly igneous and metamorphic host rocks. In spite of the relative lack of knowledge of well log response in geothermal reservoirs, several logs or log combinations have been used successfully to investigate such properties as lithology, alteration, fracturing, density, porosity, fluid flow and sulfide content, all of which may be critical in deciding how and in what intervals to complete, case, cement or stimulate the well.

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

Well logging operations are routinely performed during the drilling process at planned intervals of depth and certainly whenever casing is to be installed. The presence of casing severely compromises the ability of nearly all logs to respond to changes in the wall rock, and certain logs, such as the electric logs, are useless in cased wells. It is extremely important to have an adequate suite of logs for portions of the well that are to be cased off because they will represent the only indication of permeable zones since production and injection tests can not, of course, be performed for cased intervals. It is common practice not to repeat logs in sections of the well that have been previously logged, but simply to provide adequate overlap with the previous logging run to facilitate interpretation of logs that may be made with different instruments and different calibrations on successive logging runs.

Few developers or drilling contractors offer logging services themselves. Geophysical logging of the well is almost alway done by a separate group or contractor. State of the art contractor logging services are available throughout the free world.

In Table 5 is given a brief summary of logs that have been applied to geothermal well logging, and a brief explanation of these logs follows below. Table 6 lists the commercially available geothermal well logging services along with temperature and pressure limitations for the tools.

The caliper log, a measurement of borehole diameter, is used among other things to locate fracture zones or poorly consolidated lithologies that cave into the hole. It is also critical for correcting other borehole measurements which are sensitive to hole diameter. Multiple logging tools generally include a caliper log, and caliper correction to other logs can be made automa-

TABLE 5

LOGGING TOOLS, PROPERTY MEASURED AND GEOTHERMAL APPLICATION
(Modified from Glenn and Hohmann, 1981).

Logging Tool	Property Measured	Application
Caliper	Borehole diameter and shape	Hole completion[1], fractures[3], lithology[3], correction of other measurements[1].
Temperature	Temperature	Fracturing[3], fluid flow[1,3], oxidation[3], lithology[1,3], corrections of other measurements[1].
Resistivity/IP	Complex resistivity	Lithology identification[2,3], sulfide and clay content[2,4], correlation[3].
Spontaneous polarization	Natural voltage in the earth	Lithology[3], mineralization[4], oxidation-reduction[2,4].
Natural gamma	Natural gamma radiation, count or spectral	Lithology[1,3], correlation[1], U_3O_8[1], K_2O[1] (borehole assaying)[1].
Gamma-Gamma	Scattered gamma rays	Bulk density[1], porosity[2], lithology[2], borehole assaying[2].
Neutron	Capture gamma rays; thermal, epithermal or fission neutrons	Borehole assay[1], porosity[2], chemically bound water[2], lithology[2].
Acoustic	Acoustic velocity; interval transit time	Lithology[3] fracturing[1,3], alteration[4].
Spinner	Flow of fluids along the borehole	Production zones, zones of fluid uptake

1. Direct quantitative
2. Indirect quantitative
3. Direct qualitative
4. Indirect qualitative

TABLE 6

COMMERCIALLY AVAILABLE SLIM HOLE LOGGING TOOLS

Tool Type	Wellbore	O.D. (in)	Max. Press (ksi)	Max. Temp. °F/°C)
Schlumberger (Schlumberger Services Catalog, 1978)				
Resistivity				
Induction	Open	2-3/4, 3-7/8	20	350/175
Electrical	Open	3-3/8	25	500/260
Induction-Spherically Focused	Open	3-1/2	20	350/175
Dual Induction Laterolog	Open	3-/38, 3/-7/8	20	350/175
Ultralong Spaced Electrical	Open	3-5/8	20	350/175
Porosity				
Formation density	Open	2-3/4	25	500/260
	Open	3-/38	20	400/205
Compensated Sonic	Open	1/11/16	16.5	350/150
	Open	3-3/8, 3-5/8	20	350/175
	Open	2-3/4, 3-3/8	25	500/260
Long Space Sonic	Open	3-5/8	20	350/175
Compensated Neutron	Open	2-3/4	25	500/260
Natural Gamma	Open	3-5/8	20	350/175
Temperature				
Temperature	Open	1-11/16	15	350/175
Flowmeter-Temperature	Open	1-11/16	20	500/260
Drill String				
Electrical	Through	1-1/2	20	350-500/175-20
Induction	Drill Stem	2-3/4	20	350-400/175-20
Sonic	Drill Stem	1-11/16	16	300/150
Neutron	Drill Stem	2-3/4	25	500/260
Formation Density	Drill Stem	2-3/4	25	500/260
Gamma Ray	Drill Stem	2-5/8	25	500/160
Thermal Decay	Drill Stem	1-11/16	16.5	300/150
Production Logging				
Continuous Flowmeter	Cased	1-11/16	15	350-600/175-31
Gradiometer	Cased	1-11/16	15	350/175
High Resolution Thermometer	Cased	1-11/16	15	350/175
Fluid Sampler (650 & 836 cc)	Cased	1-11/16, 2-1/2	10	350/175
Radioactive Tracer	Cased	1-11/16	20	275/135

TABLE 6 (cont.)

Tool Type	Wellbore	O.D. (in)	Max. Press (ksi)	Max. Temp. °F/°C)
Logging in Casing				
Gamma Ray	Cased	1-11/16, 2, 2-3/8, 3-3/8	20	350-500/175-260
Neutron	Cased	1-11/16, 2, 2-3/8, 3-3/8	12-25	350-500/175-260
Thermal Nuetron Decay	Cased	-11/16	16.5	300/150
Dresser Atlas				
Electrical				
Induction-Electrolog	Open	2.0	17	350/175
Dual Induction Focused	Open	3-5/8	18	350/175
		3-3/8	25	400/204
Dual Laterlog	Open	3-5/8	20	400/204
Radioactive				
Compensated Neutron	Open	2-3/4, 3-5/8	20	300/150
Gamma-Neutron	Open	1-11/16, 3-3/8		
		2-3/4, 3-3/8	17	300/150
Compensated Densilog	Open	3	20	300/150
Epithermal Neutron	Open	3	20	300/150
Gamma Spectra	Open	3-5/8	20	400/204
Acoustic				
Acoustilog	Open	2-3/4	20	450/
		3-3/8, 3-7/8	20	350/175
Production Logging				
Nuclear Flolog	Cased	1-1/2	12	350/175
Tracerlog	Cased	1-1/2	12	350/175
Fluid Density	Open or Cased	1-3/4	15	400/204
Temperature	Open or Cased	1-11/16	17	400/204
Flowmeter	Open or Cased	1-11/16, 1-1/8	18	300/150
Fluid Sampler	Open	1-11/16	10	300/150

tically during the logging process. Three- or four-arm caliper tools may be employed to determine the shape of the borehole as well as its size.

Temperature logging can help locate zones of fracturing and fluid flow in a borehole, if the flowing waters are warmer or cooler than the rock. Open zones along which fluids can flow can be detected by comparing temperature logs made during drilling with those after the hole has reached thermal equilibrium, usually several months after drilling ceases. Permeable zones, which have taken up drilling fluids, will often reveal themselves as temperature spikes on the first surveys which disappear on equilibrium surveys.

Temperature information commonly is required to correct other logs, notably resistivity. For this reason a temperature log is generally included on the tool along with other logs. However, the requirements in sensitivity and accuracy of temperature logs used only for correction are not sufficient for the purposes detailed logging to detect zones of fluid flow. One generally needs a calibrated log with a sensitivity of ± 0.01 C° for this purpose, and so a special temperature logging tool is called for.

Conventional resistivity logs, including long- and short-normal and lateral logs, have been very useful in the petroleum environment for characterizing sedimentary sequences. These logs, however, are presently much more difficult to interpret in igneous and metamorphic rocks, due not only to lack of experience but also to inadequate measuring capabilities and calibration of such logging tools originally designed for sedimentary rocks (Keys, 1979). Nonetheless, resistivity logs, properly evaluated, can provide valuable information about aspects of a reservoir likely to affect fluid production, generalized lithology, fracturing and clay content. For example, the resistivity of many unaltered igneous rocks is several thousand ohm-meters (Keys, 1979), but if veined with sulfides, altered to clay, or fractured and saturated with conductive fluid, these rocks become very conductive.

The spontaneous potential (SP) log is a measurement of natural voltage of a borehole electrode relative to a surface electrode. In sedimentary sequences SP logs are used primarily to detect and correlate permeable beds (usually sandstones) and to give qualitative indications of bed shaliness. In igneous and metamorphic rocks SP is presently quite difficult to interpret, although it has been locally successful in detecting water entry zones, which produce the streaming potential (Keys, 1979) discussed above.

Radioactivity logging methods can also be useful for characterizing the geothermal environment. Certain of these logs are sensitive to lithologic variations, even behind casing; others are helpful in locating fractures. Both passive and active radioactivity logging techniques have been developed. Passive methods measure the natural radioactivity of rocks by detecting gamma rays. Active methods use natural or induced radiation from a logging tool to observe various kinds of scattered radiation.

The natural gamma log is a passive technique useful for identification of rock types in a borehole, for detection and evaluation of radioactive mineral deposits (such as potash and uranium) and, in some cases, for fracture identification. In sedimentary sequences, this log usually reflects shale content, since radioactive elements tend to concentrate in clay minerals. Potassium-rich rocks such as granite and rhyolite are readily detected by natural-gamma logs, which record the decay of ^{40}K to ^{40}Ar. Keys (1979) reports that fractures in altered rocks locally may be enriched in radioactive elements and therefore detectable on natural gamma logs.

Natural gamma logging tools measure either total counts above a threshold energy level, counts in selected energy windows, or counts in 1000-4000 or more individual detection channels. Total counts are a qualitative indicator of abundance of several natural radioactive elements. Measuring counts in energy windows specifically designed to detect thorium, uranium and potassium. The primary natural radioactive elements yields more useful information for interpretation in terms of the variation of geology downhole, particularly if the measurements are corrected for "dead time", borehole size, fluid composition, rock moisture and casing.

The gamma-ray density log is an active technique whereby the number of Compton scattering collisions between source gamma rays and formation electrons is measured. This number varies directly with formation electron density, which in turn directly reflects bulk rock density. One to lack of calibration, gamma ray density logging may not be presently as useful in igneous and metamorphic rocks as in sedimentary terrain. Densities of certain igneous and metamorphic rocks, for example, may exceed the calibration range of commercially available logging tools. Additionally, gamma-ray density logs are extremely sensitive to borehole size, mitigating their usefulness in highly fractured or otherwise easily caved rocks.

Another active radioactive technique is neutron logging, designed primarily to respond to variations in rock porosity, a critical variable in geothermal systems. In this technique, high-energy neutrons emitted from a source within the tool collide with nuclei of elements in the rock, thereby losing energy in an amount which is greatest when emitted neutrons and the formation nuclei with which they collide are of equivalent mass. Formation hydrogen nuclei thus cause maximum energy decay. Successive collisions slow the neutrons to thermal velocities, corresponding to energies of about 0.025 electron-volts; the neutrons then are readily captured by various elements in the rock. Either the thermal neutrons themselves or the gamma rays that are emitted when they are captured can be measured to determine relative formation hydrogen content.

Classical application of neutron logging to determine porosity assumes that this hydrogen is restricted to free water confined to pore spaces. In many rocks, however, particularly certain mafic-rich and hydrothermally altered varieties, much hydrogen occurs in bound water in hydrous framework minerals -- for example, biotite, hornblende, sericite and montmorillonite. In such cases, the neutron response reflects rock type or alteration intensity rather than porosity (Nelson and Glenn, 1975). Furthermore, neutron tools are presently calibrated for matrix effects only in sedimentary rocks and thus cannot be expected to yield accurate porosity measurements in igneous and metamorphic rocks.

Acoustic logs yield valuable information about host rock, fracturing and porosity of a deposit and its surroundings. Standard acoustic logs, measure the time required (interval transit time) for a compressional sound wave to travel through a given distance in the formation. The interval transit time can be empirically related to porosity for certain rock types (Wyllie et al., 1956). It can be also correlated with rock quality designation or intensity of fracturing (Nelson and Glenn, 1975). Fractures can be located by analyzing the full wave form of the incoming acoustic velocity signal (Myung and Helander, 1972).

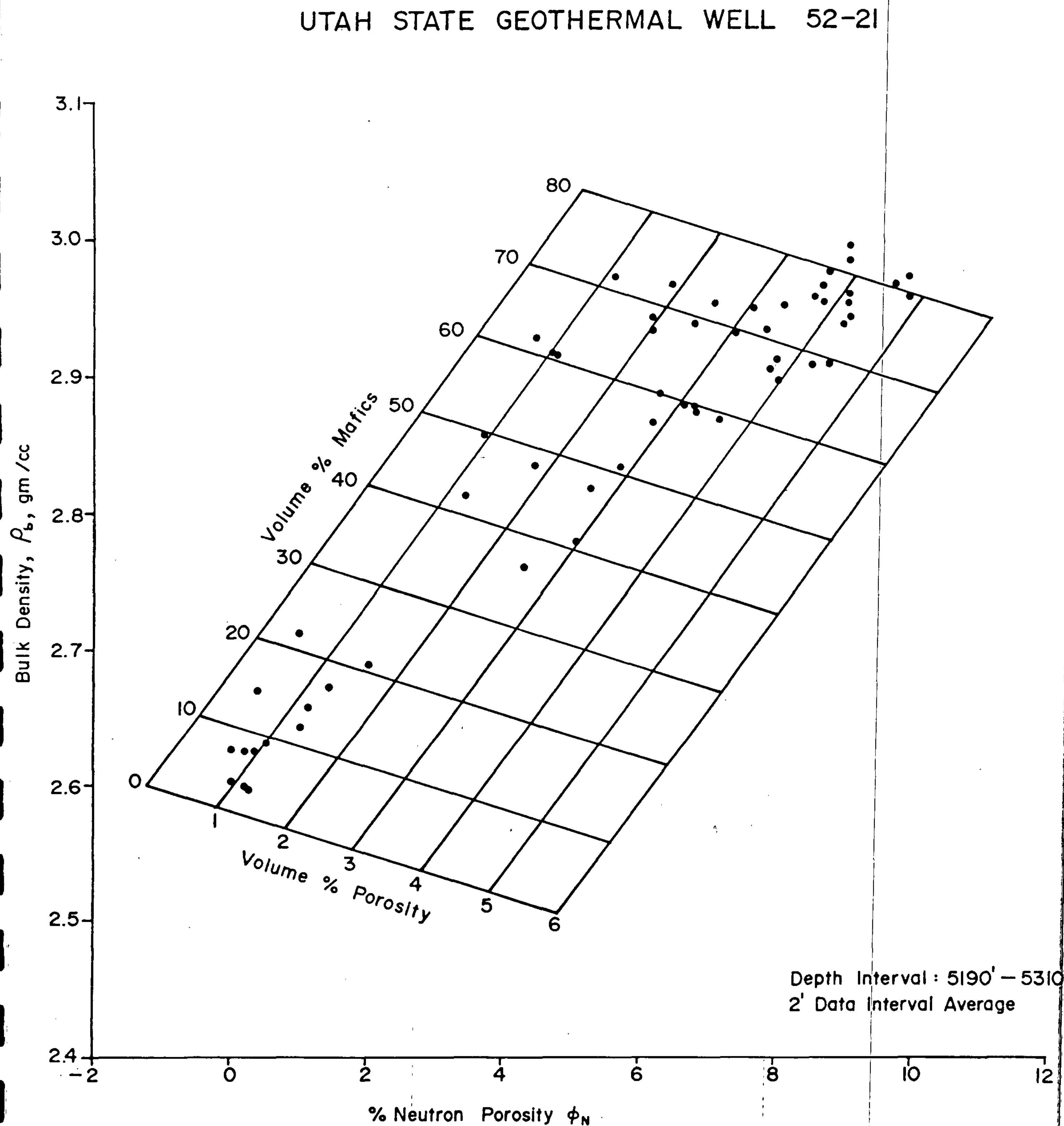
The acoustic televiewer, also known as the borehole televiewer or seisviewer, provides, through complex instrumentation described by Heard (1980), and oriented acoustic image of the borehole wall. From this image, the attitude, irregularity and aperture of borehole-intersected fractures can be determined. These fracture parameters are crucial in determining the nature of permeability in a concealed deposit to be leached or solution mined.

Cross plots of one borehole data type vs. another can greatly facilitate data interpretation, particularly for boreholes in complex igneous and metamorphic terrain (Ritch, 1975; Glenn and Hulen, 1979a,b). As an example of the utility of these plots, bulk density is plotted against neutron porosity in Figure 22 to illustrate the deceptive effect of dense, hydrous mafic minerals on tool reponses. The plotted data on the figure indicate that, contrary to expectation, bulk density increases as neutron porosity increases. The density increase is known to be due to an increase in content of the relatively dense mafic minerals hornblende and biotite (Glenn and Hulen, 1979a,b). These mafic minerals contain abundant bound water, to which, as discussed above, the neutron porosity tool readily responds. Thus, the apparent porosity increase is spurious. Superimposed on the crossplot is a grid (with origin offset from 0 to compensate for the neutron log's limestone calibration) which allows adjustment of these false porosity values. The grid shows, for example, that a rock in the borehole with bulk density of 2.71 and neutron porosity of one percent contains about 27 per cent hydrous mafic minerals and has only about 0.2% actual porosity; another rock of similar density and 3% neutron porosity contains about 32% mafic minerals and has a little less than 2% actual porosity.

Surface-to-Borehole Techniques

The class of techniques which we call surface-to-borehole require a combination of surface and in-hole sources and/or receivers. The least experimental of these is vertical seismic profiling (VSP) using both P- and S-wave surface sources (usually mechanical vibrators) arranged circumferentially around the well. Direct and reflected waves are detected by means of strings of down-hole geophones clamped to the well wall or hydrophones. VSP has been used mainly to trace seismic events observed at surface to their point of origin in the earth and to obtain better estimates for the acoustic properties

Figure 22



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of a stratigraphic sequence (Balch et al., 1982). Gal'perin (1973) presented a review of VSP research in the USSR including recent results of three-component VSP (P- and S-wave sources with 3-component detectors) to estimate compressional-shear velocity ratios and Poisson's ratio. While much of the interest in VSP has centered on better stratigraphic interpretations, particularly in difficult areas where conventional surface-to-surface reflection surveys have not proved entirely satisfactory, VSP conducted by using multiple P- and S-wave sources around a well has the potential for resolving local structural discontinuities near the well. In this regard VSP may be considered experimental. An S-wave shadow zone was detected following one hydrofrac operation at 2300 feet (Fehler et al., 1982). On the basis of data from three shot points, a finite-difference model showed that the shadow data fitted other information about the hydrofac. However, due to the low frequency S-wave source and the long wavelength of the S-wave (200 feet) in the medium, it is apparent that the fractured region must have large dimensions (a few wavelengths) for this shadow effect to occur.

A source of noise in VSP surveys are tube or Stonley waves which are high amplitude guided waves in the wellbore. Although they are excited mainly by the Rayleigh waves ("ground roll") crossing the well head (they are particularly severe if the source is close to the well), tube waves may also be excited by body waves impinging on fractures that intersect the wellbore. Consequently, there has been some interest in developing methodologies to derive fracture permeability information from the tube waves (Paillet, 1980). Lastly, Crampin (1978 and 1984) and others have argued that VSP conducted with 3-component geophones might prove extremely useful for mapping the fractured conditions of rocks if one were to extract seismic anisotropy information from the shear-wave splitting effect.

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

Borehole-to-borehole and borehole-to-surface resistivity methods also appear to be applicable to geothermal exploration. Yang and Ward (1985) presented theoretical results relating to detection of thin oblate spheroids and ellipsoids of arbitrary attitude. In this study, the effects of the surface of the earth are neglected and the body is assumed to be enclosed within an infinite homogeneous mass. The surface of the body is divided into a series of subsurfaces, and a numerical solution of the Fredholm integral equation is applied. Once a solution for the surface charge distribution is determined, the potential can be specified anywhere by means of Coulomb's law. The theoretical model results indicate that cross-borehole resistivity measurements are a more effective technique than single-borehole measurements for delineating resistivity anomalies in the vicinity of a borehole.

Beasley and Ward (1985) obtained the representative results in their mise-a-la-masse studies. The figure are self-explanatory. The dip of the body and the location of the energizing electrode within it were both varied; section and plan views of apparent resistivity are the end product of these computations. The maximum depth at which a body can be located and still produce a detectable surface anomaly is dependent upon the position of the buried electrode and upon the contrast in resistivity between the body and the host. It was found that locating the buried electrode <u>just outside</u> the body does not significantly alter the results from those when the electrode is embedded in the inhomogeneity.

From the above studies we tentatively conclude the following:

1) the cross-borehole method produces larger anomalies than does a single-borehole method,

2) the cross-borehole anomalies using a pole-pole array are smaller than those for a cross-borehole dipole-dipole array,

3) the cross-borehole mise-a-la-masse method produces larger anomalies than for the other cross-borehole methods, and

4) the anomalies due to a thin sheet were generally much smaller than those for a sphere as is to be expected (e.g. Dobecki, 1980).

VII. CRITIQUE OF SPANISH GEOPHYSICAL SURVEYS

Our critique of the geophysical surveys is based on a partial translation of four reports and our present understanding of the techniques used. A detailed evaluation of the surveys and their cost effectiveness would require a better understanding of survey aims and details, and was not understood to be a component of this contract.

1. Estudio Geotèrmico de las Montañas del Fuego (Lanzarote) por Mèthodos Magnetoteluricos y Electromagnèticos .

The electromagnetic method (Dipolo SOFREM) used at Lanzarote is one of a large number of electromagnetic sounding and/or profiling techniques which could be used for the determination of shallow resistivity structure in this environment. The very high surface resistivities and probable high electrode impedances certainly favor an electromagnetic method over contact resistivity techniques (such as dipole-dipole, or Wenner profiling) from operational and efficiency considerations. The main limitation of the method, as applied here, is the current depth penetration in lower resistivity areas. This may not be a serious shortcoming since the survey appears to be mapping the top of the sea water invasion zone at approximately 200 m depth near the coast. Most electromagnetic and contact electrical methods will be limited in the resolution of thermal waters in the presence of sea water intrusion this near the coast. It appears that east-west structural zones have been mapped where increased porosity and permeability are indicated by low resistivity zones and a decreased depth to the second (conductive) layer.

Map No. 2, Apparent resistivity determined by SOFREM at 560 Hz illustrates the presence of the east-west conductive zones but the linearity of the conductive zones may be less than indicated by direct contouring between profiles two to four km apart. It is also important to realize that the current depth penetration for approximately 50% of the survey area is less than 50 m into the second layer at the 560 Hz frequency mapped.

The MT.-5-E.x. method is one of several magnetotelluric survey methods. It is not reported or described in the major geophysical journals (Geophysical Prospecting, Geophysics, JGR) and does not seem to be frequently used in the United States, Latin America or within the Pacific region geothermal exploration. We have noted its use in Africa and Europe, however.

The accurate interpretation of apparent resistivity (ρ_a) and its counterpart conductance (C) from all magnetotelluric methods assumes a horizontally-traveling plane-wave e.m. source. The assumption is often invalid in nonlayered geologic environments but useful data may still be obtained. Proximity to an irregular sea coast (and the presence of 0.3 ohm-m sea water) and a complex shallow resistivity structure indicated by the SOFREM survey raise some question as to the accuracy of the calculated apparent vertical resistivity values (Map No. 6). In addition, the contractor notes telluric noise problems that result in a large number of stations with only "average" data quality.

A comparison between the shallow e.m. (SOFREM) results, Map No. 2, and the conductance (Map No. 5) and apparent vertical resistivity (Map No. 6) results of the M.T.-5-E.x. survey show very little agreement. This is due in part to the much wider distribution of MT stations (acknowledged by the contractor to be insufficient), and also due to the probable complex nature of the electromagnetic wave and the shallow resistivity structure. It seems unlikely that the M.T.-5.-E.x. data are appropriate for anything more than a qualitative evaluation of apparent resistivity below sea level depths, and certainly not to depths approaching 1000 m.

We have completed electrical survey (dipole-dipole) resistivity in a similar island environment (Ascension Island) and have encountered similar problems. We would regard the Lanzarote area a difficult setting in which to utilize electrical survey methods, and the magnetotelluric methods in particular seem to be inappropriate. Electrical surveys of various types have been completed on several volcanic islands (Hawaii, Azores, Fiji, etc.) and have met with varying degrees of success, largely dependent upon the distance of the thermal area from the presence of salt water invasion.

2. Estudio Mediante "Dipole Mapping De Las Anomalias Geothermicas de Caldes De Montbui Y La Garriga"

The "dipole method" used in the subject study is also referred to as the bipole-dipole or total-field resistivity method, and has been widely used for regional scale or reconnaissance type geothermal exploration. The method has been used by the U.S. Geological Survey, the Colorado School of Mines, the Earth Science Laboratory and various mining companies and geothermal company contractors. The method uses available access and irregular transmitter-receiver geometries to avoid difficult topography or cultural features and to cover large areas at a rather minimal cost for electrical resistivity methods. The long transmitter dipoles, 1.5 to 2.0 km, used in the sruvey enable a deep current penetration while the short receiving dipoles, only 50 m in length would seem to emphasize local, near surface lithologic differences.

Most of the study area is characterized by very high apparent resistivity (> 1000 ohm-m) typical for low porosity crystalline rocks. The data do indicate the presence of lithologic and/or structural features. Small areas of anomalously low apparent resistivity (< 1000 ohm-m) may possibly be related to thermal waters and/or alteration related to a thermal zone, or just an anomalous lithology. These areas would require more detailed survey types (i.e. TDEM or dipole-dipole resistivity profiles) to better define the nature of the low-resistivity zone and subsequently establish the relationship, if any, to the hot springs, geothermal fluids and possible reservoir areas.

3. Estudio Magnetotelluric Y Audio-MT De Las Anomalias Geotermicas De Caldes De Montbui Y La Garriga

The audiomagnetotelluric (AMT) method has been used in many geothermal exploration programs in the United States, principally by the U. S. Geological Survey. Geothermal companies seem to prefer the galvanic electrical resistivity methods. The technique may be appropriate for reconnaissance level exploration and is well suited to moderate-to-high resistivity environments (i.e., $\rho_a \geq 50$ ohm-m) where the depth of exploration is sufficient, as long as natural field strengths are adequate. If this is not the case, the controlled source AMT (CSAMT) method could be used but at additional cost.

In the IGME AMT study, several zones of relatively low apparent resistivity ($\rho_a < 135$ ohm-m) were delinated by the lower frqeuency AMT channels, i.e. 39.5 and 8.3 Hz. These areas likely correspond to a greater thickness of sediments overlying the granodiorite rocks, and in some areas may indicate the presence of warm fluids themselves. The apparent resistivities typically vary from 55-200 ohm-m, and are lower by a factor of 5 to 20 than the apparent resistivities determined in the "dipole-dipole" study of the same area. This suggests the possibility of a systematic error, such as calibration constant, in one data set.

The survey design, execution and interpretation all appear to have been completed in a professional manner. Due to the widespread lateral variations in resistivity, the depth of exploration may be less than stated for most frequencies. The estimated thickness of sediments, to 3000 m, as determined from the AMT data, should be considered accurate in a relative rather than quantitative sense.

4. Estudio Microsismico Y De Ruido Sismico De La Fosa Del Valles (Barcelona)

Passive seismic surveys have been used in several different ways in geothermal exploration. In the most general sense, it is known that most high enthalpy geothermal resources occur in zones of active tectonism as indicated by the historic record for earthquakes of magnitude $M \geq 3$. Many major faults have been located to a precision of several kilometers using this data base of infrequent earthquake occurrence. It is well documented that the frequency of seismic events increases exponentially with decreasing magnitude. Thus microearthquake surveys with a closely spaced array of stations will serve to detect a much greater number of events, and the hypocenters can be located with a much greater accuracy, in the survey area of interest. The loci of hypocenters define the active structures and these are the likely conduits for geothermal fluids.

It has also been documented that a lower level of seismic activity, seismic noise or seismic emissions, is associated with many high enthalpy geothermal occurrences. The utility of the seismic noise or seismic emissions methods have been actively debated in the United States, with the main proponents being the contractors who offer the surveys as a commercial service. These surveys may be of little value in delineating or detecting moderately deep, low-temperature systems.

The microearthquake surveys at La Fosa Del Valles detected a small number of natural events during a two month recording period, and due to funding and equipment limitations these were recorded at a single station only in a noisy location. The survey was not carried out with an adequate number of stations to determine hypocenter locations and may not have had an adequate sensitivity to detect significant local events. With these limitations, it is difficult to evaluate the seismicity of the area and the cost effectiveness of the method.

The microearthquake study was undertaken as a test or calibration of the applicability of the technique for low enthalpy geothermal system exploration. The Compañia General De Sondeos and the affiliated investigators have indicated a proper understanding of the technique. They chose to proceed with this preliminary study even when it was apparent that only one seismograph was available, thus precluding epicenter and hypocenter determination. It also appears that this station was located at a rather noisy (seismically) site further hindering the study. We concur with the recommendations of the study, i.e.:

- A minimum instrumental system amplification of 10^6
- Pre-selected, quiet sites for seismograph stations
- Sites on crystalline rocks rather than fill materials
- A minimum of 4 operating stations for a period of 4 months or more (The array geometry should be designed to insure high accuracy for hypocenter determinations)

We further agree that the method could be a cost effective technique for the location of active zones along the fault planes. Several geothermal exploration contractors have developed efficient procedures for site selection, instrument deployment and tape recording or radio relaying of data. These procedures should be adopted to greatly reduce the survey cost and increase the survey effectiveness.

The seismic noise study was also limited by available funding. One hundred fifty stations were occupied for only 20 minutes each, and the noise spectrum recorded. The recordings were taken during the daylight hours rather than a more likely quiet period late at night. The duration and time of day of the recordings were not appropriate for a sensitive seismic noise study. Most of the seismic noise seems related to industrial and cultural activities, and appears to have contributed little to the geothermal evaluation of the area. This result is similar to that of many surveys in the United States where highway and railroad noise, wind noise through vegetation, etc. have obscured the noise pattern sought as an indication of movement of geothermal fluids.

VIII. EXPLORATION STRATEGIES FOR SPANISH RESOURCES

Introduction

Geothermal development is an interdisciplinary endeavor. Figure 23 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 developed earth science tools and techniques to solve their particular exploration problems in an optimum way, and this has required the expenditure of literally tens of billions of research dollars. By contrast, relatively little has been spent in developing earth science tools and techniques especially to solve problems in the geothermal environment. Because the geothermal industry is so young, it is, for the most part, unable to fund the research and technology development needed. Geothermal developers have had to resort to application of existing earth science tools, which are not generally optimum for geothermal application. In some cases, these simply are tools or techniques to solve a particular problem. Limitations of the common geophysical techniques have been discussed elsewhere in this report.

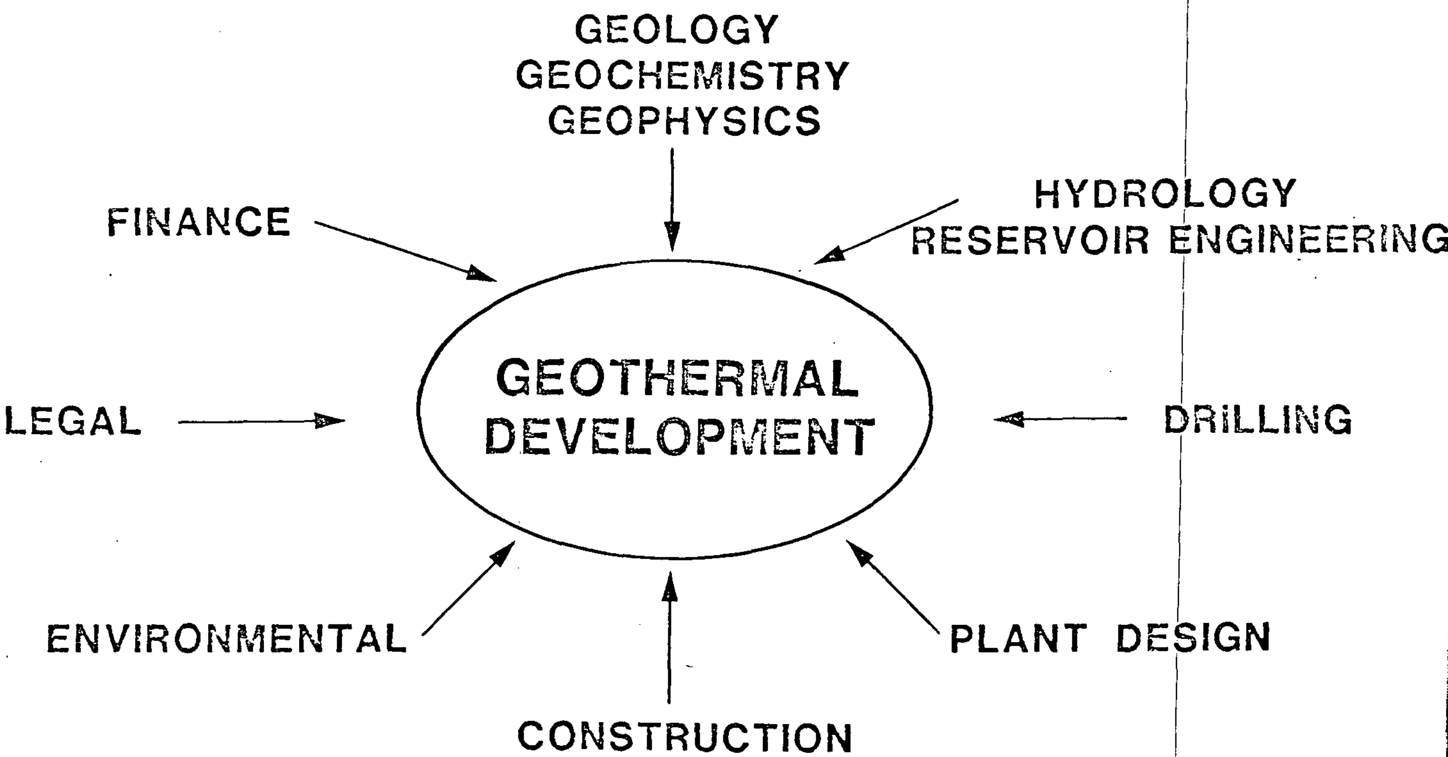
Role of the Earth Sciences in Geothermal Exploration

In this section we will briefly consider the various types of earth science data that are usually brought to bear on geothermal exploration problems.

Geology. Collection of geologic data through surface geologic mapping and through logging of drill cuttings and core provides the basic data

GEOTHERMAL DEVELOPMENT

AN INTERDISCIPLINARY ENDEAVOR



BECAUSE GEOTHERMAL OCCURRENCES ARE GEOLOGICAL PHENOMENA, EARTH SCIENCE INFORMATION IS NEEDED FOR ALL PHASES OF DEVELOPMENT

THE DEVELOPMENT TEAM MUST WORK CLOSELY TOGETHER FOR THE PROJECT TO SUCCEED

Figure 23

required for interpretation of all other exploration data. Surface geologic mapping or field evaluation of existing geologic maps should be the first step undertaken in any geothermal exploration problem. The field geologist (1) identifies separate rock units, (2) maps the structure within and among rock units (faults, fractures, folds, rock contacts), (3) studies the age relationships among rock units as shown by their mutual field relationships, (4) searches for evidence of subsurface geothermal activity, which evidence may range from obvious thermal springs, geysers and fumaroles to very subtle indications such as hydrothermal alteration of rocks or thermal spring deposits of sinter (SiO_2) or travertine ($CaCO_3$), (5) studies the geologic relationship of the particular prospecting area to regional geology, (6) collects samples of rocks and minerals for subsequent microscopic examination, age dating, geochemical analysis or geophysical characterization, and (7) collects samples of fluids from wells and springs for geochemical studies. This work helps provide first answers to many questions about the prospective geothermal area such as: (1) is there direct evidence of geothermal activity in the area?, (2) are there young (less than say, 3 million years old) volcanic rocks in the area that would indicate an underlying mass of hot rock that could provide a source of heat?, (3) are there porous and permeable rock units or are these active faults or open rock contacts that could constitute a plumbing system?, and from an overall viewpoint, (4) is this a viable geothermal prospect area and if so what exploration techniques should be used next?

<u>Geochemistry</u>. A geothermal system is a highly complex large-scale, natural chemical system. Geothermal fluids are complex brines of varying composition, concentration, acidity (pH), oxidation potential (Eh), temperatures and pressure. As they move through rocks, carried along by

hydrothermal convection and/or by hydrologically induced pressure gradients, these fluids interact chemically with the reservoir rocks, which themselves can be 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 others within a mineral. These chemical/mineralogical changes in the reservoir rocks may or may not cause volume changes, but, obviously, if the rock 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 over short distances, minerals may be precipitated into the open spaces of the plumbing system, resulting in plugging.

This complex chemical 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 geothermal systems a state of chemical equilibrium or near-equilibrium is observed to exist between reservoir fluid and reservoir rocks. Lack of equilibrium could be evidence for rapid movement of fluid through the reservoir.

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 more important role in geothermal exploration than it does in, say, petroleum exploration.

Chemical methods can be used to estimate subsurface reservoir temperature. This information is of obvious interest prior to availability of direct information obtained by drilling, but is also very important during the drilling process because (1) accurate temperature measurements cannot be made in a well until after thermal effects of the drilling process have been disrupted, and (2) fluids encountered during drilling may indicate that higher temperatures may be found elsewhere.

Geophysics. The use of geophysics in the exploration for geothermal resources is the topic of this report, and will not be elaborated here. In general terms, the geophysical surveys will map and attempt to interpret physical property distributions at depth. When integrated with the geological, geochemical and hydrological data base, it contributes to the evolving conceptual model and identifies in a cost effective manner those portions of the subsurface most appropriate for a drilling effort. The selection of which methods to apply is dependent upon the local geology and the anticipated physical property contrasts.

Hydrology. A thorough understanding of the regional and local hydrology of the prospecting area is necessary in geothermal exploration. The primary question for the hydrologist is the nature of the expected porosity and permeability at depth. Will the permeability be controlled by faults and fractures or is it expected to be intergranular in nature? Where is highest permeability likely to be found? If geothermal fluids are produced from an area, will be the reservoir be recharged or will be the fluid supply decrease? To obtain answers to these and similar questions, the hydrologist works closely with the geologist.

Geothermal Exploration - General Considerations

The geosciences have two primary applications in geothermal development:

1. Exploration for geothermal systems, and
2. Exploration within geothermal systems.

Figure 24 indicates one suggested series of steps for this exploration.

The reconnaissance stage is designed to identify prospect areas and to prioritize them for detailed exploration. This stage refers to (1) above, i.e. exploration _for_ geothermal systems. Once a geothermal system has been located, exploration becomes more detailed _within_ the system. The primary objective of both exploration phases is to _select drill sites_--drill sites to locate a resource area, to confirm the presence of a resource, and then to obtain production of fluids for the utilization plant and to dispose of spent fluids through injection. Because the drilling of geothermal wells is so costly, refinement of exploration techniques has great potential for lowering development costs by avoiding wasted drill holes.

Exploration Strategy. Figure 24 is an exploration strategy in its most basic form. Before such a strategy can become truly useful, much more detail must be added to each of the steps. In this chapter we will start with this basic strategy as a framework to develop more complex strategies. Several aspects of Figure 24 merit discussion. First, exploration proceeds from the consideration of large areas, perhaps 10,000 km^2 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 onto detailed studies of a few small areas. During this process, it is prudent to use lower unit-cost exploration techniques during the earlier stages of the program and reserve higher-cost techniques for use later when the area of interest has been reduced. Another feature of the exploration strategy is that there are a number of decision points along the way, at the end of each stage, when one may elect to terminate the project. By considering all aspects of the project and

EXPLORATION AND EVALUATION SEQUENCE

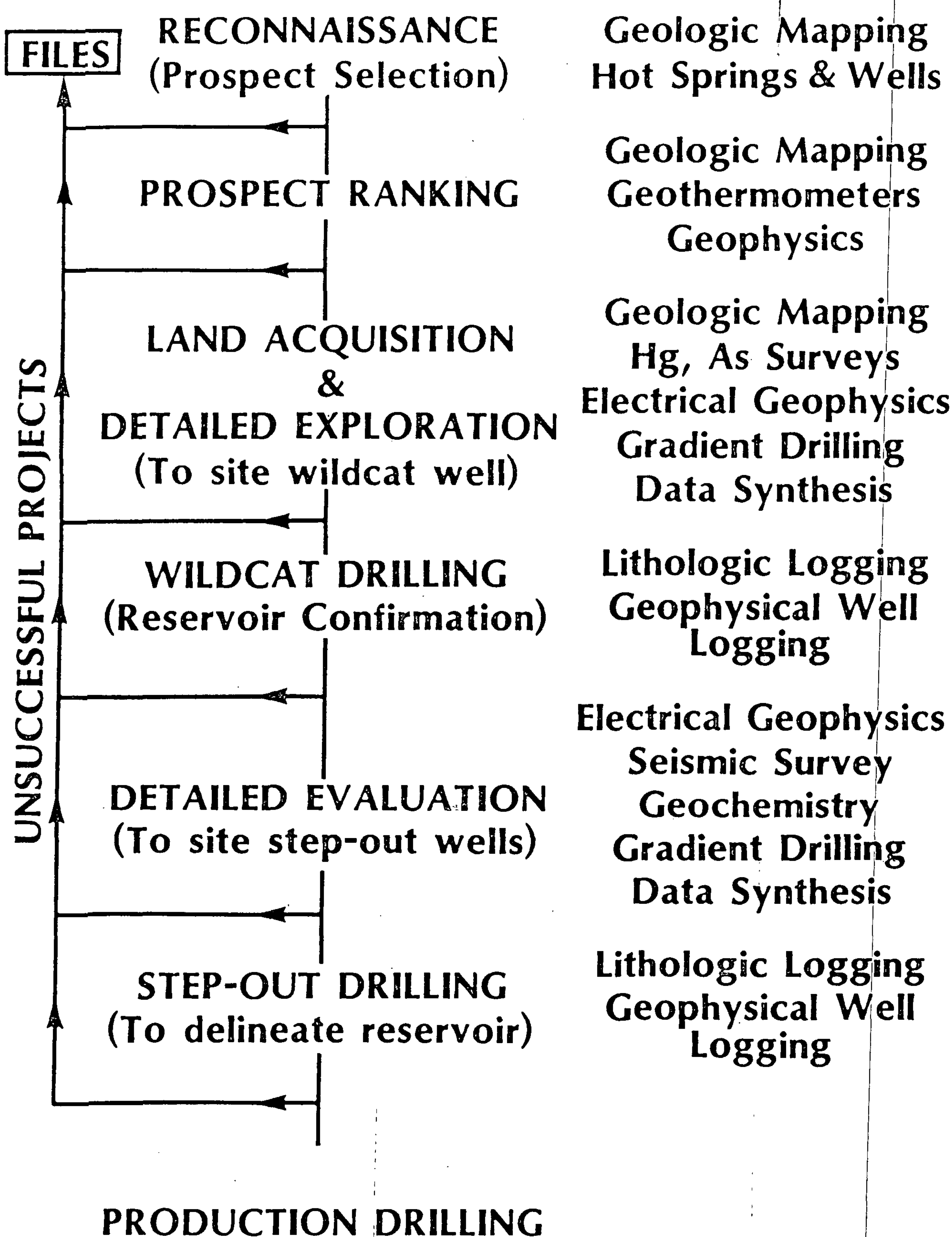


Figure 24

assessing its odds for success at each decision point, and then comparing the project to others or other uses of the money and manpower, optimum exploration results and the risks and costs of exploration are minimized.

We assume, as exploration progresses in an area, that several favorable prospect areas will be identified. The relative priorities among these areas for further exploration must always be considered if the exploration program as a whole is to be most cost effective. In the development that follows, we discuss exploration strategies as applied mainly to a single project, but we must always bear in mind that various prospects will 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.

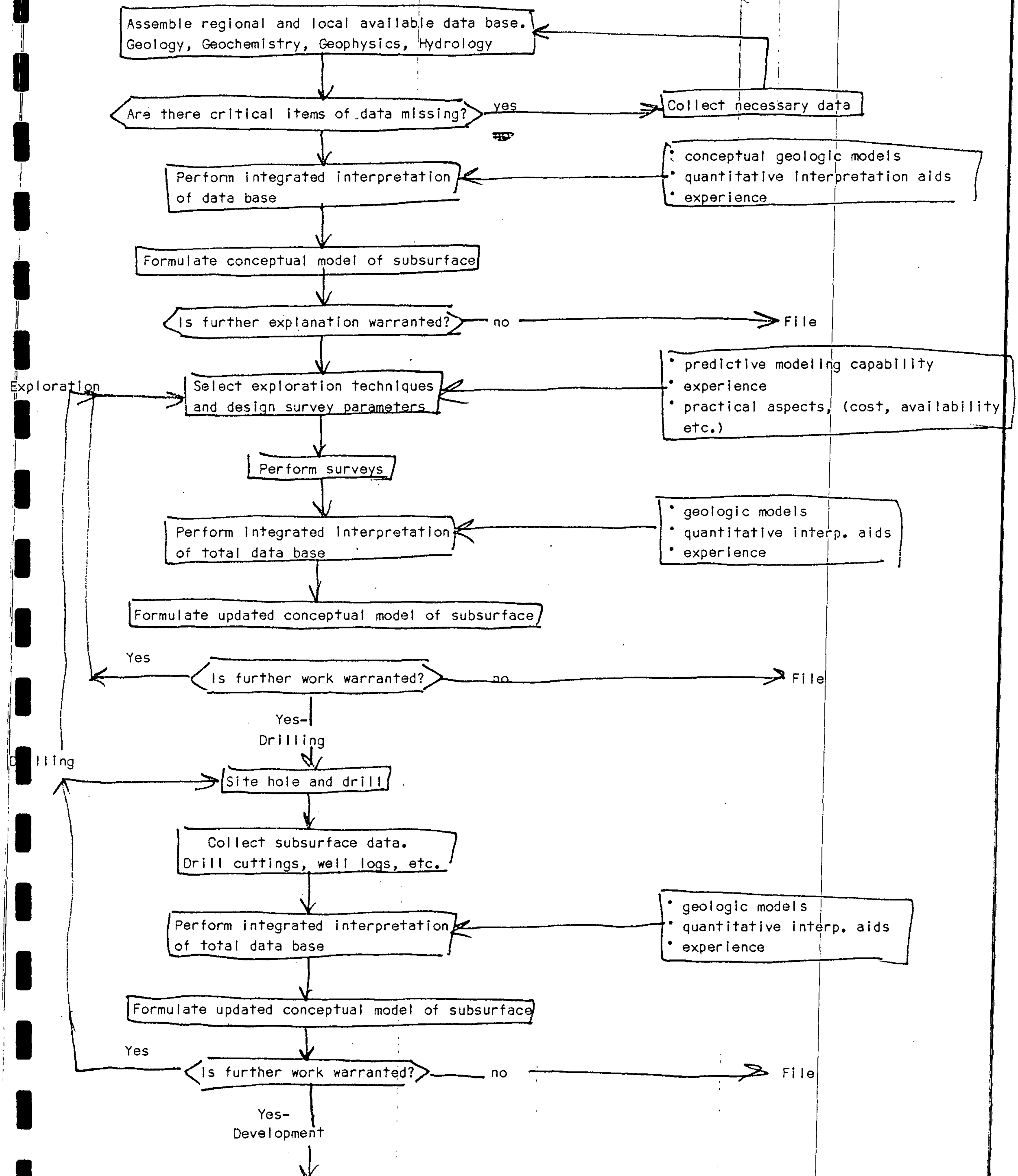
Limitations of Exploration Strategies. It is very important to understand that because geothermal resources are so varied in detail, even within resources of the same general type, it is not possible to specify a certain sequence of exploration techniques that will work or be the most cost-effective in all circumstances. **Stated differently, there is no exploration strategy that can be blindly applied with the expectation of success. The exploration strategy to be followed in any specified area must be designed specifically for application to that area by the geoscientists who are performing the work and interpreting the data.**

Basic Generic Exploration Strategy

Figure 25 illustrates an elaboration of our basic exploration strategy that is generic in that this basic strategy is applicable to all geothermal exploration. It is possible to formulate such a generic strategy precisely because it has no details regarding types of surveys to perform, methods of interpretation, etc. We will discuss the various elements of this strategy individually. The numbers in parantheses below refer to corresponding

Figure 25

BASIC GENERIC EXPLORATION STRATEGY



locations on Figure 25.

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, and 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 on it and must agree 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 to conceptualize about the study area; (b) data interpretation aids such as computer modeling programs and type curves for geophysical data and geochemical data; and, hopefully, (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 to answer 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 becomes very useful in formulating questions whose answers will help to establish the presence or absence of a resource. Examples might be the idea that a geothermal resource somewhere in the exploration area should cause a lowering of electrical resistivity, or that if a geothermal resource exists at depth, one might expect to find thermal springs or wells in a certain region. If found, these thermal springs or wells would help confirm the model.

<u>Exploration Techniques and Survey Design</u> (7). There are several important aspects to selection of exploration techniques. First, if geophysical surveys are being considered, there must be some reason to believe that the geothermal system, or some feature associated with the geothermal system, will cause a change in one or more of the basic physical properties that geophysical surveys measure, i.e. density, magnetic susceptibility, electrical resistivity, induced polarization, sonic velocity, etc. 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 must 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 km x 1 km x 1/2 km thick buried 1/2 km 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, then, 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 is optimum, and other survey design 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 an interpretable response from a geothermal system if it exists and thus help locate the hot waters. Also, if no such resistivity response is detected, then the model of the subsurface must be changed accordingly.

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 geophysical data. The geophysicist should interpret the geophysical data in terms of subsurface variation in the physical property being measured, as discussed before in Section VI. This interpretation will naturally contain ambiguity, but through discussion with the geologists, geochemists and hydrologists working on the project, the geological plausibility of the geophysical interpretation should be examined, and the interpretation modified as needed to arrive at the most geologically plausible interpretation.

In order to perform his interpretation task, the geophysicist 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 to a greater degree and should be more quantitatively accurate because of the survey(s).

With an updated model, one is in a position to decide what the next step is (12). 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 (< 300 m) 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, etc. must be carefully considered, as must the need for blow-out prevention equipment.

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 every 3 to 10 meters, and carefully labeled and preserved in sample bags. These will be used to help define lithology, petrography and hydrothermal alteration and for measurement of physical properties. Conventional geophysical well logs should be measured in the hole, with a minimum logging suite probably being temperature, caliper, resistivity and SP. If the well is flowed or if there is a drill-stem formation test, samples of the fluids from the well should be carefully collected and preserved for analysis. Specific geothermal fluid sampling and preservation techniques must be followed if the results are to be reliable. 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 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 again 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).

The Conceptual Geologic Model

We have seen that the process of exploration is essentially one of working in stages to improve a conceptual geologic resource model (Fig. 26). Data for the model come from the fields of geology, geochemistry, geophysics and hydrology. These data are preferably detailed enough to be stated as a function of three space coordinates and of time (x,y,z,t). The conceptual resource model is, in turn, used to make predictions for use in further exploration and, if a resource is discovered, in reservoir engineering and management. These predictions are tested against the growing data base, and the conceptual model is continually refined so that agreement with the data base is optimum.

The details of the conceptual model of the resource may be difficult to document. Such details are to be found on geologic maps and cross sections, in computer-generated models or data bases and in descriptions in reports.

DEVELOPMENT OF RESOURCE MODEL

INPUT ◄── REPEATED COMPARISON TO REFINE MODEL ──► OUTPUT

(OBSERVATIONS COLLECTED ON SURFACE & IN DRILL HOLES)

(PREDICTIONS BY GEOSCIENTISTS & ENGINEERS)

GEOLOGY (X,Y,Z,T)
- ROCK TYPES
- STRUCTURES - FRACTURES, FAULTS
- MINERALOGY
- THERMAL MANIFESTATIONS

GEOCHEMISTRY (X,Y,Z,T)
- FLUIDS- GEOTHERMOMETRY
  ISOTOPE STUDIES
- ROCKS- HYDROTHERMAL ALTERATION
  TRACE ELEMENTS

GEOPHYSICS (X,Y,Z,T)
- ELECTRICAL- SEISMIC
- MAGNETIC - HEAT FLOW
- GRAVITY - WELL LOGGING

HYDROLOGY (X,Y,Z,T)
- REGIONAL AND LOCAL
- WELL TEST DATA

CONCEPTUAL RESOURCE MODEL* (X,Y,Z,T)

EXPLORATION (X,Y, Z,T)
- BEST WELL LOCATIONS
  (PRODUCTION, INJECTION)
- TECHNIQUE RESPONSES
- GENESIS

RESERVOIR ENGINEERING/ MANAGEMENT (X,Y,Z,T)
- POROSITY, PERMEABILITY
- FLUID FLOW
- WELL PRODUCIBILITY
- DECLINE , LONGEVITY
- RECHARGE
- REINJECTION/ INJECTION
- OPTIMUM FIELD DEVELOPMENT
- WELL COMPLETION
- CHEMICAL INTERACTION

X,Y,Z, = SPACE

T = TIME

* MODEL EXISTENCE:

MAPS
SECTIONS
COMPUTER
DOCUMENTS

} INTEGRATED IN THE MINDS OF A FEW GEOSCIENTISTS AND ENGINEERS WORKING TOGETHER

Figure 26

Basically, the best and most useful versions of the model will exist in the minds of the geoscientists and engineers working on the project.

We have seen that there is no single geological model that can be applied to all geothermal resources. It is, therefore, imperative that the regional and local geology is well understood in order that the opportunity for discovery can be evaluated. Not every geothermal system has manifestation of its existence at the surface that is obvious enough to lead to easy discovery. It is the job of the explorationist to observe, measure and correctly interpret subtle geological, geochemical, geophysical and/or hydrological signs of a geothermal reservoir at depth and to help prescribe a drilling and well testing program that will lead to discovery.

Siting successful geothermal wells is far from easy. Even within a well-known geothermal area such as The Geysers, Calfornia, where the experience of locating and drilling approximately 700 wells is available, the success rate for production is only about 80 percent. For wildcat geothermal drilling in relatively unknown areas the success ratio is much lower -- about 15 percent for the Basin and Range Province of the western United States. The low success rate revolves not so much around finding heat as it does around finding fluids in producible amounts that are sufficient to supply a utilization system and to pay for well drilling, testing and maintenance. In many geothermal reservoirs, this means drilling into one or more fractures that are connected to other fractures and permeable horizons within the reservoir and to the ultimate source area for the geothermal fluids. There is ample evidence in numerous articles that fractures of the order of millimeters in aperature can support sufficient fluid flow to make a producer from an otherwise unsuccessful well. Although large blocks of rock in nature are nearly all cut by fractures and faults that vary in spacing from centimeters to tens of meters, most of these fractures do not persist far 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 hot fluid. Because there is no known way to detect from the surface the particular, narrow fractures that carry geothermal fluids at depths of hundreds to thousands of meters, exploration techniques are indirect and usually provide only circumstantial evidence of the existence and location of the reservoir.

Recommended Exploration Strategy-Sedimentary Basins in Spain

Without a detailed knowledge of the geologic setting, our recommended exploration strategy must be generalized. Our data base and knowledge of the resource areas are less complete than those of the IGME staff. Our recommended exploration strategy for sedimentary basin areas is outlined in Figure 27. This strategy focuses on geoscience and assumes potential users for moderate temperature (i.e. not electric quality) resources are located in close proximity to the resource. Although the strategy clearly indicates an interdisciplinary exploration effort, this report is directed toward the geophysical methodology.

The compilation of regional geologic, geophysical, thermal gradient, thermal spring and oil/water well data is essential for an initial regional scale evaluation of resource potential. The IGME has recognized this need and already completed this step of the recommended strategy. Following the identification of probable resource types and areas of occurrence, it is necessary to identify key missing elements in the regional scale data base (Step 2). At this stage of exploration, large basin areas are still being considered and it is important to develop an understanding of basin hydrology, fluid geochemistry and stratigraphy. To the extent possible existing boreholes should be utilized to obtain thermal gradient, stratigraphic, hydrologic and geochemical

IGME - EXPLORATION STRATEGY - BASINS

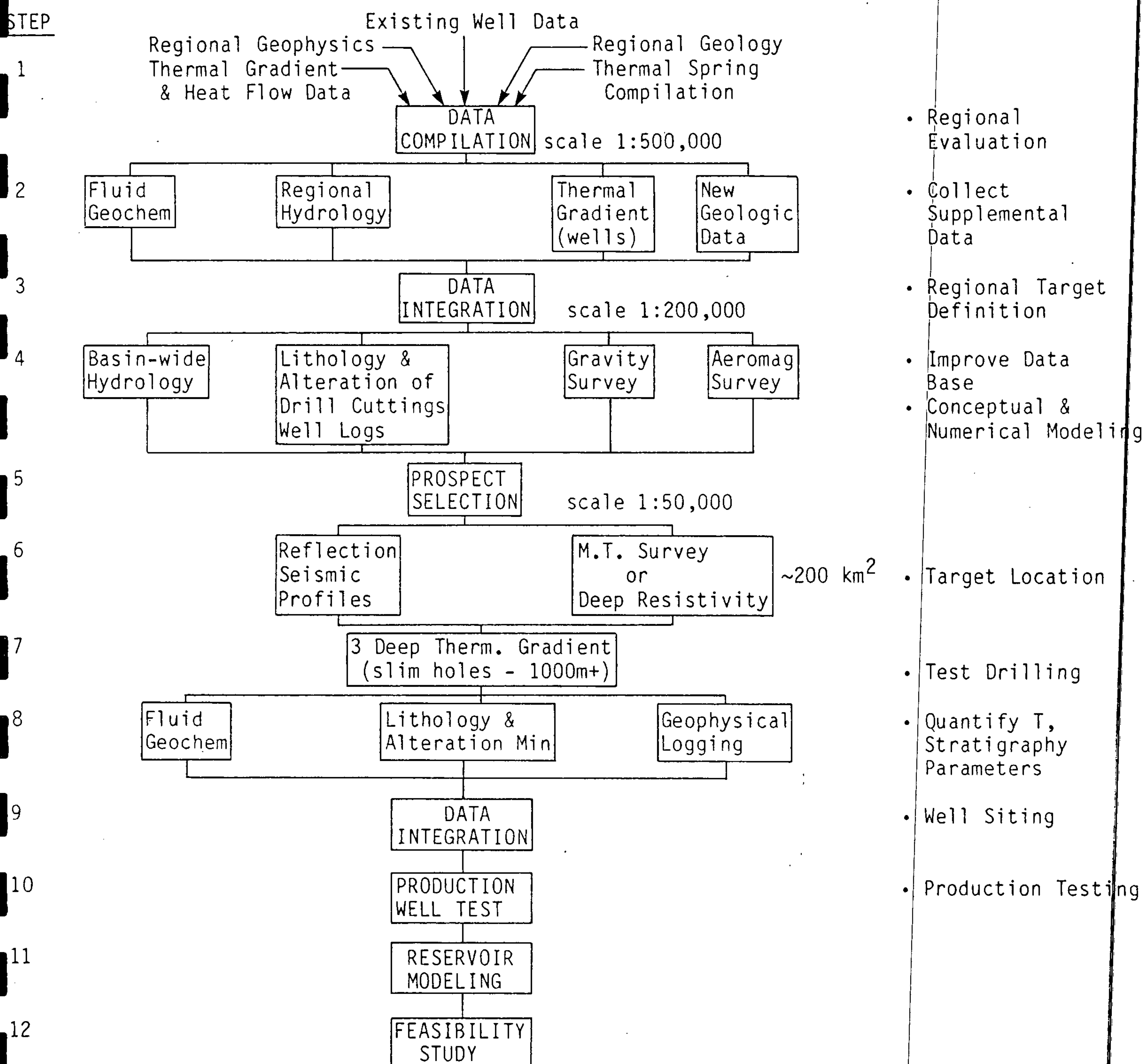


Figure 27

data. New geophysical surveys should be deferred for the present.

Step 3 calls for a period of study and integration of the improved data base. The output for this effort should be the identification of smaller, more promising subareas within the basin, and the identification of new data needs to be obtained in Step 4.

In the basin environment, reconnaissance geophysical methods can be used in a highly effective manner. Aeromagnetic and gravity data should be strongly considered as techniques to map the borders of basin areas, the thickness of sedimentary fill and the position of structures which may be important as conduits for geothermal fluids. The magnetic data may also map the presence of volcanic flows, dikes and intrusive bodies along the margins of the basins. An evaluation of possible reservoir units would be enhanced by lithologic and alteration studies of existing well cuttings and of geophysical well logs. The interpretation of the gravity and magnetic data is enhanced by detailed numerical modeling and correlation with the geologic data. Computer-based modeling techniques should be used to the extent possible.

The integration and interpretation of new data (Step 5) from Step 4 will lead to the selection of definite prospect areas, appropriate for study at an expanded scale, perhaps 1:50,000. It is now appropriate to consider the cost effectiveness of the more specific and costly geophysical methods. The reflection seismic method offers the most precise method for mapping horizons and faults within the basin environment. The typical high cost of the surveys, $5,000-$10,000 per line mile, may be inconsistent with the development of low- to moderate-temperature resources for direct heat application. The economies of the anticipated end use of the geothermal resource must be considered to determine if this expensive step is warranted. The optimum siting of deep thermal gradient holes to follow however, may warrant this cost. Needless to say, more specific target definition will be required for drill holes intended to intersect faults than for those aimed at stratigraphic horizons. Magnetotelluric (MT) or some other form of deep resistivity mapping (VES, bipole-dipole, TDEM) may be appropriate if low resistivity thermal fluids are expected, and if the sedimentary stratigraphy (as indicated by cuttings and geophysical logs) indicates suitable physical property contrasts. Computer modeling to determine expected anomaly magnitude should be undertaken before the decision is made to proceed with these geophysical techniques. Detailed interpretation of Step 6 data should focus on the siting of thermal gradient/stratigraphic test drilling.

The drilling of new wells for thermal gradient/heat flow and stratigraphic information has been delayed until now, in our recommended basin strategy, because of the high cost of drilling. It is important to realize that, in a complex basin environment, the drill hole may test a very small area and may not properly test the resource potential. Thus, we envision a deep thermal gradient program which would include a minimum of three slim holes to depths consistent with the geologic target and basin hydrology, perhaps of the order of 1000 m deep.

Step 8 attempts to maximize the geologic information available as a result of the thermal gradient drilling. Fluid chemistry, lithology and alteration mineralogy, and geophysical logging will aid in the quantification of the stratigraphy, and thermal resource potential. Subsequent efforts in the exploration strategy are straightforward. A re-evaluation of the total data base in light of the slim hole results may or may not require a substantial effort. The well-log data may enhance earlier seismic or electrical interpretation. An interim evaluation of resource potential should be completed and the probability of success (temperature, quality and quantity of fluids) weighed against the cost of completing the exploration and development program. If this evaluation is favorable, a deep production well would be sited from the available data base.

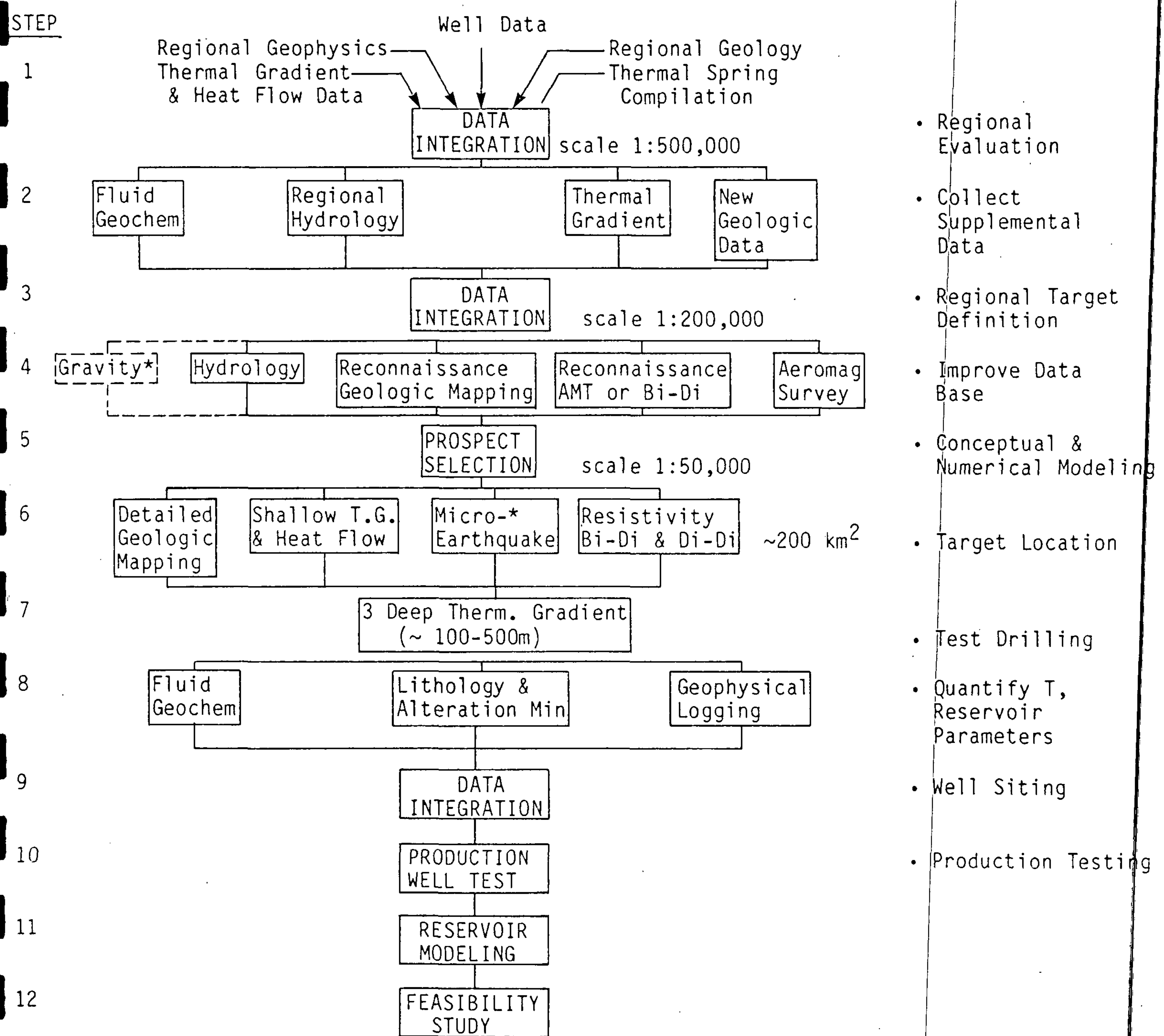
Additional geophysical work includes geophysical logging and interpretation for the production well test. More detailed reflection seismic work may be cost effective if the development of a major geothermal resource is indicated. The cost/benefit of additional geoscience work and resource development must be evaluated by the knowledgeable workers for each resource area.

The basin exploration strategy presented here is somewhat idealized and simplified. The principal geophysical elements have been identified, but may not always be cost effective. Some survey techniques have been ignored because in our judgement they lack the spatial resolution, physical property contrast, or probable cost-effectiveness desired in the typical basin environment. Among these techniques considered, but not included in our strategy, are the microearthquake, seismic emissions, self-potential and detailed shallow resistivity methods. Particular geologic conditions may make one or another of these methods useful in isolated instances. Variations to the recommended strategy may best be addressed by those scientists more familiar with the local geology, expected resource type, and intended utilization.

Recommended Exploration Strategy-Igneous/Fault Resources in Spain

Figure 28 presents our recommended generalized exploration strategy for the igneous (granitic) areas. Many elements of the strategy are common to the corresponding basin strategy, especially in the early part of the evaluation (Steps 1 through 3). The IGME has already completed many aspects of the preferred strategy shown here. The strategy addresses large areas of granitic and other crystalline and metamorphic rocks where identified thermal springs are associated with faults and fractures, and little interstitial porosity and

IGME - EXPLORATION STRATEGY - IGNEOUS/FAULT RESOURCES



* Included only if appropriate for geologic setting

Figure 28

permeability are present.

Step 4 addresses the acquisition of additional reconnaissance-scale geologic and geophysical data. Aeromagnetic data must be considered as perhaps the most cost-effective method to identify different intrusive or metamorphic units and geologic structures which cut across them. The desirability of gravity data should be considered but the method would not routinely be included due to the need for precise terrain corrections and elevation control, and the probable low density contrasts. Reconnaissance electrical surveys such as AMT or bipole-dipole would perhaps be appropriate for the location of any large areas of hydrothermal alteration and increased fluid content related to major fracture zones. A possible alternate electrical resistivity mapping approach would be electromagnetic profiles run perpendicular to already identified fracture zones.

Additional geophysical work would be conducted at the prospect exploration stage, for areas a few hundred km^2 in size. These methods complement and extend the geologic mapping effort. A microearthquake survey may be appropriate for the identification of active structures. The seismicity in many areas is episodic, however, so a commitment to a long (6 months) monitoring period and operation of many (4-12) recording seismographs, would be required. A lesser commitment could be misleading and counterproductive. A more detailed electrical resistivity survey may be warranted, but the cost-effectiveness would have to be evaluated in view of the local geology. The bipole-dipole or dipole-dipole arrays should be considered for delineation of alteration areas and fracture zones. Vertical loop electromagnetic profiles may map these features at shallow-moderate (100-500 m) depths.

A program of shallow thermal gradient/heat flow drilling would be recommended in this dominantly conductive gradient environment. A program of 10 to 20 holes of 30-100 m depth may be appropriate in many igneous environments.

The integration and interpretation of these prospect scale data would generally be adequate for the selection of the deep thermal gradient program.

Deep thermal gradient drilling (Step 7) is recommended to test the validity of the most promising shallow gradients in areas found to be favorable by other data sets. The depth of the holes will depend on the local hydrology and indicated resource type.

Other geophysical aspects of this strategy follow the basin strategy, as modified by drilling results and the differences in geologic environment.

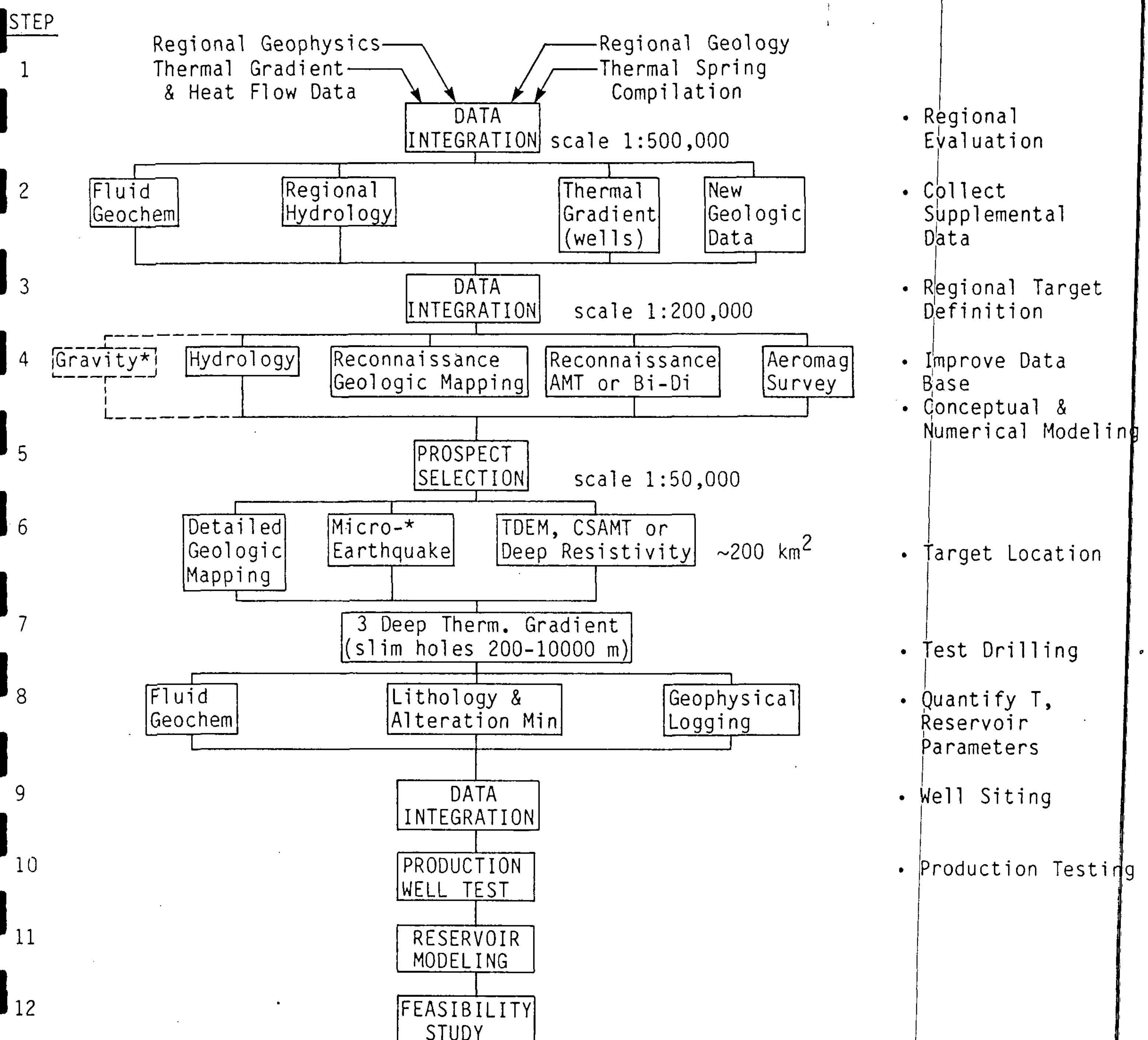
Recommended Exploration Strategy - Volcanic Areas in Spain

Figure 29 diagrams the generalized strategy recommended for volcanic area geothermal resources. The role of geophysics is similar to that for igneous/ fault resources.

Step 4 indicates the probable use of aeromagnetic and reconnaissance resistivity surveys for the delineation of rock units, geologic structures, and possibly the detection of low resistivity fluids and/or alteration. The gravity method should be considered, cognizant of the need for precise elevation control and terrain corrections. The association of a deep thermal source with a young, low density intrusive would encourage the inclusion of gravity surveys. Without this incentive, gravity surveys may not be effective in a volcanic environment.

Geophysical exploration at the prospect scale, Step 6, may vary from the igneous and basin programs. Time domain electromagnetics (TDEM) or controlled source AMT (CSAMT) may provide cost-effective alternatives to galvanic resistivity in areas of high surface impedance. Microearthquake surveys should be considered but would not routinely be included in volcanic area exploration efforts. Shallow thermal gradient data would likely be disturbed by the near

IGME - EXPLORATION STRATEGY - VOLCANIC AREAS



* Included only if appropriate for geologic setting

Figure 29

surface hydrologic regime and, in general, would be deleted in favor of an expanded deep thermal gradient program.

Thermal gradient holes should be drilled deep enough to penetrate the water table and the near surface hydrologic regime. These holes should provide considerable information on potential reservoir porosities and on volcanic lithology.

We emphasize again the very site specific nature of the exploration strategy. The geophysical strategies recommended here are at best a generalization of the approach we would consider and then evaluate, for each of the specific resource types. The careful integration of all the data, geologic, geophysical, and geochemical is essential for an effective resource exploration program.

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APPENDIX I

PHYSICAL PROPERTIES OF GEOTHERMAL SYSTEMS

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VOL. 82, NO. 36 JOURNAL OF GEOPHYSICAL RESEARCH DECEMBER 10, 1977

A PRELIMINARY ANALYSIS OF INTRINSIC FLUID AND ROCK RESISTIVITY IN ACTIVE HYDROTHERMAL SYSTEMS

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Abstract. Electrical resistivity data are utilized in interpretations of subsurface environments and to explore for geothermal and mineral resources. Abnormally low resistivity data are alternatively interpreted to indicate the presence of high-temperature fluids or conductive minerals (metal sulfides) at depth, even though relative contributions of thermal, porosity, and fluid composition effects appear to be poorly known. An analysis of intrinsic rock resistivities, calculated electrical porosities, and two-dimensional heat and mass transfer computations indicates that the host rock resistivity distribution around igneous intrusives is directly related to the mode of dispersion of thermal energy away from the pluton. Comparisons between numerical results and field observations in geothermal areas indicate that resistivity values in the vicinity of thermal anomalies are a complex function of fluid circulation patterns, fluid composition, and the distribution of conductive minerals produced by the reaction between circulating fluids and rocks; therefore in many cases, low near-surface resistivity anomalies cannot be entirely accounted for by hot circulating saline fluids, and observations of high thermal gradients associated with low-resistivity anomalies are not unique indications of a high-energy geothermal resource at shallow crustal depths.

Introduction

The nature of rocks in the upper crust is often deduced from apparent electrical resistivity data. The relationship between these data and the intrinsic resistivities is poorly known, and therefore correlation of the electrical resistivity measurement of rocks with variations in rock and pore fluid properties is usually speculative. Although interpretations are based on resistivity data measured in deep drill holes and laboratory measurements on rocks and fluids, the correlation of laboratory measurements, even in well-controlled laboratory experiments, with rock properties has not been satisfying. Better understanding of this correlation would facilitate mapping subsurface conditions with the aid of electrical survey data and is particularly relevant in regions of active hydrothermal activity, where there is considerable interest in energy resources.

The electrical resistivity variations in upper crustal rocks have been inferred from various electrical methods. The results of these surveys indicate that average resistivity values in stable crustal regions range from 10^5 to 10^2 ohm m [Keller and Frischknecht, 1966; Keller et al., 1966]. An analysis of laboratory experimental data on the resistivity of fluid-saturated crustal rocks coupled with considerations of regional heat flow data predict similar ranges in resistivity to a 40-km depth [Brace, 1971]. Resistivity surveys in regions of geothermal activity indicate anomalously low resistivities, which range between 10 and 100 ohm m [Sato, 1970; Cheng, 1970; Risk et al., 1970; Zohdy et al., 1973; Keller, 1970]. These anomalous values are often attributed to the presence of prospective thermal energy resources.

The properties and conditions in geothermal systems which contribute to resistivity values are fluid and mineral composition, porosity, temperature, and pressure [Brace, 1971; Brace and Orange, 1968; Brace et al., 1965; Keller and Frischknecht, 1966]. The effect of porosity and fluid resistivity on the bulk rock resistivity of sedimentary rocks was deduced by Archie [1942] and extended to crystalline rocks by Brace et al. [1965]. The empirical relationship derived by Archie defines bulk rock resistivity ρ_r as

$$\rho_r = a\rho_f\phi^{-n} \qquad (1)$$

in terms of pore fluid resistivity ρ_f, a proportionality constant a, porosity ϕ, and a factor which depends on the degree of rock consolidation, n. Experimental data by Brace et al. [1965] and Brace [1977] suggest that for fractured media, a = 1 and n = 2, values which apparently agree with theoretical electrical network models of Greenberg and Brace [1969] and Shankland and Waff [1974]. The porosity value normally used in (1) is that of total rock porosity. However, only those pores which contribute to current flow should be included in this term, and in fractured media the total porosity is usually not totally composed of interconnected pores, as is indicated by studies of ion transport in these types of rocks [Norton and Knapp, 1977]. Ranges in rock resistivity of 6 orders of magnitude may be realized for reasonable variations in the abundance of interconnected pores in fractured media [Moskowitz, 1977].

The transient thermal history of rocks in hydrothermal systems related to cooling igneous bodies has been simulated, over large regions and for long time periods, by numerical methods [Norton and Knight, 1977]. Since the variation in resistivity of rocks relates directly to subsurface temperature and pressure conditions, their numerical models provide a basis with which to analyze intrinsic resistivity of hydrothermal systems. The purpose of this communication is to present the results of a first-order approximation to the nature of intrinsic resistivity in such systems. The study considered variations in permeability and porosity, heat sources, rock and fluid properties, including variation in pore fluid resistivity as a function of temperature, pressure, and concentration of components in solution, as well as the time dependence of these parameters in a two-dimensional domain.

Porosity

The distribution of porosity in the crust varies in response to changes in pore fluid pressure. Effective pressure P_e is the difference between confining pressure P_c and pore fluid pressure P_f:

$$P_e = P_c - P_f \tag{2}$$

The variation of effective pressure with depth in the crust shows that in many geologic environments, increases in pore fluid pressure, as a result of temperature increases, will cause the effective pressure to decrease [Knapp and Knight, 1977]. As a consequence of the low tensile strength of rocks, when effective pressure is reduced to zero, the rock will fracture. Thus an increase in porosity is expected at zero effective pressure.

The total porosity in fractured media may be represented by

$$\phi = \phi_F + \phi_D + \phi_R \tag{3}$$

where ϕ_F, the effective flow porosity represents those pores through which the dominant mode of fluid and aqueous species transport is by fluid flow, ϕ_D, the diffusion porosity, represents those pores through which the dominant mode of transport is by diffusion through the aqueous phase, and ϕ_R, the residual porosity, represents those pores not connected to ϕ_F or ϕ_D. Field observations and experimental studies indicate that ϕ_R apparently accounts for more than 90% of the total porosity observed in crystalline rocks at ambient conditions [Norton and Knapp, 1977]. Our studies indicate that when ϕ_R values are used in (1), intrinsic resistivities of saturated rocks are predicted reasonably well, whereas ϕ_F and ϕ_D predict values many orders of magnitude higher than observed values [Moskowitz, 1977].

The correlation between porosity values consistent with electrical diffusivity and ion diffusivity determined by Norton and Knapp [1977] is unclear. We have assumed that pore fluid thermal expansion in residual pores produces fractures which contribute to electrical porosity. The pore characteristics at which the fluid will simply flow from the pore in response to thermal expansion and not increase the porosity are considered to be typical of the flow porosity normally found in crystalline rocks. Therefore any increases in total porosity due to temperature occur approximately as the result of changes in residual porosity. These assumptions are justified by the fact that $\phi_R \simeq 0.9\phi$ and that ϕ is used in Archie's law. Residual porosity is certainly an upper limit to the actual electrical porosity of crystalline rocks, and subsequently, the intrinsic resistivity calculated from these assumptions represents minimum values. Also, conversion of residual porosity to flow or diffusion porosity which relates directly to permeability is not considered in the fluid flow models to be discussed below.

The concept presented by Knapp and Knight [1977] can be used to relate porosity change at zero effective pressure to temperature. The total derivative of the rock-pore volume at constant composition is

$$dV = \left(\frac{\partial V}{\partial T}\right)_P dT + \left(\frac{\partial V}{\partial P}\right)_T dP \tag{4}$$

where $V = V_r + V_f$, V_r is rock volume, and V_f is pore volume. The coefficients of isobaric thermal expansion α and isothermal compressibility β for the bulk rock are defined as

$$\alpha \equiv \frac{1}{V}\left(\frac{\partial V}{\partial T}\right)_P \tag{5a}$$

$$\beta \equiv -\frac{1}{V}\left(\frac{\partial V}{\partial P}\right)_T \tag{5b}$$

Substitution of (5a) and (5b) into (4) defines the total volume change in terms of α and β:

$$dV = V\alpha\, dT - V\beta\, dP \tag{6}$$

This total derivative can also be expressed in terms of the individual thermal expansions and compressibilities of pore fluid and rock:

$$dV = [V_f\alpha_f + V_r\alpha_r]\, dT - [V_f\beta_f + V_r\beta_r]\, dP \tag{7}$$

However, when rocks fracture as a consequence of pore fluid expansion, infinitesimal increases in pore fluid pressure will produce further fracturing. Therefore $dP \cong 0$, and (7) may be simplified to

$$dV = [V_f\alpha_f + V_r\alpha_r]\, dT \tag{8}$$

Typical values for α_r, for common silicate minerals, over a temperature range of 0°-800°C, are of the order of 10^{-6} °C^{-1} [Clark, 1966]. The thermal expansion coefficient for pure water, over the same temperature span, is of the order of 10^{-3} °C^{-1}. As long as pore volume V_f is greater than or equal to 0.01, $V_f\alpha_f \gg V_r\alpha_r$, and (8) becomes

$$dV = V_f\, \alpha_f\, dT \tag{9}$$

The total volume change, according to (9), occurs as a result of pore volume changes, the rock volume remaining essentially constant. Rearranging (9) with the approximation that $dV \simeq dV_f$ yields an integral equation relating pore volume and temperature:

$$\int_{V_f^o}^{V_f} \frac{dV_f}{V_f} = \int_{T_b}^{T} \alpha_f(T)\, dT \tag{10}$$

In (10), V_f^o is the initial residual pore volume, and T_b is the temperature at which the rock fractures. Integrating (10) gives the pore volume as a function of temperature:

$$V_f = V_f^o \exp\left[\int_{T_b}^{T} \alpha_f(T)\, dT\right] \tag{11}$$

where $T > T_b$. The initial residual porosity ϕ_R^o is defined as

$$\phi_R^o = \frac{V_f^o}{V^o} = \frac{V_f^o}{V_r + V_f^o} \tag{12}$$

and the fluid and rock volumes are

$$V_f^o = \phi_R^o V^o \qquad V_r = (1 - \phi_R^o)V^o \tag{13}$$

respectively. Substituting (12) and (13) into (11) defines a porosity temperature function in terms of the initial residual porosity:

$$\phi_R = \frac{\phi_R^o F(T)}{(1 - \phi_R^o) + \phi_R^o F(T)} \tag{14}$$

where

$$F(T) = \exp\left[\int_{T_b}^{T} \alpha_f(T)\, dT\right] \tag{15}$$

For the purposes of this discussion we will consider that (14) defines increases in the effective electrical porosity. That is, all porosity increases due to thermal effects are assumed to contribute to increased electrical current flow in the rocks.

The temperature at which the rocks initially fracture, T_b, may be defined as

$$T_b = T_a + \Delta T \tag{16}$$

where T_a is the ambient temperature and ΔT is the temperature increment necessary to reduce effective pressure to zero. The value of ΔT depends on the geothermal gradient, and the maximum value of ΔT along a gradient of 20°C/km is 20°C [Knapp and Knight, 1977].

Porosity, defined by (14), was computed for depths of 1, 2, 3, and 4 km below the earth's surface (Figure 1). At a depth of 1 km and initial temperature of 40°C, large increases in porosity are predicted for temperature changes of the order of 300°C. However, the porosity increases are small for changes in temperatures of less than 100°C at this same depth. At greater depths, e.g., 4 km, much smaller increases in porosity are predicted for these same temperature conditions, owing to increased confining pressure.

The relationship among bulk rock resistivity, fluid resistivity, and electrical porosity is

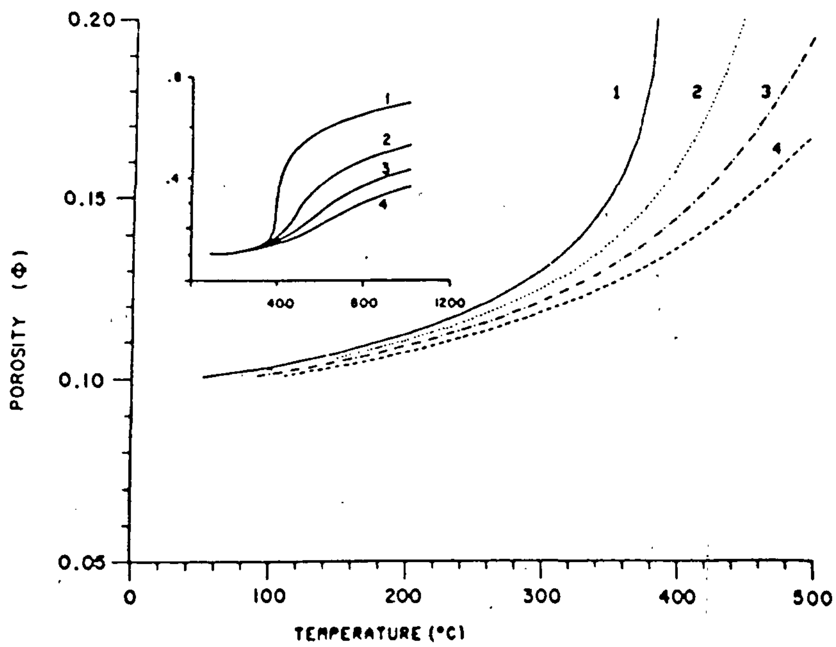

Fig. 1. Porosity as a function of temperature at depths of 1, 2, 3, and 4 km, as computed from (14). Initial porosity ϕ_o is 0.1, and background temperatures are consistent with a surface temperature of 20°C and a temperature gradient of 20°C/km. Pressures were computed for a rock density of 2.75 g/cm³. Insert shows porosity values consistent with temperatures up to 1200°C.

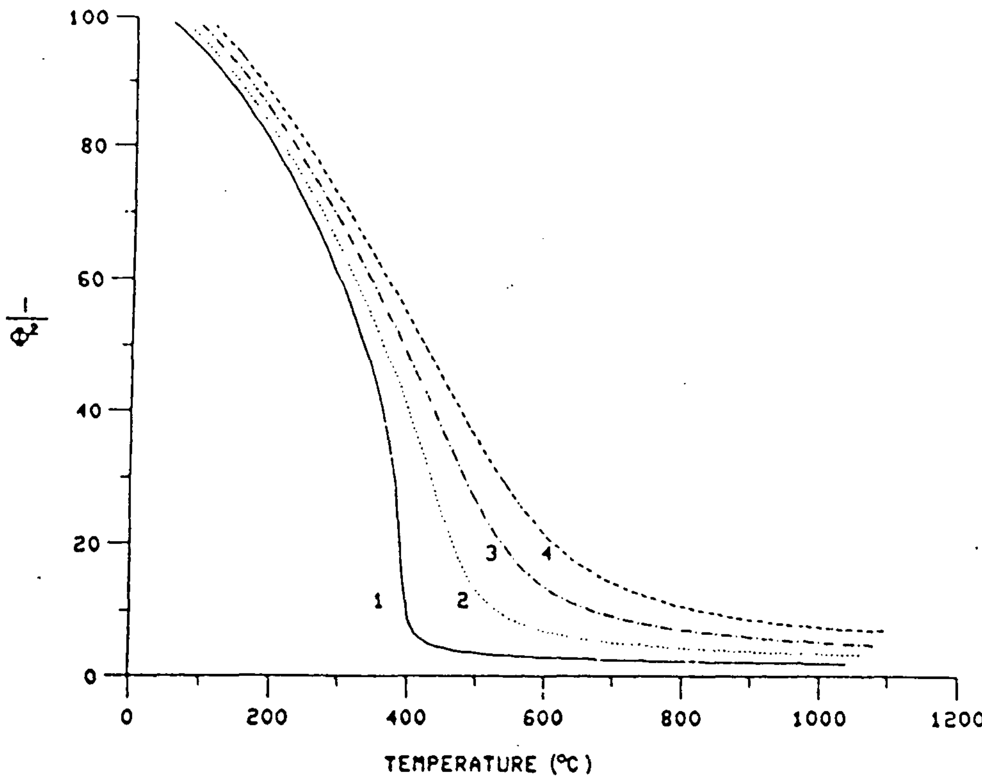

Fig. 2. Parameter ϕ^{-2} as a function of temperature at depths of 1, 2, 3, and 4 km. Parameters used are same as those in Figure 1.

poorly known. Archie's law, equation (1), has been assumed as an adequate first approximation to rock resistivity. Therefore the important parameter in predicting resistivity from (1) is ϕ^{-2}, and therefore small increases in porosity will result in a significant decrease in ρ_R (Figure 2).

Fluid Resistivity

The resistivity of natural groundwaters varies as a function of temperature, pressure, and composition. Since the dissolved constituents in natural waters are often dominated by sodium and chloride, and the resistivity values of $NaCl-H_2O$ fluids are similar, within a factor of 2.5, to those of other common fluids, the compositional effects of fluid resistivity are approximated by the system $NaCl-H_2O$ [Quist and Marshall, 1968; Chambers, 1957; Gunning and Gordon, 1942]. The variation in resistivity of a 0.1 m NaCl solution with temperature and pressure exhibits a steady, pressure independent decrease in resistivity to approximately 300°C, then an order of magnitude increase to 12 ohm m at 500°C and 500 bars (Figure 3). As can be seen, the dominant pressure effect is to shift the resistivity minimum to higher temperatures with increasing pressure. Increasing the NaCl concentration results in a decrease in resistivity that varies from 100 to 0.01 ohm m for concentrations ranging from 10^{-4} to 2 m.

Fluid temperatures in geothermal reservoirs range up to 300°C, and pressures to 1 kbar. Total ionic strength of these fluids ranges from 1 m, such as was observed in the Imperial Valley system [Meidav and Furgerson, 1972] to 10^{-2} m, such as was observed in the Broadlands, New Zealand, system [Browne and Ellis, 1970]. Typical resistivities of geothermal reservoir fluids range from 0.01 to 10 ohm m [Cheng, 1970], which is similar to the range in resistivity of pore fluids in a variety of geologic environments [Keller and Frischknecht, 1966].

Temperature-Pressure Distribution

Notions of temperatures and pressures in geothermal systems are primarily derived from production or exploration wells, and consequently, information is restricted to small portions of the total system. Knowledge of these parameters over the entire hydrothermal system is necessary in order to analyze the time dependence of resistivity in the region of a cooling pluton. Simulation of cooling plutons by numerical methods is one method by which these parameters can be defined for an idealized geothermal system.

Fluid flow caused by thermal anomalies related to igneous plutons is effectively scaled and represented in two dimensions by partial differential equations which describe the conservation of mass, momentum, and energy for the fluid-rock system [Norton and Knight, 1977]:

Conservation of energy

$$\gamma\frac{\partial T}{\partial t} + q\nabla H = \nabla\cdot\kappa\nabla T \qquad (17)$$

Conservation of momentum

$$\nabla\cdot\frac{\nu}{k}\nabla\Psi = R\frac{\partial\rho}{\partial y} \qquad (18)$$

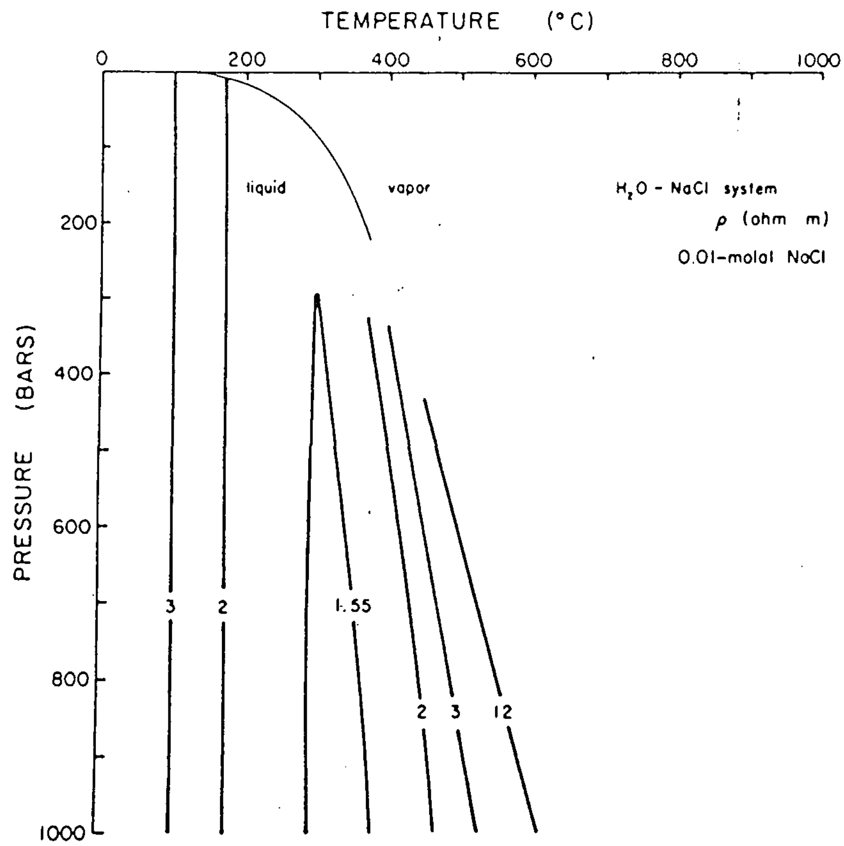

Fig. 3. Temperature-pressure projection of the two-phase surface (liquid and vapor) in the H_2O-NaCl system at 0.01-m NaCl concentration depicting fluid resistivity isopleths.

where T is the temperature, Ψ is the stream function, q is the fluid flux, t is the time, H, ρ, and ν are the enthalpy, density, and viscosity of the fluid, k is the permeability of the rock, κ is the thermal conductivity, and γ is the volumetric heat capacity of the fluid-saturated media, R is the Rayleigh number, ∇ is the gradient operator, and y is the horizontal distance in the two-dimensional section to which these equations apply.

Equations (17) and (18) are approximated by finite difference numerical equations which permit computation of the values of the dependent variables at discrete points in the domain from initial and boundary values specified for the system. The numerical analysis provides the option to include variable transport properties of the fluid (H_2O system) and rock, general boundary and initial conditions, and radioactive and volumetric heat sources in a two-dimensional domain. The transport process related to the transient thermal anomaly is approximated by a time sequence of steady state numerical solutions to (17) and (18), computed at explicitly stable time intervals. An alternating direction implicit finite difference method is used to approximate the spatial derivatives at discrete intervals of the order of 0.1-0.5 of the system height. Fluid pressure in the system is computed at each steady state step by integration of Darcy's law, in which the fluid properties, viscosity and density, are expressed as a function of temperature and pressure.

The methods used by Norton and Knight [1977] were used to define the temperature variation in the environment of a cooling pluton as a function of time. The hypothetical system is characterized by a dominance of convective heat transport over conductive heat transport as a result of relatively large host rock permeabilities (Figure 4). As a consequence of fluid circulation the temperature distribution in the host rocks evolves into a plumose pattern at $\sim 10^5$ years (Figure 5) and results in broad regions of uniform temperature above the pluton.

Initial temperatures in the host rocks at this depth are 110°C, as defined by the 20°C/km geothermal gradient and 20°C surface temperature. At 190,000 years after pluton emplacement the 200°C isotherm is at approximately a 0.5-km depth (Figure 5), and the temperatures between the top of the pluton and the 200°C isotherm have increased by at least 90°C. The pore fluid resistivity reaches a minimum at temperatures between 200°C and 300°C (Figure 3), and the porosity increase defined by (14) is of the order of 15% of the initial value for temperature increases of 100°-200°C. Therefore the zone between the 200°C isotherm and the top of the pluton in the system will be characterized by maximum porosity increase and the maximum decrease in pore fluid resistivity.

Porosities in host rocks at depths of <2 km directly over the pluton have significantly increased approximately 20% of the initial value at 190,000 years after pluton emplacement. This porosity increase persists uniformly to a 4-km depth. Time variations in porosity, calculated

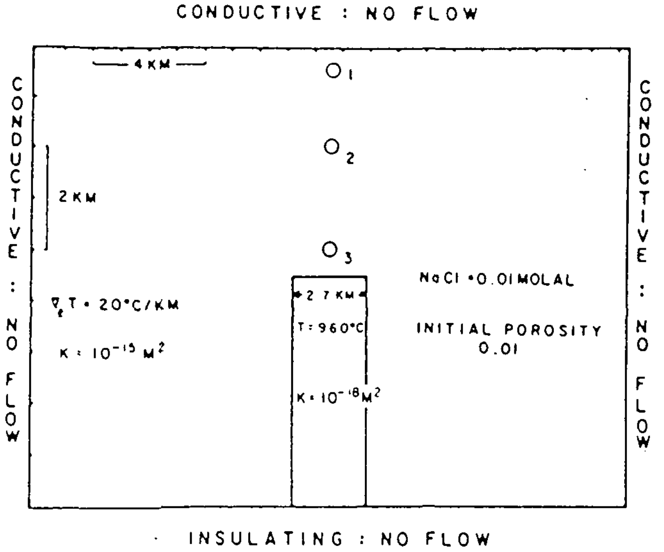

Fig. 4. Two-dimensional cross section of a pluton intruded into uniform permeability host rocks depicting initial and boundary conditions for numerical simulation of heat and mass transfer. Domain was discretized into 120 grid points such that Δz = 0.9 km and Δy = 1.75 km. The initial conditions include a background temperature consistent with a surface temperature of 20°C and a thermal gradient $\nabla_z T$ = 20°C/km. The pluton's initial temperature is 960°C. Permeabilities are 10^{-15} m^2 and 10^{-18} m^2 for the host and pluton rocks, respectively.

at fixed points 4, 2.5, and 0.5 km above the top of the pluton, predict a maximum 0.5 km above the pluton at 4 x 10^4 and 10^5 years after emplacement (Figure 6). However, as a result of convective transport of thermal energy to the surface a porosity maximum is observed at depths of <2 km.

The spatial and temporal distribution of temperature in the system will directly determine the host rock resistivity distribution. The resistivity isopleths closely parallel the isotherms at 50,000 and 190,000 years (Figures 7 and 8, respectively), which also illustrate the displacement in the resistivity isopleths between 50,000 and 190,000 years. By 190,000 years the lateral extent of the isopleth displacement at a 1-km depth spans the entire width of the system (~22 km).

In summary, the calculations indicate that the dispersion of thermal energy away from a pluton will directly affect the host rock resistivity. When pluton emplacement is into permeable host rocks, significant decreases in resistivity between the surface and depths of <0.5 km are predicted. These resistivity values then persist uniformly in a vertical zone, extending from 0.5 km to approximately 4 km above the pluton by 190,000 years after pluton emplacement. The maximum decrease in resistivity is less than a factor of 10, as compared to surface values. The range in host rock resistivity is from 10^4 to 10^5 ohm m. These values are quite high with respect to values obtained on real rocks. However, our calculations only account for a conductive fluid in a nonconductive matrix.

Discussion

The temperature variations in hydrothermal systems account for changes in electrical porosity and electrical resistivity of pore fluids. Results of our analysis suggest that resistivity anomalies caused by thermal events are several times broader in extent than the thermal source, and the lateral resistivity gradients at the margins of the anomaly are much lower than the vertical resistivity gradients directly above the pluton. The side and top margins of the resistivity anomaly correspond closely to the 200° isotherm, as a consequence of the fluid properties. A resistivity minimum occurs at relatively shallow depths, e.g., 0.5 km, and extends to 4 km. However, the magnitude of these resistivities is considerably greater than values measured in geothermal systems.

The magnitude of ρ_R is defined by the pore fluid concentration and initial host rock porosity, while the distribution of ρ_R is defined by the temperature distribution. To determine the change in magnitude of ρ_R, due to varying molality of pore fluids and host rock porosities, a series of calculations was made with different initial values of porosity and NaCl molalities. The isopleths of resistivity as a function of porosity and NaCl molality at constant temperature are defined by (1) and shown for T = 300°C in Figure 9. The results of the calculations summarized as the minimum resistivities predicted for the cooling pluton environment are comparable to actual values realized in geothermal systems and in saline groundwater systems. In order to explain the observed resistivities in geothermal areas (<10 ohm m), high-molality pore fluids and/or high-porosity host rocks must occur for large vertical and horizontal zones within the geothermal system.

The results of this study indicate that rock resistivities characteristic of active hydrothermal systems are considerably less than can be accounted for by simple changes in fluid resistivity or rock porosity. The discrepancies be-

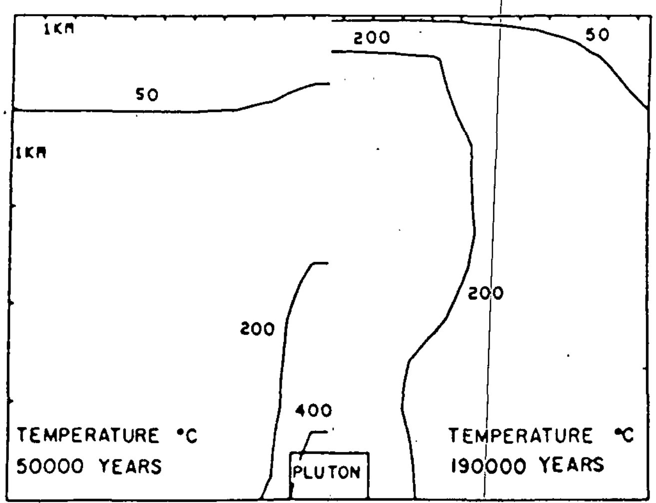

Fig. 5. Temperature distribution in an idealised hydrothermal system, defined by Figure 4, for (left) 5 x 10^4 and (right) 1.9 x 10^5 years elapsed time.

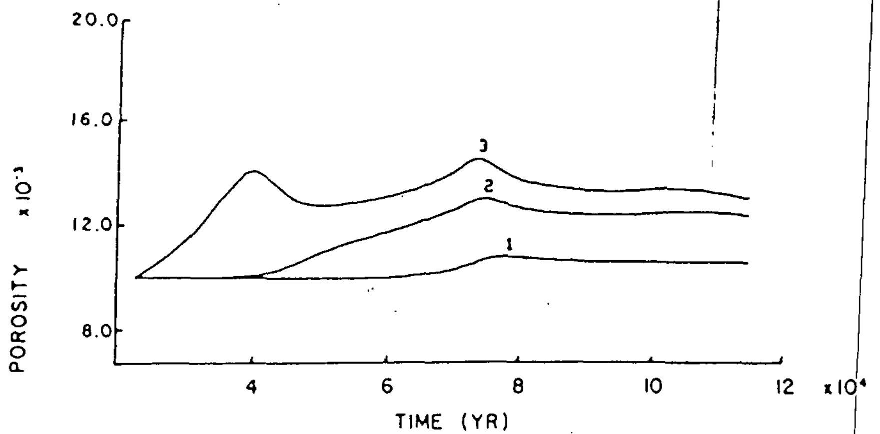

Fig. 6. Porosity as a function of time resulting from thermal energy transport into host rocks from the pluton at positions directly over the pluton, 0.5, 2, and 4 km below the surface (Figure 4).

tween the numerical resistivity models and the field resistivity observations in geothermal systems may be accounted for by the presence of conducting minerals, since pyrite and conductive clay minerals are typically found in the region of hydrothermal systems over the top of the thermal anomaly. If one uses a conservative estimate of a factor of 10 decrease in host rock resistivities resulting from conducting minerals, a geologically reasonable range in porosity and fluid composition can produce the anomalously low resistivity values observed in geothermal areas. Therefore except in anomalously high salinity and high porosity environments the presence of hot fluids alone is not sufficient to generate the low resistivity values observed in geothermal areas.

Considerable interest has been given to exploration techniques that might be useful in detecting high-energy geothermal systems. Commonly used techniques include measurements of heat flow and electrical resistivity. High heat flow in combination with anomalously low electrical resistivity data have been used as a justification for drilling of exploratory wells. Sedimentary basins and young, old, and mature geothermal systems in fractured rocks constitute a set of geologic environments within which the correlation of high thermal gradients, low near-surface resistivities, and surface thermal effects may lead to nonunique interpretations of the potential for geothermal energy resources at moderate depths. In the basin and range province of the western United States, concentrated brines associated with evaporite deposits in the high-porosity basins can produce lateral density gradients which cause fluid circulation. Exothermal hydration reactions that produce local thermal anomalies, coupled with the fluid circulation, are often sufficient to cause high surface heat flux and surface thermal springs. The high salinity and high porosity in these sedimentary basins would result in anomalously low near-surface resistivity. This particular environment appears

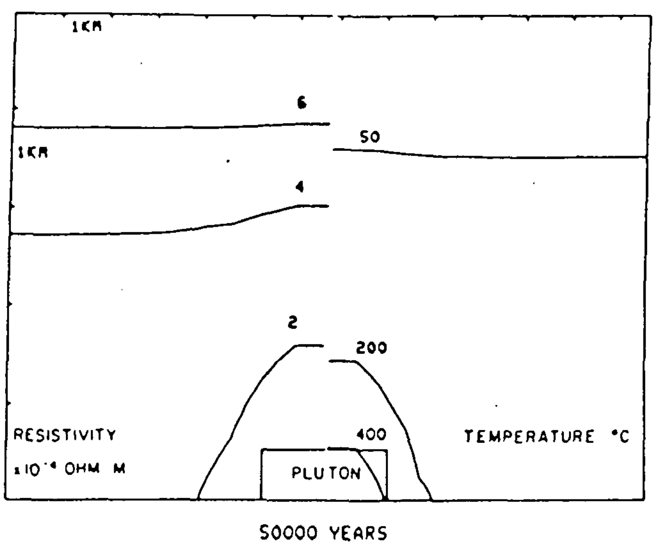

Fig. 7. Resistivity and temperature values in a hydrothermal system at 5×10^4 years elapsed time, depicting the temperature control on intrinsic resistivities.

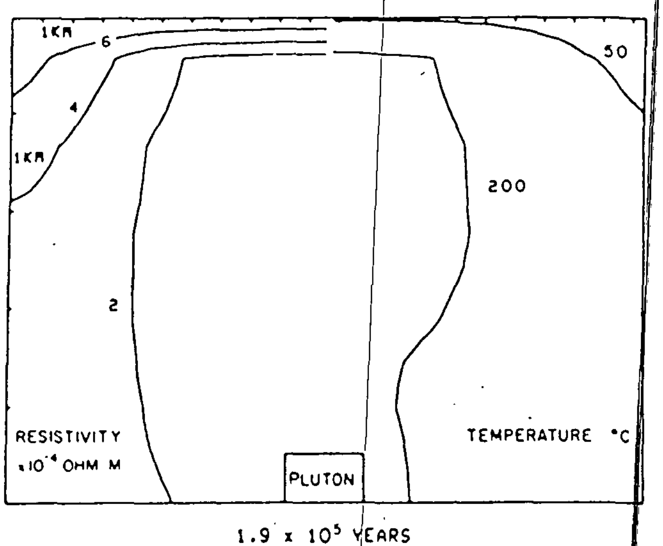

Fig. 8. Resistivity and temperature values in a hydrothermal system at 1.9×10^5 years elapsed time, depicting the temperature contro on intrinsic resistivities.

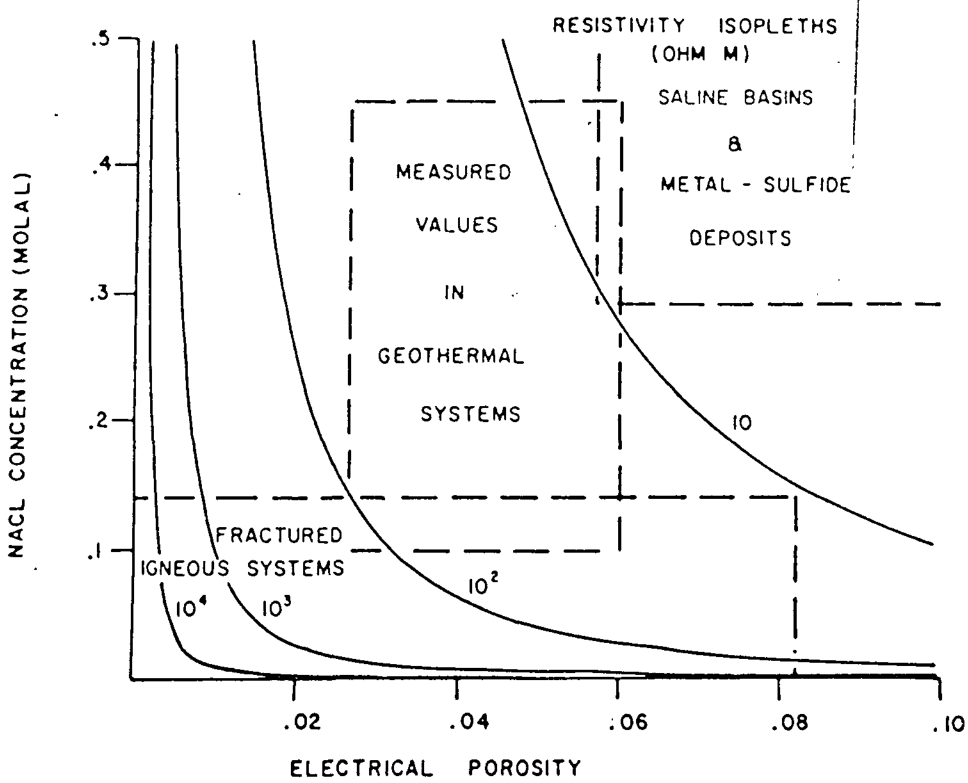

Fig. 9. Concentration of NaCl in pore fluids and electrical porosity effect on intrinsic rock resistivities, with isopleths of 10, 10^2, 10^3, and 10^4 ohm m. Values are for T = 300°C, at which the minimum in fluid resistivity occurs, and for P = 500 bars. Note that the minimum in fluid resistivity is nearly independent of pressures (Figure 3). Regions delimited by dashed lines represent ranges in values of porosity and fluid compositions observed in the respective geologic environments and for idealised systems considered in this study. The latter are labeled 'fractured igneous systems.'

to occur in the Safford Basin, southeastern Arizona [Norton et al., 1975].

Geothermal systems which have nearly cooled to regional background temperatures may be characterized by large conductive heat fluxes [Norton, 1977] as a result of remnant thermal energy that has been transported from the heat source to near-surface environments. Conducting minerals will undoubtedly have been formed above the pluton, and thermal springs will still be prevalent on the surface. In this environment, low resistivity would be associated with the conducting minerals and, in part, with the circulating saline fluids.

Geothermal systems in their early stages of formation have not been studied; however, their characteristics have been numerically simulated. The transport of thermal energy away from a pluton may be rapid with respect to the mass flux of reactive components in solution to the surface. This means that hot saline fluids will dominate changes in host rock resistivities because not enough time has elapsed to produce significant quantities of conducting minerals. High heat flux and surface thermal effects will probably form relatively early in the life cycle of a geothermal system. The calculated resistivity values resulting from increased temperatures are anomalous with respect to background values but are relatively high (10^4–10^5 ohm m). Therefore this environment is characterized by high heat flux and thermal surface effects but probably an undetectable resistivity anomaly, even though there is a high-energy thermal source at depth.

Active, mature geothermal systems are abundant worldwide where high heat flux, thermal surface effects, and low resistivities are associated with a productive thermal source at depth. However, low-resistivity anomalies, <100 ohm m, are probably caused by the presence of conductive minerals which may be coincident with hot thermal fluids.

The four geologic environments presented serve to illustrate the problems which can be encountered in attempting to interpret near-surface resistivity anomalies. Also a combination of heat flux measurements, surface thermal effects, and low resistivity can be characteristic of both productive high-energy geothermal systems and unproductive low-energy geothermal environments. The observations are also manifested in that electrical methods are used in prospecting for both sulfide mineral deposits and thermal energy.

Acknowledgments. We wish to acknowledge the Energy Research Development Administration for support of this study through contract E-11-1-2763. We thank Robert F. Butler, R. Knapp, and J. Knight for discussions and penetrating questions, L. McLean for editing and typing the manuscript, and the University Computer Center personnel for their assistance.

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(Received March 15, 1977;
revised August 15, 1977;
accepted August 26, 1977.)

DOE/ID/12079-90
ESL-108

RESISTIVITY, INDUCED POLARIZATION, AND SELF-POTENTIAL METHODS IN GEOTHERMAL EXPLORATION

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Work performed under contract DE-AC07-80ID12079

Prepared for

U.S. Department of Energy

Division of Geothermal Energy

3.0 ELECTRICAL PROPERTIES OF EARTH MATERIALS

3.1 Introduction

Bulk resistivities from the surface to in excess of 15 km depth in a normal crust are controlled by aqueous electrolytic conduction via pores, fractures, and faults. A slight increase in resistivity with depth in this region is the result of decreasing pore, fracture and fault porosity due to increased hydrostatic load. Fractures and faults are known to remain open to depths in excess of 5 km due to departures from hydrostatic loading. From about 15 km to the Moho, mineral semiconduction dominates and the resistivity decreases downwards. Semiconduction will remain the dominant conduction mechanism in excess of 100 km into the normal upper mantle.

In spreading centers (e.g. Iceland), intraplate melting zones (e.g. Hawaiian Islands), hot spots (e.g. Yellowstone, USA), subduction zones (e.g. Cascades volcanoes, USA and Canada), extensional continental regions (e.g. eastern Basin and Range, USA), and rift zones (e.g. East African Rift), the crust and mantle are abnormal in that they then contain melt or partial melt at any depth from surface to 100 km. Thus in geothermal areas, which abound in the tectonically active areas, one must be concerned with three basically different conduction mechanisms: aqueous electrolyte conduction, semiconduction, and melt conduction.

3.2 Aqueous Electrolyte Conduction

3.2.1 Normal mode of conduction

Conduction in near-surface rocks is largely electrolytic, taking place in pore spaces, along grain boundaries, in fractures and in faults but negligibly through the silicate framework.

The ions which conduct the current result from the dissociation of salts, such dissociation occurring when salts are dissolved in water. Since each ion is able to carry only a definite quantity of charge, it follows that the more ions that are available in a solution and the faster they travel, the greater will be the charge that can be carried. Hence, the solution with the larger number of ions will have the higher conductivity. Thus, in general, a rock which contains saline water within its pores will have a greater conductivity when the salinity of the water is high than when it is low; salinity is a major factor in determining the resistivity of a rock.

An increase in temperature lowers the viscosity of water, with the result that ions in the water become more mobile. The increased mobility of the ions results in an observed resistivity decrease with increase in temperature according to the formula

$$\rho_t = \frac{\rho_{18}}{1 + \alpha\ (t - 18)} \tag{34}$$

in which α is the temperature coefficient of resistivity (usually given as about 0.025 per degree centigrade), t is the ambient temperature, ρ_t is the resistivity at this temperature, while ρ_{18} is the resistivity at 18°C.

Archie's Law,

$$F = \frac{\rho_r}{\rho_w} = \phi^{-m} , \tag{35}$$

usually is satisfied for aqueous electrolytic conduction. In (35), F is formation factor, ρ_r is the resistivity of the rock, ρ_w is resistivity of the saturating electrolyte, ϕ is porosity, and m is the cementation factor which usually varies between 1.5 and 3.

3.2.2 The effect of clays on rock resistivity

A clay particle acts as a separate conducting path additional to the electrolyte path. The resistance of this added path is low. The origin of this abnormally high clay mineral conductivity lies in the double layer of exchange cations as shown in Figure 1. The cations are required to balance the charge due to substitution within the crystal lattice, and to broken bonds (Grim, 1953). The finite size of the cations prevents the formation of a single layer. Rather, a *double layer* is formed; it consists of a *fixed layer* immediately adjacent to the clay surface and a *diffuse layer* which drops off in density exponentially with distance from the fixed layer.

The diffuse layer, in contrast to the fixed layer, is free to move under the influence of an applied electric field. The cations of the diffuse layer add to the normal ion concentration and thus increase the density of charge carriers. The net result is an increased *surface conductivity*. Although clay minerals exhibit this property to a high degree because of their large ion exchange capacity, all minerals exhibit it to a minor extent. All rocks containing clay minerals possess an abnormally high conductivity on this account.

The effect of disseminated clay or shale on rock resistivities becomes increasingly important as the conductance through the pores diminishes. In a geothermal environment, hydrothermal alteration converts feldspars to kaolinite, montmorillonite and other clay minerals, especially in silicic rocks. In basic rocks, chlorite and serpentine may also be produced. All of these alteration products exhibit high surficial conductivity. As the concentration of the electrolyte increases, the relative contribution of the electrolyte conduction path to the clay conduction path increases as may be seen from the formula

$$\sigma_r = \frac{\sigma_e + \sigma_s}{F} \qquad (36)$$

in which σ_r, σ_e, and σ_s represent the observed conductivities of the rock, the electrolyte, and the clay surface path. Ward and Sill (1976) demonstrate that $\sigma_s \sim 3\ \sigma_e$ for altered rocks at Roosevelt Hot Springs, Utah, USA, despite the presence of an electrolyte containing 7000 ppm total dissolved solids.

3.3 Induced Polarization in Geothermal Areas

3.3.1 Introduction

Pyrite and clay minerals often are found as alteration products in geothermal areas. Hence the induced electrical polarization mechanisms of electrode polarization and membrane polarization might be expected in these areas.

3.3.2 Electrode polarization

Whenever there is a change in the mode of current conduction, e.g. from ionic to metallic, energy is required to cause the current to flow across the interface. This energy barrier can be considered to constitute an electrical impedance.

The surfaces of most solids possess a very small net attraction for either cations or anions, as we mentioned earlier for clay minerals. Immediately adjacent to the outermost solid layer is adsorbed a layer of essentially fixed ions, one or a few molecular layers in thickness (Figure 2a). These are not truly exchangeable and, hence, constitute the fixed layer.

Adjacent to the fixed layer of adsorbed ions there is a group of relatively mobile ions, either of the same or opposite charge, known as the diffuse layer. The *anomalous* number of ions in this zone decreases exponentially from the fixed layer outward to the normal ion concentration of the liquid. (The normal balanced distribution of anions and cations has been deleted from Figure 2 for clarity). The particular distribution of ions shown is only one of several possible distributions, but it is the most common. The electrical potential across the double layer has been plotted also; the potential drop across the diffuse layer is known as the Zeta potential (Z).

While the fixed layer is relatively stable, the diffuse layer thickness is a function of temperature, ion concentration in the *normal* electrolyte, valency of the ions, and the dielectric constant of the medium. Most of the anomalous charge is contained within a plane distance d from the surface, where (Grahame, 1947)

$$d = \left[\frac{\varepsilon_0 K_e k T}{2ne^2 v^2}\right]^{1/2} , \qquad (37)$$

n = normal ion concentration of the electrolyte,

v = valence of the normal ions,

e = elementary charge,

K_e = the dielectric constant of the medium,

k = Boltzman's constant,

and

T = temperature.

The thickness is, therefore, governed by the balance between the attraction of unlike charges at the solid surface and the thermal redistribution of ions. Obviously, increasing n, the salinity, or v, the valence, decreases the diffuse layer thickness.

Returning now to polarization at electrodes, it may be stated that there are two paths by which current may be carried across an interface between an electrolyte and a metal (Figure 3). These are called the faradaic and nonfaradaic paths. Current passage in the faradaic path is the result of an electrochemical reaction such as the oxidation or reduction of some ion, and may involve the diffusion of the ions toward or away from the interface. The charge is carried physically across the interface by an electron transfer. In the latter, i.e. nonfaradaic, case, charged particles do not cross the interface; rather, current is carried by the charging and discharging of the double layer. The nonfaradaic component, thus, may be represented by a simple capacitance insofar as the variation of its impedance with frequency is concerned.

In the faradaic path, the impedance associated with the electron transfer is represented by the reaction resistance. The ion diffusion process is not representable in so simple a fashion and, in fact, may not be adequately represented by any combination of fixed capacitors and resistors. It is customarily referred to as the Warburg impedance W and its magnitude varies inversely with the square root of the electrical frequency.

The interfacial impedances of many metal-electrolyte interfaces may be described roughly as follows. Above 1,000 Hz the major part of the electric current is carried across the interface by means of the non-faradaic path; hence, the interfacial impedance varies with frequency as approximately f^{-1}. As the frequency is lowered, more and more current is carried via the faradaic path, and so the low frequency impedance varies with frequency in the range $f^{-1/2}$ to f^0 depending on the magnitude of the impedance ratio W/R.

All of the above discussion applies to an ideal electrode in a pure electrolyte. The concepts, however, are important in understanding the processes occurring when current is passed through a rock. Any rock sample is *dirty* from the viewpoint of the physical chemist since the electrodes (metallic mineral grains) and electrolytes (pore solutions) are anything but pure. Nevertheless we perhaps are justified in employing equivalent circuits based on pure systems since a phenomenological explanation for rock behavior results. With this caution, one might suggest the equivalence of the elementary rock system of Figure 4a with the equivalent circuit of Figure 4b, where

W is the Warburg impedance

[= $k(1 - i)/ f^{1/2}$; k is a constant],

C_F is the double layer capacitance,

C_{CH} is the chemical capacitance,

R is the reaction resistance,

R' is the resistance representing a higher order reactions,

R_i is the resistance of ionic path,

and

R_m is the resistance of metallic vein path or particle.

In noting these circuit elements, it must be appreciated that one chemical reaction at the interface may lead to a chain of subsequent reactions involving electrons, ions, and molecules of all reaction products present. At each point of the reaction chain, the accumulation of the reaction product represents a capacitance C_{CH} to the electrode. The escape of the product is achieved either by diffusion, represented by a Warburg impedance W, or by a reaction represented by a resistor R. The product of this reaction in turn follows a similar circuit behavior which we have omitted for simplicity, except to lump all such products as R'.

While the circuits of Figure 4b and 4c satisfy the expected physical/chemical processes in mineralized rock, they are too complicated for practical use. Thus, the simple circuit of Figure 5a is used to predict induced polarization, of both electrode and membrane type, in a rock. The frequency and time domain responses of the circuit of Figure 5a are shown in Figures 5b and 5c, respectively. This is the Cole-Cole model of relaxation used by Pelton et al. (1978).

3.3.3 Membrane polarization.

In rocks containing a few percent clays distributed throughout the rock matrix, membrane polarization is of importance. Membrane polarization arises chiefly in porous rocks in which clay particles (membranes) partially block ionic solution paths [Figure 6a]. The diffuse *cloud* of cations (double layer) in the vicinity of a clay surface is characteristic of clay-electrolyte systems. On application of an electrical potential, positive charge carriers easily pass through the cationic cloud but negative charge carriers accumulate [Figure 6b]; an ion-selective membrane, therefore, exists.

Consequently, a surplus of both cations and anions occurs at one end of the membrane zone, while a deficiency occurs at the other end. This is because the number of positive charges cannot deviate significantly from the number of negative charges at any one point in space due to the large electric fields which would result if they did so deviate. These *ion concentration gradients oppose the flow of current.* The overall mobility of ions is reduced by this process. This reduction in mobility is most effective for potential variations which are slow (e.g., 0.1 Hz) with respect to the time of diffusion of ions between adjacent membrane zones. For potential variations which are fast (e.g., 1,000 Hz) with respect to the diffusion time, the mobility of ions is not substantially reduced. Hence, the conductivity of a membrane system increases as electrical frequency increases.

3.4 Semiconduction

The *intrinsic* conductivity of a solid at temperature T is computed from the relation

$$\sigma = |e| \, [n_e \mu_e + n_h \mu_h] \qquad (38)$$

where n_e, n_h are the electron and hole equilibrium concentrations, and μ_e, and μ_h are the mobilities of electrons and holes respectively while e is the elemental charge.

Kinetic theory leads us to expect a temperature dependence of the form $e^{-E/kT}$ for the concentration of electrons in the conduction band of a solid. Assuming a relatively small variation of mobility with temperature, we are then led (Kittel 1953) to predict a conductivity dependence of the form

$$\sigma = \sigma_0 e^{-E_g/2kT} \tag{39}$$

in which E_g is the gap energy, σ_0 includes the mobility function, and, in this form, is the conductivity as $T \to \infty$. Boltzmann's constant is k. Thermal, electrical, or optical excitation of electrons across the band of forbidden energy renders the solid conducting.

Impurities and imperfections in the material produce extrinsic conductivity. Above some temperature, impurities may be unimportant so that we define the temperature range above extrinsic conductivity as the intrinsic range in which the previous mechanism is operative.

However, below the intrinsic range, certain types of impurities and imperfections markedly alter the electrical properties of a semiconductor. Extrinsic semiconduction arises by thermal excitation of electrons (occupying intermediate energy levels in the forbidden gap produced by impurities in solid solution) into the unoccupied conduction band, or by the excitation of electrons from the occupied valence band into unoccupied impurity levels.

Ionic conduction in a solid occurs as a result of mobile ions moving through the crystal lattice as a result of defects in it. The simplest imperfection is a missing atom or lattice vacancy (Schottky defect). The diffusion of the vacancy through the lattice constitutes transport of charge. The conduction mechanism above 1,100°C is recognized as ionic because, when an iron electrode is used in contact with a magnesium orthosilicate, iron diffuses into the silicate replacing the magnesium.

Table 1 illustrates the temperature ranges important to extrinsic, intrinsic, and ionic conduction.

3.5 Melt Conduction

A silicic magma chamber can be expected to exhibit a resistivity two to three orders of magnitude lower than its solid rock host as the experiments of Lebedev and Khitarov (1964) have demonstrated. Duba and Heard (1980) measured resistivity on buffered olivene while Rai and Manghnani (1978) measured electrical conductivity of basalts to 1550°C; these latter measurements establish that mafic rocks can demonstrate low resistivities also. Resistivities of order 1 Ω m are to be expected in either silicic or basic melts due to ionic conduction.

For partial melts, the melt phase will serve as an interconnection of low resistivity in a residual crystal matrix of resistivity two or more orders greater and will determine the bulk resistivity (Shankland and Waff, 1977). An Archie's Law dependence is hence expected.

13.0 FIGURE CAPTIONS

Fig. 1. Schematic representation of ions adsorbed on clay particle (after Ward and Fraser, 1967).

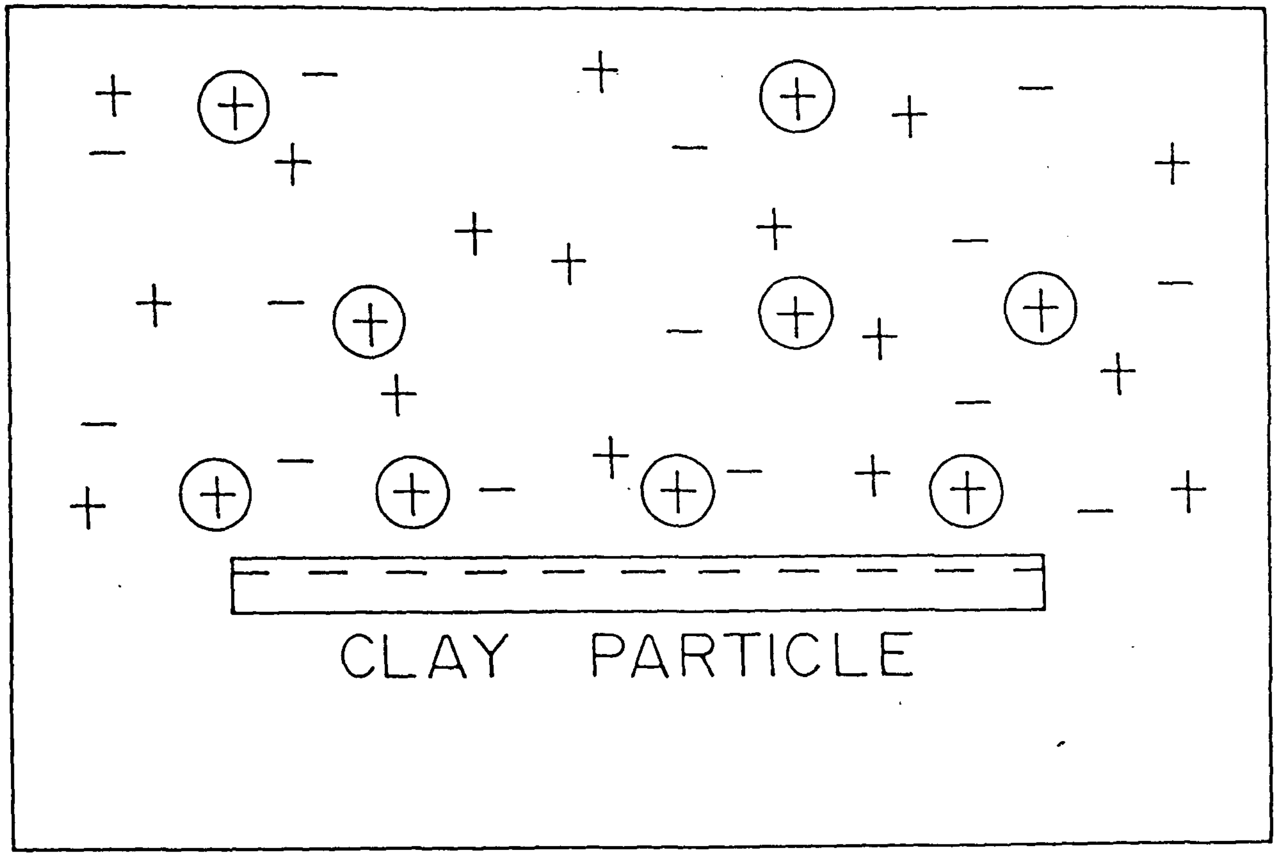
Fig. 2. (a) Hypothetical anomalous ion distribution near a solid-liquid interface; (b) Corresponding potential distribution (after Ward and Fraser, 1967).

Fig. 3. Circuit analog of interfacial impedance (after Ward and Fraser, 1967).

Fig. 4. Simplified representation of mineralized rock, (a) and the corresponding equivalent circuit (b) and (c) equivalent circuit of all mineralized rocks (after Ward and Fraser, 1967).

Fig. 5. Simplified analog circuit model of rock. (a) Elementary circuit, (b) frequency response of elementary circuit, (c) transient response of elementary circuit, and (d) a generalization of the elementary circuits.

Fig. 6. Depiction of ions in a pore space forming an ion concentration barrier which creates membrane polarization: (a) Pore path before application of an electric potential, (b) Pore path after application of a potential (after Ward and Fraser, 1967).



⊕ ADSORBED AND EXCHANGE CATIONS
\+ NORMAL CATIONS
− NORMAL ANIONS

Fig. 1. Schematic representation of ions adsorbed on clay particle (after Ward and Fraser, 1967).

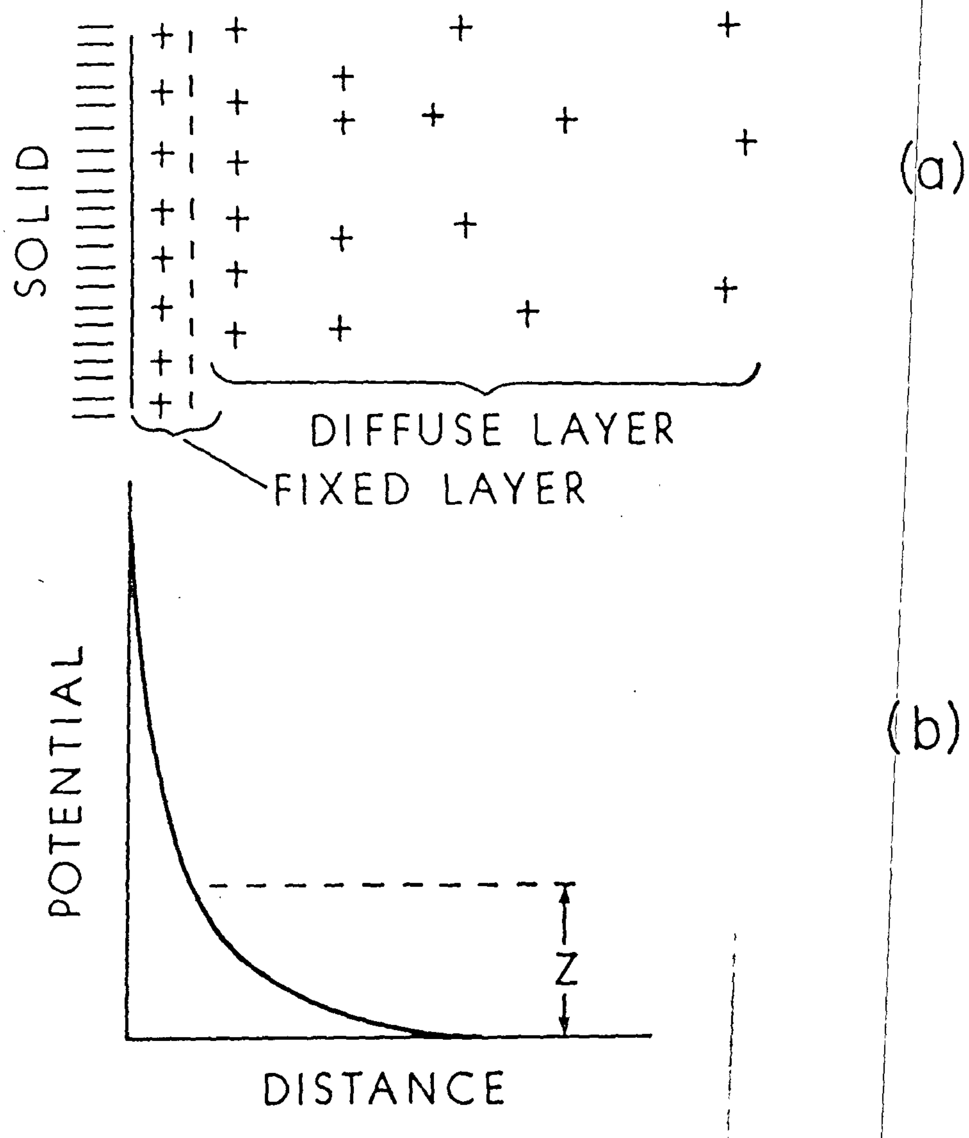


Fig. 2. (a) Hypothetical anomalous ion distribution near a solid-liquid interface; (b) Corresponding potential distribution (after Ward and Fraser, 1967).

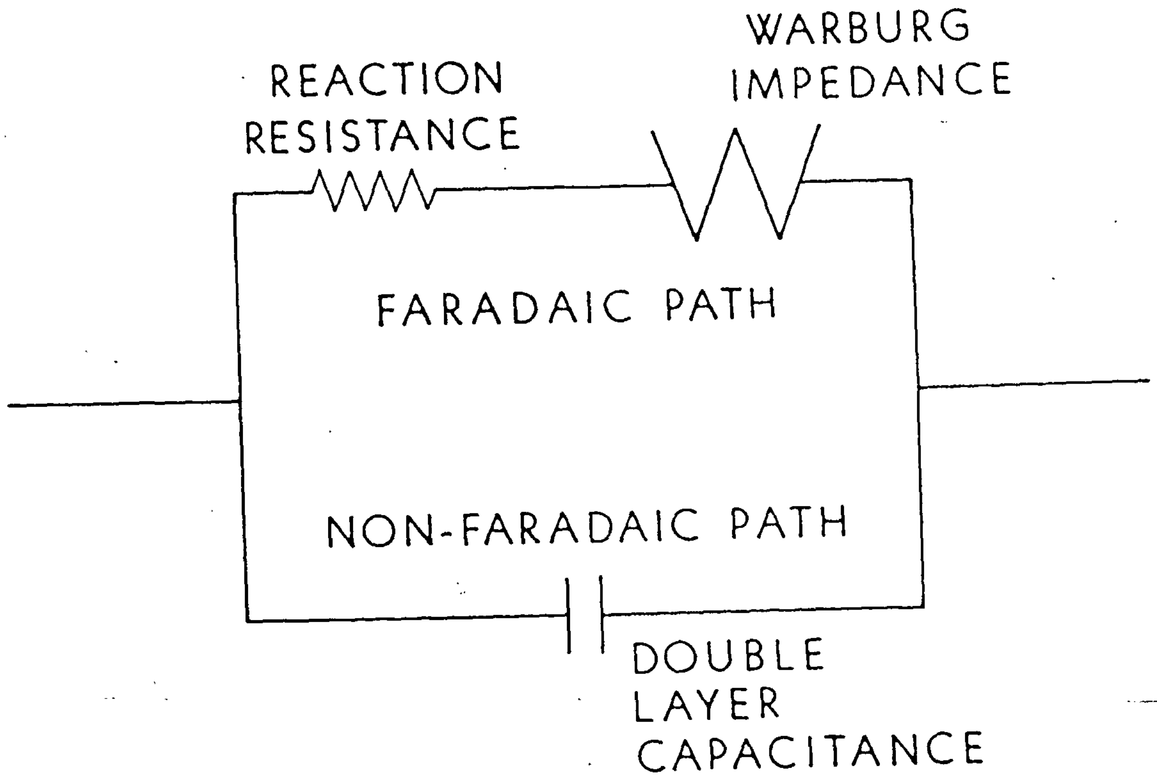


Fig. 3. Circuit analog of interfacial impedance (after Ward and Fraser, 1967).

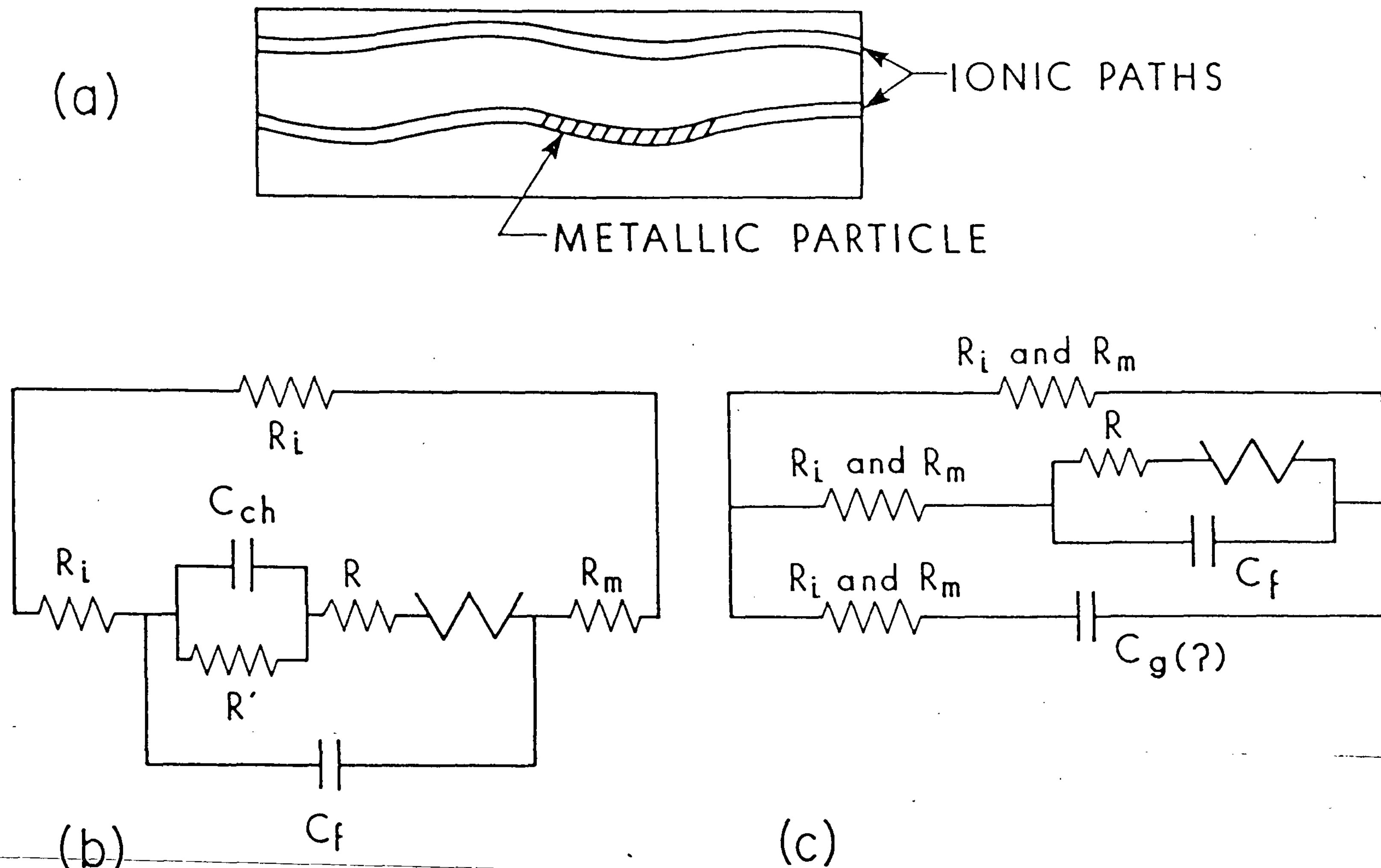


Fig. 4. Simplified representation of mineralized rock, (a) and the corresponding equivalent circuit (b) and (c) equivalent circuit of all mineralized rocks (after Ward and Fraser, 1967).

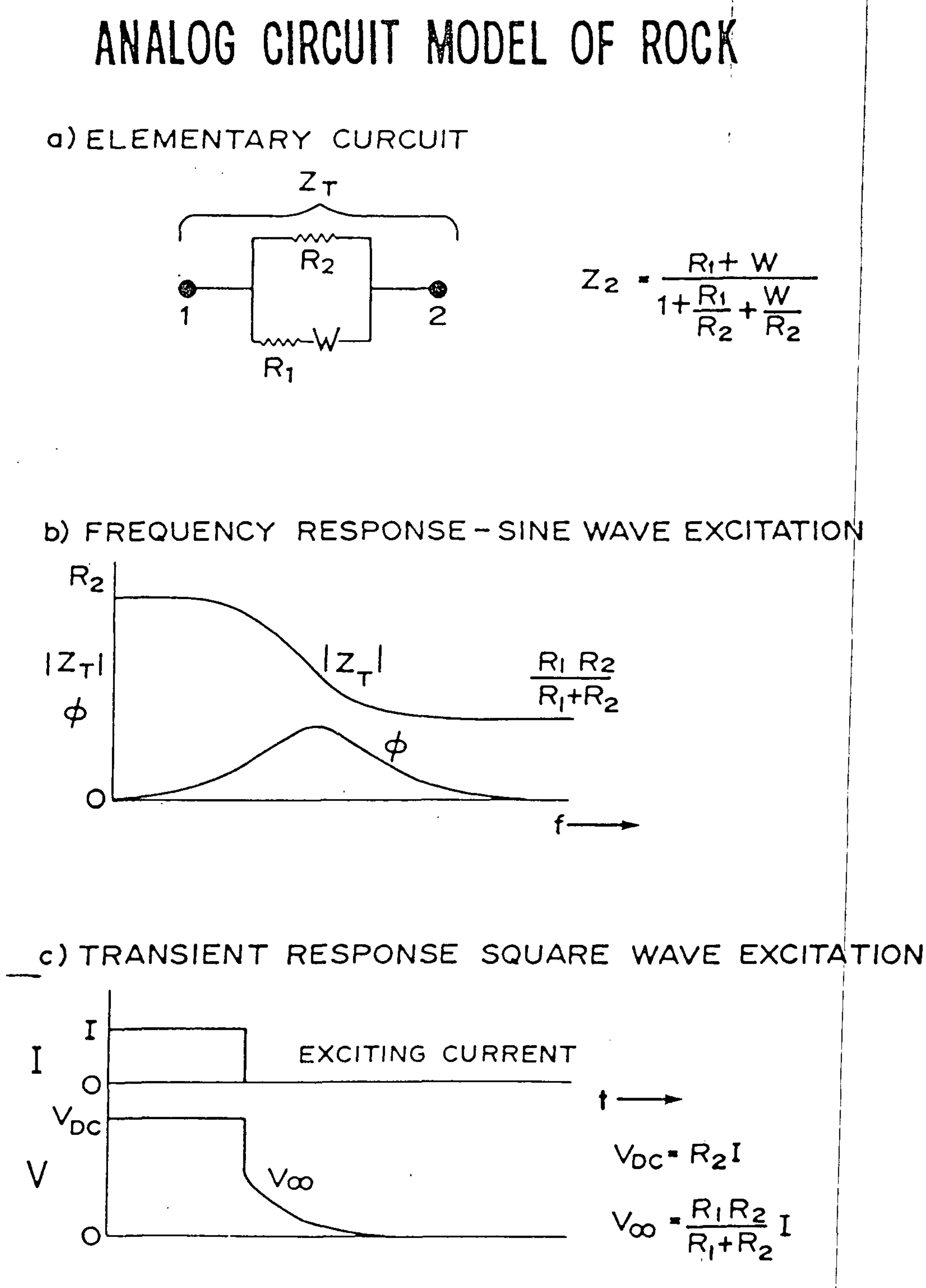


Fig. 5. Simplified analog circuit model of rock. (a) Elementary circuit, (b) frequency response of elementary circuit, (c) transient response of elementary circuit, and (d) a generalization of the elementary circuits.

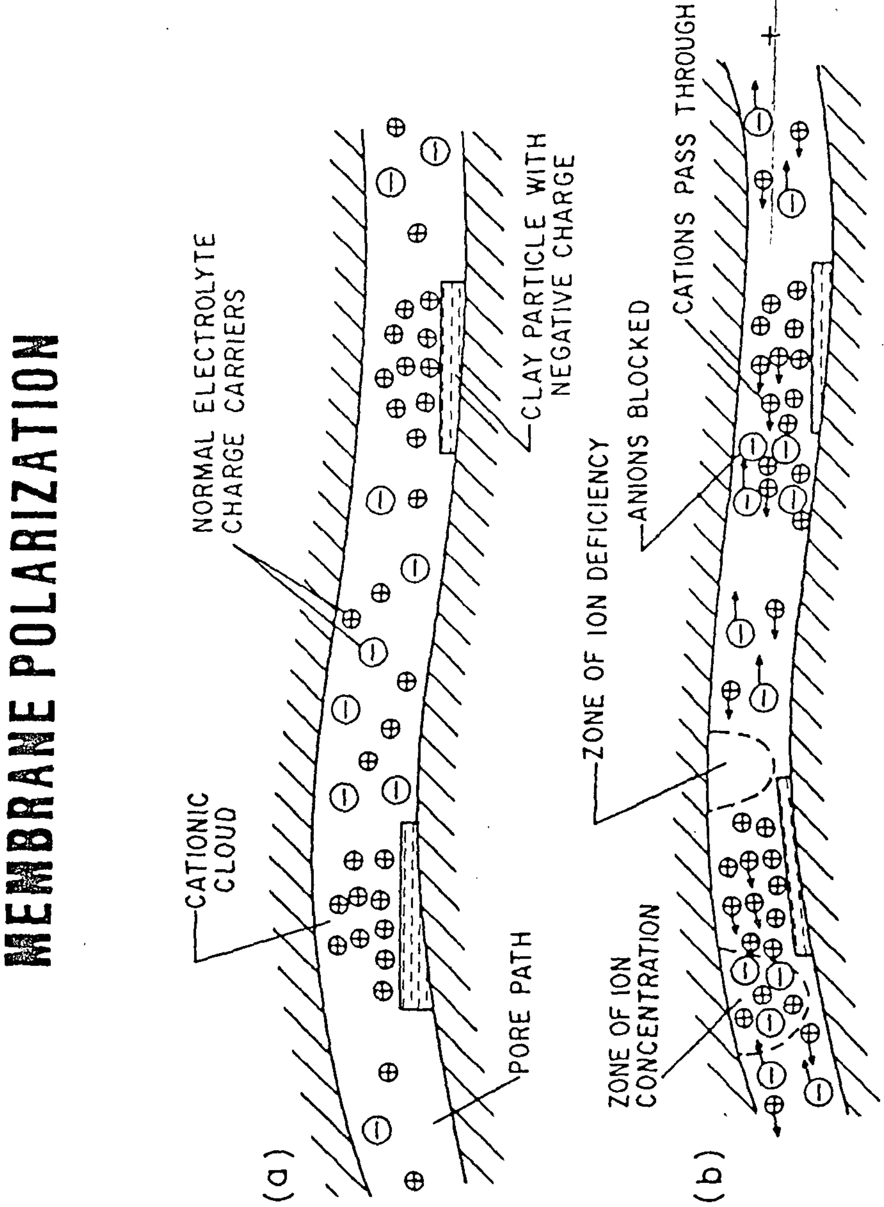


Fig. 6. Depiction of ions in a pore space forming an ion concentration barrier which creates membrane polarization: (a) Pore path before application of an electric potential, (b) Pore path after application of a potential (after Ward and Fraser, 1967).

TABLE 1

TYPE OF SEMICONDUCTION	σ_o	E	RANGE OF IMPORTANCE
EXTRINSIC	10^{-6} mho/m	1 ev	600°C
INTRINSIC	10^{-3} mho/m	3.3 ev	600 to 1,100°C
IONIC	10^{3} mho/m	3.0 ev	1,100°C

Table 1 Semiconduction follows the formula $\sigma=\sigma_0 e^{-E/kT}$ but σ_0 and E are different for each conduction mechanism. The values of σ_0 and E are stated here as are the temperature ranges of importance to each of the three mechanisms; extrinsic electronic, intrinsic electronic and ionic.

ELECTROMAGNETIC PROPERTIES OF ROCKS

by

William R. Sill

The electromagnetic properties of rocks appear in the constitutive relationships for the field vectors of Maxwell's equations. The constitutive equations are

$$B = \mu \cdot H \,, \tag{1}$$

$$D = \varepsilon \cdot E \,, \tag{2}$$

and

$$J = \sigma \cdot E \,, \tag{3}$$

where μ is the magnetic permeability, ε is the dielectric permittivity and σ is the electrical conductivity. In general, the relations in equations (1) to (3) can be nonlinear and the constitutive parameters are complex tensors which can be functions of frequency (time), temperature, pressure and composition. Under the more usual conditions they are treated as real or complex constants.

Measurements of these parameters are often carried out by forming the sample into a convenient geometry and measuring the capacitance, inductance, and resistance of the sample. Under the appropriate assumptions these measurements and the known geometry can be used to calculate the intrinsic values of the dielectric constant ($\varepsilon/\varepsilon_0$), the relative magnetic permeability (μ/μ_0) and the electrical conductivity or resistivity ($\rho = 1/\sigma$). The interpretations of laboratory measurements of these properties is not always as straightforward as the above discussion might indicate. Certain interfacial effects that can occur at the boundaries of the sample can invalidate the interpretation and steps must be taken to reduce them. Electrode polarization in dielectric and conductivity measurements is an important example of these effects. Also in the case of electrical measurements, dielectric effects can be superimposed on conduction effects and this can complicate the interpretation.

Another approach to the description of electromagnetic properties involves the use of mixture formulas for heterogeneous systems. These are usually developed in conjunction with laboratory measurements of the properties and they often present a useful insight into the physical processes involved. In a rock the heterogeneous system consists of the rock forming minerals and the material filling the pore space. The pore filling is usually a water solution or less often a gas or a liquid hydrocarbon. In a rock, the contributions of the various components differ widely depending on the property and the physical conditions as we shall see in the following discussions.

Magnetic Permeability

In rocks the principal magnetic minerals and their permeabilities are given in Table 1.

Table 1

Magnetic Minerals

Mineral	μ/μ_o
Magnetite	5
Pyrrhotite	2.6
Ilmenite	1.6
Hematite	1.05
Pyrite	1.0015

Except for massive orebodies, these minerals are usually minor constituents of rocks. A useful mixing law for rocks expresses the relative magnetic permeability (μ_r) as

$$\mu_r = 1 + (\mu_m - 1)v_m$$

where μ_m is the relative magnetic permeability of the magnetic mineral and v_m is its fractional volume. For one percent magnetite in a granite the relative magnetic permeability is around 1.04. In most electromagnetic prospecting problems it is therefore assumed that the magnetic permeability is the same as free space.

Dielectric Constant

The dielectric constant ($\varepsilon/\varepsilon_o$) of most rock forming silicates falls in the range from 1 to 10. The next most important constituent is water with a low frequency dielectric constant of about 80 (Figure 1). Small amounts of adsorbed water on the pore surfaces of nominally "dry" rocks can have profound and complicated effects on the electrical properties because the interfacial water is conductive and polarizable. Under terrestrial conditions where the more usual saturation is in the range from a few percent to completely saturated, the effects of the water become more tractable since the conduction and dielectric polarization in the bulk pore water dominate over the interfacial region. True dielectric measurements are usually made at relatively high frequency (> 10^4 hertz) first to minimize electrode polarization effects and second to ensure that displacement current is a dominant or at least a significant mechanism. When the effects of the interfacial water layers and the effects of conducting minerals in the rock can be ignored a useful formula for the dielectric constant (K_r) is given by the geometric mixing law,

$$K_r = \prod_i K_i^{v_i}, \tag{1}$$

where K_i is the dielectric constant of the i^{th} component and v_i is its volume fraction. In a typical rock with an average silicate dielectric constant of 5, containing 10 percent water, the rock dielectric constant is about 7.

Measured values on rocks fall in the range from around 4 to 25 (Telford et al., 1976).

Electrical Conductivity

The electrical conductivity of most silicates is very small ($< 10^{-9}$ s/m) at room temperature while the conductivity of pore water is much larger, usually in the range from 10^{-3} to 10^{1} s/m. Under the usual upper crustal conditions the pore water exists as an interconnected network and the conductivity of rocks is controlled solely by the conductivity of the pore water and the conductivity of the pore water-silicate interface. When the bulk pore water conduction dominates a useful mixing formula is given by Archie's law,

$$\sigma_r = \sigma_w \phi^m , \tag{2}$$

where σ_r is the rock conductivity, σ_w is the pore water conductivity and ϕ is the functional porosity of the interconnected pore spaces. The exponent m is sometimes referred to as the cementation index and it has a value in the range from 1 to 3 depending on the rock type. Usually larger values for m are associated with "tighter" rocks and low porosity. The effects of the geometry of the pore network can be combined into the formation factor (F) which is given by

$$F = \phi^{-m} . \tag{3}$$

The remaining factor determining the rock conductivity is then the conductivity of the pore water which can be expressed as,

$$\sigma_w = \sum_i n_i Z_i e U_i \tag{4}$$

where n_i is the concentration of the i^{th} ion, Z_i is its valence, e is the electronic charge and U_i is the mobility. The mobility of most ions in water solutions is similar ($U \sim 5 \times 10^{-8}\ m^2 S^{-1} V^{-1}$) so that the major factor affecting the conductivity is the concentration. The mobility of ions in solution is dependent on the temperature and this effect can be incorporated into the conductivity as

$$\sigma_r(T) = \sigma_r(T_o)[1 + \alpha (T - T_o)] \tag{5}$$

where T_o is the reference temperature and the temperature coefficient α is around $0.025°C^{-1}$. At higher temperatures the dependence is nonlinear similar to the change in viscosity. At temperatures near the triple point, the conductivity starts to decrease with temperature (Figure 2). This decrease is mostly due to an increase in ion association caused by the rapidly decreasing dielectric permittivity.

Interfacial effects on the conductivity arise because of the excess charge in the electrical double layer at the pore water-silicate interfaces (Figure 3). At most silicate interfaces the interaction with an electrolyte of moderate pH leads to a net negative charge on the interface and a compensating net positive charge in the diffuse portion of the double layer. With clay-type silicates the most important effect results from a net negative charge in the clay lattice and a compensating positive charge of exchange cations in the diffuse layer. In either case, the excess charge in the diffuse layer is relatively free to move under the influence of an electric field and this movement gives rise to the surface conductivity. The mobility of these ions is not as large as that of the same ions in a bulk solution and the temperature dependence is somewhat different (Waxman and Smits, 1968; Waxman and Thomas, 1974). The Waxman-Smits model provides a useful equation to describe the combined bulk and surface effects. This equation can be written as

$$\sigma_r = F^{-1}(\sigma_w + \sigma_s) \ , \qquad (6)$$

where σ_s is the contribution from the surface conduction. For clay minerals in sedimentary rocks Waxman and Smits (1968) give the surface conductivity as

$$\sigma_s = B\, Q_v \ , \qquad (7)$$

where B is the mobility of the clay exchange ions and Q_v is the cation exchange capacity per unit pore volume. When Q_v is express in equivalents per liter, which is a measure of the average concentration of the exchange ions in the pore volume, B is given by

$$B = 3.83\ (1 - .83\ \exp(-2\sigma_w)) \ , \qquad (8)$$

where σ_w is expressed in s/m. Equation (9) shows that the mobility in the Waxman-Smits model is an increasing function of the pore water conductivity.

A similar model for the surface effects due to non-clay silicates (Sill, 1982) gives σ_s as

$$\sigma_s = 2q^* U^* S_v \ , \qquad (9)$$

where q^* is the charge in the diffuse layer, U^* is the mobility of these ions and S_v is the surface area to volume ratio.

The major difference between the clay and normal silicate interface is the much larger surface charge on the clay. The charge density on montmorillonite is about 10^{-1} coulomb/m^2 while for a silicate with a surface potential of about 50 mV in a solution of concentration 10^{-2} molar the surface charge would be about 10^{-2} coulombs/m^2.

Semiconduction

The *intrinsic* conductivity of a solid at temperature T is computed from the relation

$$\sigma = |e| \, [n_e\mu_e + n_h\mu_h] \qquad (10)$$

where n_e, n_h are the electron and hole equilibrium concentrations, and μ_e and μ_h are the mobilities of electrons and holes respectively while e is the elemental charge.

Kinetic theory leads us to expect a temperature dependence of the form $e^{-E/kT}$ for the concentration of electrons in the conduction band of a solid. Assuming a relatively small variation of mobility with temperature, we are then led (Kittel 1953) to predict a conductivity dependence of the form

$$\sigma = \sigma_o e^{-E_g/2kT} \,. \qquad (11)$$

in which E_g is the gap energy, σ_o includes the mobility function, and, in this form, is the conductivity as $T \to \infty$. Boltzmann's constant is k. Thermal, electrical, or optical excitation of electrons across the band of forbidden energy renders the solid conducting.

Impurities and imperfections in the material produce extrinsic conductivity. Above some temperature, impurities may be unimportant so that we define the temperature range above extrinsic conductivity as the intrinsic range in which the previous mechanism is operative.

However, below the intrinsic range, certain types of impurities and imperfections markedly alter the electrical properties of a semi-conductor. Extrinsic semiconduction arises by thermal excitation of electrons (occupying intermediate energy levels in the forbidden gap produced by impurities in solid solution) into the unoccupied conduction band, or by the excitation of electrons from the occupied valence band into unoccupied impurity levels.

Ionic conduction in a solid occurs as a result of mobile ions moving through the crystal lattice as a result of defects in it. The simplest imperfection is a missing atom or lattice vacancy (Schottky defect). The diffusion of the vacancy through the lattice constitutes transport of charge. The conduction mechanism above 1,100°C is recognized as ionic because, when an iron electrode is used in contact with a magnesium orthosilicate, iron diffuses into the silicate replacing the magnesium.

Table 2 illustrates the temperature ranges important to extrinsic, intrinsic, and ionic conduction.

Table 2

Values for the constants in equation (11) and the temperature region where the mechanism is important.

Conduction Mechanism	σ_0(s/m)	E_g(eV)	Range of Importance
Extrinsic	10^{-6}	1.0	600°C
Intrinsic	10^{-3}	3.3	600°-1100°C
Ionic	10^{3}	3.0	1100°C

Melt Conduction

A silicic magma chamber can be expected to exhibit a resistivity two to three orders of magnitude lower than its solid rock host as the experiments of Lebedev and Khitarov (1964) have demonstrated. Duba and Heard (1980) measured resistivity on buffered olivene where Rai and Manghnani (1978) measured electrical conductivity of basalts to 1550°C; these latter measurements establish that mafic rocks can demonstrate low resistivities also.

Resistivities of order 1 Ω m are to be expected in either silicic or basic melts due to ionic conduction.

For partial melts, the melt phase will serve as an interconnected phase of low resistivity in a residual crystal matrix of resistivity two or more orders greater. This situation is similar to ordinary electrolytic conduction and an Archie's Law dependence is expected (Shankland and Waff, 1977).

Summary

The previous discussions have concentrated on the real part of the material property, treating them as scalar constants. This treatment is an oversimplification but the objective was to illustrate the major factors controlling the gross electromagnetic properties of rocks. In the case of induced polarization this treatment is inadequate and the details of this phenomena will be addressed later.

The treatment of the magnetic permeability is the simplest. Usually the permeability is taken to be that of free space unless the earth contains massive amounts of magnetic minerals.

The dielectric constant of saturated rocks is also relatively simple to treat when the effects due to the polarization of the adsorbed water can be neglected. The contribution of the dielectric displacement current ($\omega\varepsilon E$) relative to the conduction current (σE) can be evaluated by a consideration of the total current (J_T) given by

$$J_T = \sigma E + i\ \omega\varepsilon E \ . \qquad (12)$$

The ratio of the displacement current to the conduction current is then

$$\frac{\omega\varepsilon}{\sigma} = 10^{-10}\ \frac{\varepsilon f}{\varepsilon_0 \sigma} \ . \qquad (13)$$

Since the typical dielectric constant is about 10, the displacement current is relatively unimportant for frequency below a megahertz when the conductivity is larger than 10^{-2} s/m.

In the treatment of the conductivity the major factors were shown to be the water content, the geometry of the water network, the conductivity of the pore water and the surface conductivity. Surface conduction depends on the amount of clay present or the pore surface area to volume ratio for non-clay silicates. In both cases the surface conduction can become unimportant if the pore water conductivity is large enough.

The porosity of rocks can vary from a few tenths of a percent to around fifty percent giving rise to formation factors which span the range from 10^{-4} to almost 10^{0}. Since the conductivity of pore water ranges from 10^{-3} to 10^{1} s/m, it is not surprising that the observed range of rock conductivity in the upper crust is from 10^{-5} to 10^{1} s/m.

As rocks are subjected to greater lithostatic loading with depth, the porosity decreases. In the study of this decrease in porosity it is useful to separate the porosity into various categories depending on size and geometry. At the small end of the scale there are pores (rounded or tubular openings) and micro-cracks (thin and somewhat planar) and at the large end there are macro-cracks such as joints and fractures. The closing of these large scale joints and fractures with depth has been estimated by Brace (1971) to occur at depths of the order of hundreds of meters. However, we should probably discriminate between these types of fractures and joints and fault zones which are known to be open and permeable to much greater depths. For small scale features the laboratory studies of Brace and his co-workers (Brace et al, 1965; Brace and Orange, 1968; Brace, 1971) have shown the effects that crack and pore porosity have on the conductivity of rocks. In general the crack porosity can be an important contribution at low pressures where the cracks are largely open. At pressures above a few killobars most of the cracks close and the remaining porosity is due mostly to the more equidimensional pores. As the cracks close in the low pressure region the conductivity may decrease by an order of magnitude in a few killobars. At higher pressures, the rate of decrease due to pore closure levels off to about ten percent per killobar. Effective pressures of a few killobars correspond to depths in the range from 5 to 10 killometers so the rapid change in conductivity due to crack closure is expected to take place in this depth range. In the crust temperature also increases with depth and this will initially lead to an increase in the water conductivity. However, the decrease in porosity can usually be expected to be large enough that the net effect is a decrease in conductivity. At still greater depths and temperatures solid state semiconduction in the silicate framework will ultimately lead to an increase in the conductivity. These effects are illustrated in the models in Figure 4 from Brace (1971). The initial decrease in conductivity at essentially zero depth is due to the closure of large scale joints and fractures and the next region of rapid decrease which extends to about 8 killometers is due to the closure of micro-cracks. Below 8 killometers the more gradual decrease is due to the closure of pores and finally the rapidly increasing conductivity is due to the contribution from solid state semiconduction.

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EM PROPERTIES OF ROCKS

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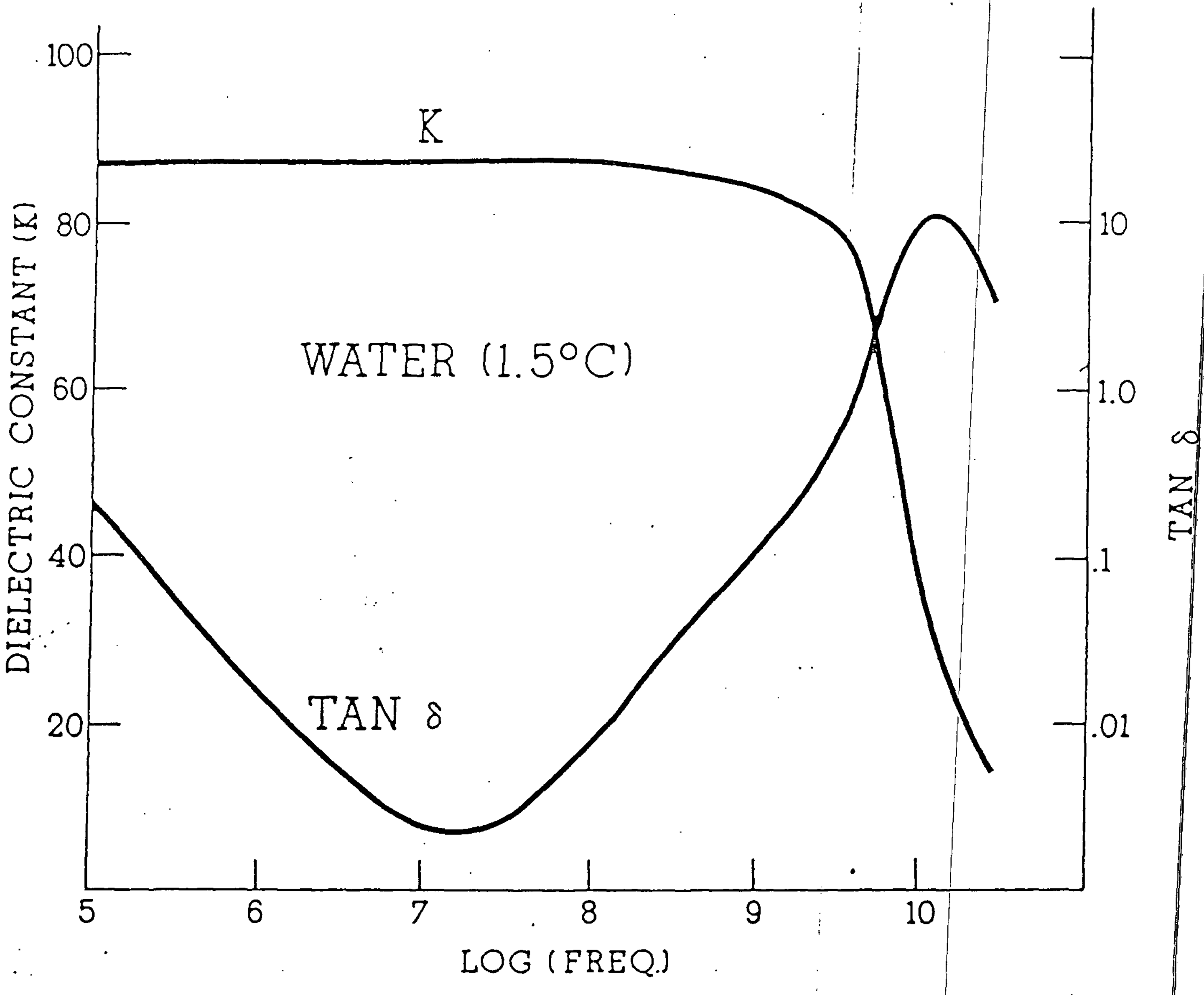

Figure 1. Dielectric constant (K) and loss tangent (Tan δ) of water.

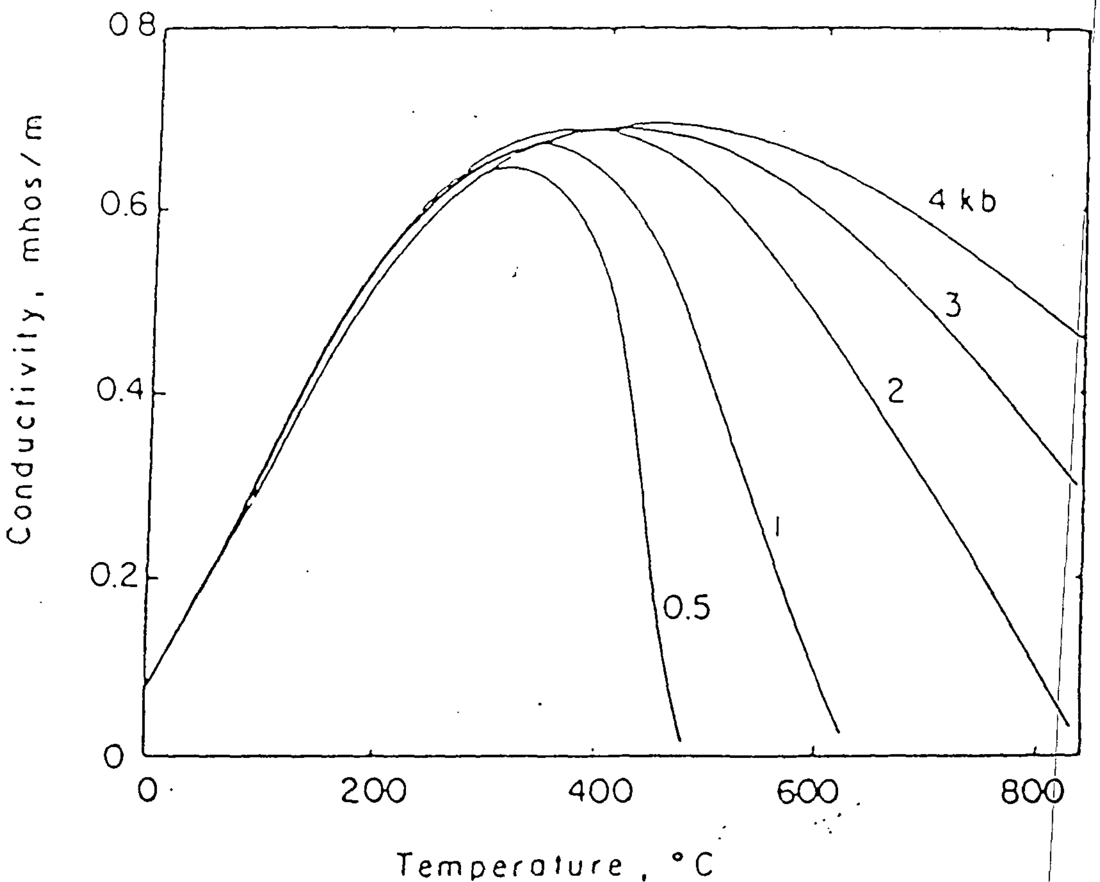

Figure 2. Conductivity of 0.01 molar NaCl as a function of temperature. The number on each curve is P_{H_2O}. (Quist and Marshall, 1968).

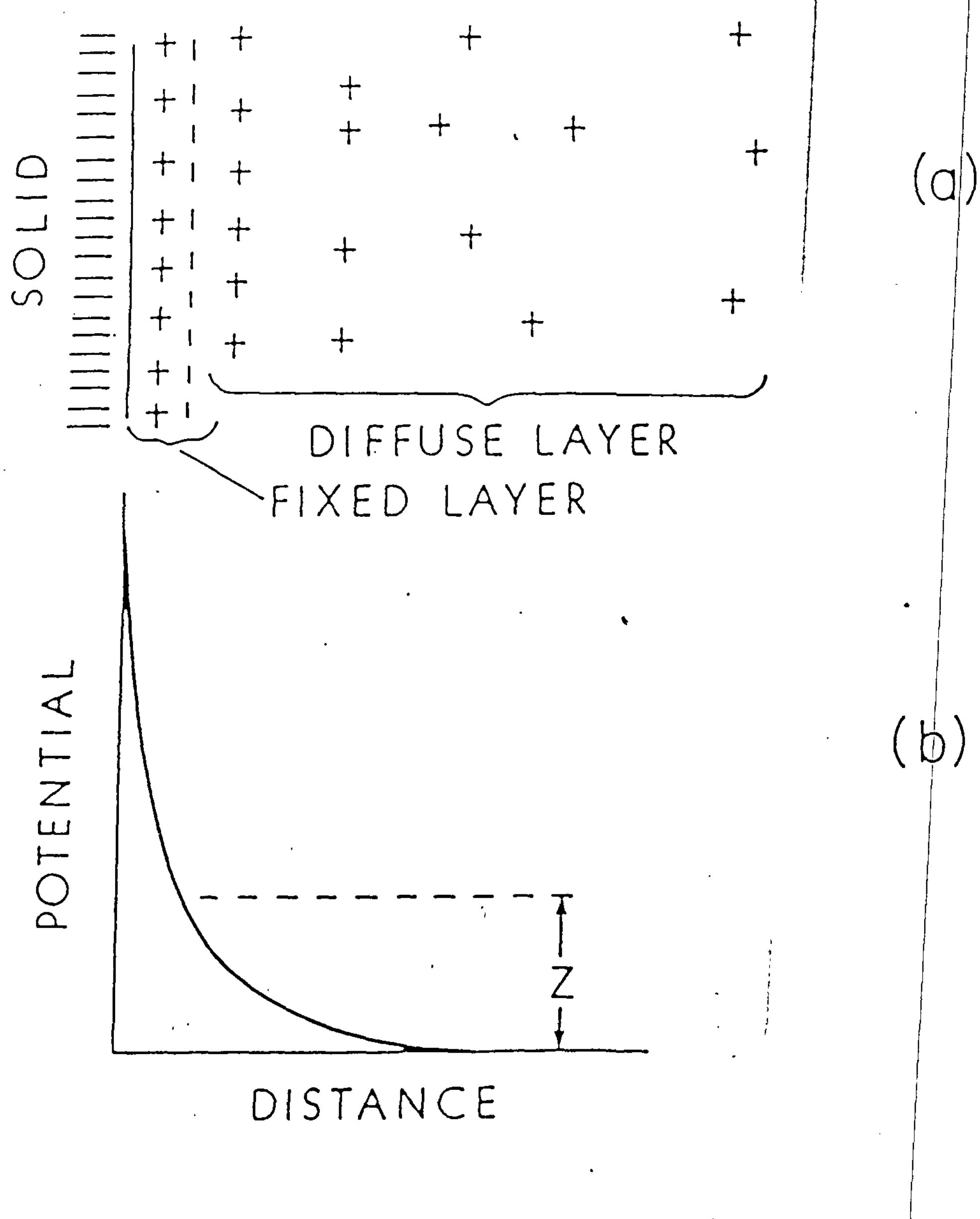

Figure 3. a) Ion distribution near a solid-liquid interface;
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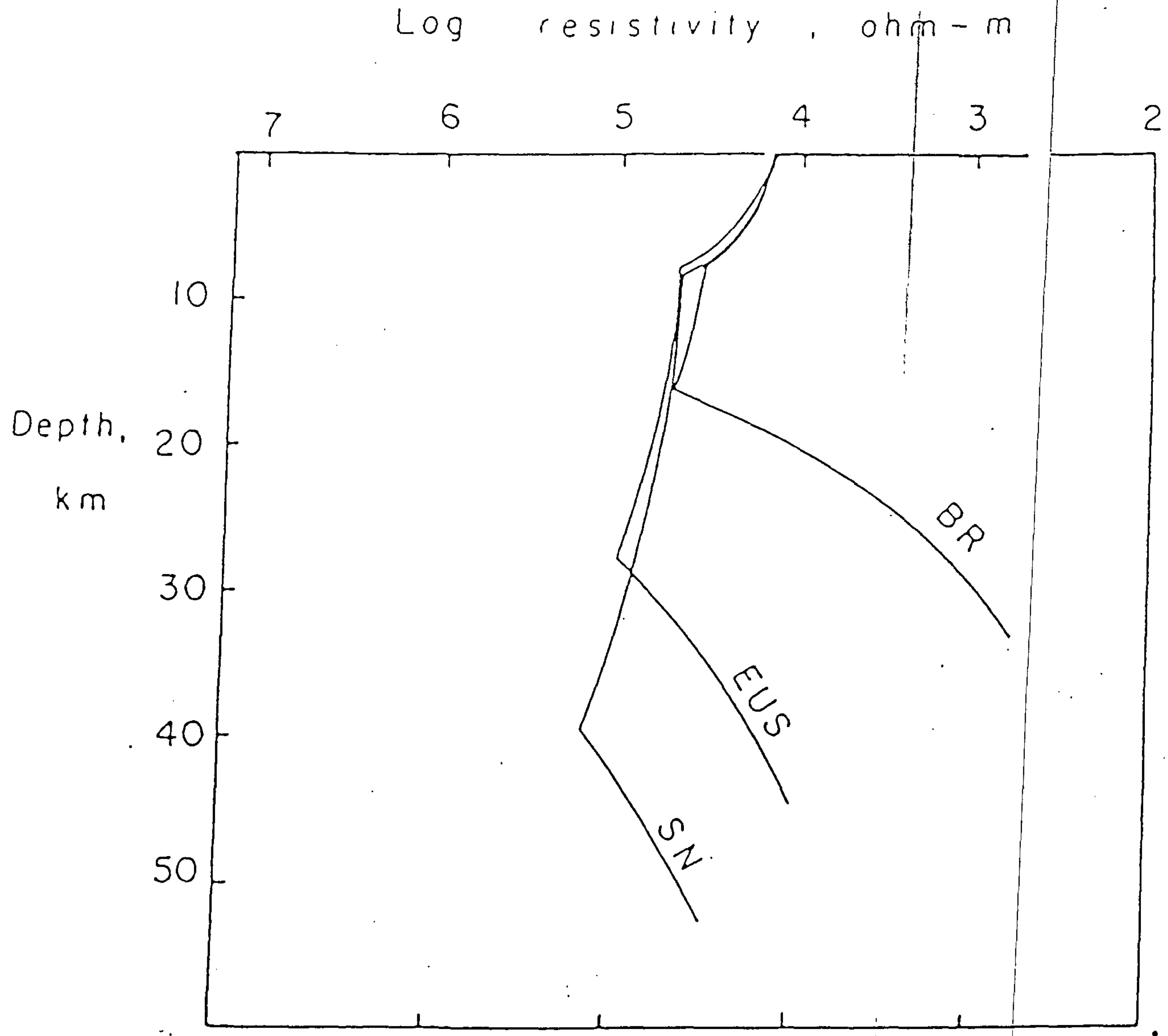

Figure 4. Comparison of resistivity-depth profiles for three heat flow provinces. (Brace, 1971).

GEOPHYSICAL METHODS IN GEOTHERMAL EXPLORATION

BIBLIOGRAPHIC SEARCH

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January, 1985

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INTRODUCTION

A bibliographic search of geophysical and geothermal journals, published reports and transactions from technical meetings was undertaken in order to establish and document the application of various geophysical techniques used worldwide for the exploration of geothermal resources. Over 700 pertinent references were assembled and are listed at the end of this report. A data base which indicates the worldwide application of various geophysical methods for geothermal exploration was created by reviewing the more significant publications within the bibliography. This data base was then used to evaluate the effectiveness of the geophysical methods within specific geologic and tectonic settings.

PRINCIPAL LITERATURE SOURCES

A computer-aided bibliographic search was conducted using the GEOREF data base of Dialog Information Services, Palo Alto, California. This search resulted in an extensive listing of technical articles which describe the application of geophysical methods for the exploration of geothermal resources around the world. A total of 554 listings was obtained which included references from technical journals, transactions and extended abstracts from technical meetings, government publications, doctoral and masters theses and geothermal texts. Approximately 200 additional references were obtained through a specific literature search, so that the total number of bibliographic references exceeds 700.

The GEOREF bibliographic entries are included in Appendix A of this report and are listed according to a GEOREF identification number which includes the year that the reference was placed in the data base. The most recently published references are generally listed first with the article title shown in boldface lettering. Unfortunately, there is no author cross-reference to aid in the search for a particular article. A bonus with this reference list is the list of key words that accompanies each reference entry. These key words provide valuable information regarding the articles. Abstracts are also included for some of the more recent listings.

The 200 additional references came from transactions of selected technical meetings, texts and technical journals. These references are listed by the publication and then by the author and article title in Appendix B of this report. Time did not permit a complete reorganization of the reference material. Because of the awkward method of listing GEOREF entries, there is some duplication between the GEOREF and the supplemental reference listings.

Approximately half of the references could not be reviewed in detail because of time and cost considerations, but this did not pose a limitation to our evaluation because of the duplication of information published within the literature (e.g., some authors have published essentially the same article in three separate publications such as the USGS Open File Reports, the GRC Transactions of Meetings and a technical journal such as Geophysics). The principal literature sources for the articles that were reviewed are given in Table I along with the number of articles reviewed. It is apparent from this table that technical meetings provide a greater wealth of information about exploration techniques than do technical journals which tend to emphasize theoretical advancements.

Those references that were not reviewed first-hand still provided information regarding the application of geophysical methods because of the list of key words that were included with each reference of the GEOREF bibliography.

Types of Articles

The references within the bibliographic list are divided into four basic categories and listed according to the predominance in the reference list: (1) reconnaissance exploration, (2) technique development, (3) theoretical and (4) generalized case history references. Most of the articles that were reviewed described reconnaissance geophysical surveys in areas where very little was known about the subsurface geology. Usually a limited suite of geophysical methods were utilized and little drilling information was available to confirm the interpretation of the geophysical data. In many of these articles, a development in a particular technique is illustrated by the application of the technique in a geothermal area. Again, this type of article rarely presents an integrated interpretation and testing of the interpretation by the drill bit. Theoretical articles are the next most prevalent type and they generally do not provide much information regarding the utilization of geophysical methods for exploration. Unfortunately, the most important type of article, the case history, is also the rarest in the literature.

Organization of Tabulated Results

The geophysical methods commonly used for geothermal exploration can be divided into ten basic categories: passive seismic, active seismic, passive electromagnetic, active electromagnetic, electrical resistivity, radiometric, thermal gradient, remote sensing, borehole geophysics and potential (gravimetric and magnetic) methods. Borehole methods utilize most of the other geophysical methods within the confines of existing boreholes for reservoir analysis and fracture detection. Within each basic category are specific geophysical methods, such as the gravimetric and magnetic methods within the potential methods category. In all, twenty-seven geophysical methods were identified and are shown in Table II with their designated abbreviations and acronymns.

The geologic and tectonic settings of the geothermal areas that were reviewed were divided into five basic types in Table III: rift valley, basin and range, intrusive volcanic, extrusive volcanic and basin settings. The use of geophysical methods within an area is based on the consideration of the geology, surface terrain, accessibility, the expected reservoir type and the survey cost versus resource profitability. These factors are variable for each geothermal area, but they are influenced by the geologic environment of the area; consequently the geologic setting exerts great influence on the choice of geophysical method applied in an area.

Both rift valley (RV), and basin and range (B&R) settings are defined by their particular style of tectonic activity and geomorphology. Examples of rift valley settings are the Rio Grande Rift in the USA, the East African Rift and the Baikal Rift in the USSR; the Basin and Range Province, USA is the classical example of a basin and range setting. A geothermal resource not occurring within an RV or B&R setting will generally be in close proximity to Tertiary to Recent age volcanism or will occur within deep basins. An intrusive volcanic (VI) setting is one in which a near-surface intrusive body acts as the heat source for a reservoir. The recently emplaced dike in the Puhimau thermal area of the Kilauea volcano, Hawaii is an example of this type of setting. A special case of the VI setting is the hot dry rock (HDR) resource, such as the older silicic intrusion in the Jemez Mountains of New Mexico. The intrusive body not only supplies the heat but also becomes the reservoir, either as a result of natural or man-made fracturing within the body. When the intrusive magmatic body is much deeper within the crust and a more conventional geothermal reservoir exist, then the area is classified as an extrusive volcanic (VE) setting. Calderas, such as the Long Valley, and Yellowstone calderas and extensive volcanic fields related to subduction zones (El Tatio, Chile) are embraced by this category. Basin settings (B) are deep basins with generally low- to moderate-temperature geothermal resources that are a result of the deep circulation of meteoric waters within the basin. The Paris Basin of France is an example of this type of setting. The geopressured resources of Texas and Louisiana, USA, are also in this general category.

Information regarding the application of the geophysical methods in various geologic settings and temperature regimes in countries and regions around the world is listed in Table III. A total of 47 countries or regions (e.g., the Caribbean Sea region) and 88 geothermal resource areas (e.g., Cerro Prieto, Mexico) are represented within Table III. The country/region listings are organized according to the geologic setting and then are arranged alphabetically within each geologic group. Only significant geothermal areas with a substantial number of accessible references were listed separately within the table (e.g., the Coso Hot Springs resource area is listed separately from the other Basin and Range resource areas of the USA, while all of the resource areas in India were listed together).

Geothermal resources within each region are classified as low-temperature ($T < 100°C$), moderate-temperatue ($100°C < T < 200°C$) or high-temperature ($T > 200°C$) resources. The three rows comprising each entry for a region correspond to the low-, moderate- and high-temperature classifications (L, M, H), respectively. This classification is similar to the generally accepted temperature classification given by White and Williams (1975), but does differ from it in the choice of the boundary temperature between the moderate- and high-temperature regimes (200°C versus 150°C). The present classification scheme was used in order to provide a more even distribution of resource areas among the three temperature categories.

The symbol 'X' is used in Table III for known information (i.e., geologic setting, resource temperature, and geophysical method), the symbol '+' is used for information derived or inferred from details within an article and the symbol '?' is used for uncertain interpretations made by the authors of an article. Multiple usage of a particular geophysical technique within a region is not noted.

The site-specific information in Table III was gathered together and assembled in Table IV according to geologic setting and resource temperature so that the utilization of geophysical methods could be more easily analyzed according to those two important criteria. The number of resource areas that were surveyed by a geophysical method and the total number of available areas for each geologic/temperature category are shown along with the subtotals for each geologic setting within each of the five categories. Totals for the number of resource areas and the utilization of geophysical methods for the three resource temperatures are shown at the bottom of the table. An alternate presentation format for the utilization of geophysical methods is a percent utilization table (Table V) which provides the percentage of resource areas in which a geophysical method was applied (i.e., the number of areas where a particular method was used divided by the total number of resource areas of that type and multipled by 100). With this data format, the utilization of geophysical methods in different geologic settings and temperature regimes can be compared directly since the number of occurrences has been normalized by the total number of possible occurrences. This table can be further simplified by replacing the numeric data with symbolic data that represents four categories of percent utilization: utilization $\geq$ 50%, 25% $\leq$ utilization < 50%, utilization < 25% and 0% utilization. This summary (Table VI) of the percent utilization of geophysical methods graphically delineates those geophysical methods popularly applied in various geologic settings and resource temperatures.

DISCUSSION AND CRITIQUE

Before noting and discussing the findings of this study, a few comments regarding the data base are required. A large variation in the number of articles per country/resource area is found within the GEOREF reference list. Table VII lists the number of references per country/region and these figures indicate that a majority of the references cover only a few of the countries. This reference list consists of the GEOREF bibliography plus the 23 articles contained within the geoelectric and geothermal studies of the USSR and the eastern bloc countries (Adam, 1976). This point is better illustrated by Table VIII which lists the seven countries with the most references. Not surprisingly, the geothermal exploration in the USA was referenced 340 times out of a possible of 575 references, thus comprising the 59% of the reference list. The next most referenced country is Italy with 30 articles or 5% of the reference list. The other five countries, the USSR, Japan, Mexico, Iceland and New Zealand, all are referenced fewer than 30 times (less than 5% of the list). References for these seven countries comprise 80% of the reference list; consequently there is a definite bias in the data set towards geothermal exploration in the USA. To help de-emphasize this bias, all but three of the Basin and Range geothermal areas were lumped into one category (the Basin and Range region) in order to cut down on the number of US resource areas. Even so, there are 20 separate resource areas in the USA that are listed in Table III.

Another observation regarding this study is that not all results of geothermal exploration are published in the literature. This is especially true in the USA where much of the geothermal exploration was funded by private companies that have kept the survey findings proprietary. The usage of such techniques as bipole-dipole, microearthquake and controlled source AMT is more widespread in the USA and would be better represented if this proprietary information were available. However, we do not feel that this problem will affect the significant findings of this study. Most of the statistical findings of this study are obtained directly from Tables IV to VI and these results will now be introduced.

The column totals of Table IV Indicate that 88 resource areas were reviewed and a total of 562 entries regarding the usage of geophysical methods were made to Table III. These tabulations do not include the 7 resource areas and the corresponding 11 geophysical entries that could not be correlated with one of the five geologic settings because of deficiencies in the published literature. These are referred to as the 'unclassified category' in Table III. The ratio of the total number of entries to resource areas indicates that an average of 6 different geophysical methods were applied in each of the reviewed geothermal areas around the world. Using the totals for each resource temperature, the average number of geophysical methods applied per resource area becomes approximately 5, 5 and 8 for the low-, moderate- and high-temperature resources, respectively. The range in the number of methods applied for the three temperature regimes is 3 to 9, 3 to 11 and 5 to 16 respectively. The maximum number of methods per area (16) occurred in the high-temperature basin and range setting and the minimum (3) occurred in the low-temperature extrusive volcanic and the moderate-temperature rift valley geologic settings. It is reasonable to conclude that more exploration effort is expended in the more profitable high-temperature resource areas, as indicated by the number of methods applied in the three temperature regimes. The third column of Table IV also shows that the number of resource areas reported for a particular geologic setting is proportional to the resource temperature in all but the basin geologic setting. Of the five geologic settings, the extrusive volcanic setting contains the most documented resource areas (47) as compared to fewer than 14 areas for each of the other four settings.

Tables V and VI are used to determine the most popular geophysical methods for the different geologic settings and temperature regimes. Table VI is the easiest of the two tables to use since the percent utilization of a geophysical method is divided into 4 categories representing significant ($\geq$ 50%), moderate (25% $\leq$ utilization < 50%, low (< 25%) and non (0%) utilization of the method. Considering all resource areas and temperatures, only three methods saw significant utilization: VES (59%), gravimetric (52%) and temperature gradient (50%) methods. The popularity of the VES method is due to its use as a low cost reconnaissance method. The Schlumberger sounding is the most popular of the VES methods, but Wenner and dipole-dipole (e.g., equatorial dipole-dipole) soundings have also been used. Popularity of the gravity method is also due to its low cost and because of its usefulness in defining geologic structure. The widespread usage of the TG method is obvious, since it is the only geophysical technique that actually measures the property that is being sought. Seven other methods were moderately used around the world: heat flow (48%), magnetic (39%), MT (35%), dipole-dipole resistivity (33%), reflection seismology (33%), MEQ (32%), remote sensing (28%), and bipole-dipole (26%). The least used methods included CSAMT, IP, pole-dipole and geomagnetic soundings, all of which have values of percent usage less than 7%.

The subtotal row for each of the five geologic settings in Tables VI (and V) indicates the distribution of different geophysical methods in the different settings. These results can be summarized as follows:

rift valley: significant - VES method

moderate - MEQ, gravimetric, magnetic, MT, dipole-

dipole, bipole-dipole and heat flow and TG methods

B and R: significant and moderate - all of the methods with the exception of geomagnetic soundings, CSAMT, HEP, SP and BG (borehole geophysical) methods

IV: significant - gravimetric, magnetic, VES, and temperature gradient methods
moderate - reflection seismology, AMT, MT, dipole-dipole and heat flow methods

EV: significant - gravimetric and VES methods
moderate - MEQ, reflection seismology, magnetic MT, dipole-dipole, bipole-dipole, SP, heat flow, TG, and remote sensing methods

basins: significant - gravimetric, VES, heat flow and TG methods
moderate - reflection seismology, MT and telluric methods

Examining the columns of Table VI, the VES and TG methods are clearly the most popular geophysical methods employed in all of the geologic settings and temperature regimes. The rows of Table VI indicate that the greatest utilization of geophysical methods occurs in the moderate- to high-temperature basin and range geothermal resource areas; this result is largely due to the extensive geothermal exploration of the Basin and Range Province in the USA.

One obvious criticism of Tables V and VI is that the popularity of a method does not necessarily indicate its value as an exploration tool. Too often a technique that has been successfully employed in one environment is then tried in other geologic settings and reservoir types with much poorer results. The bipole-dipole technique is a good example of the blanket usage of a technique in areas where it is not well suited. The original success of the technique in outlining the boundaries of the Broadlands Field, New Zealand (Risk et al., 1970) led to its use in such areas as the Olkaria Field, Kenya where it found little success in mapping the geothermal resource.

The evaluation of the usefulness of the various geophysical techniques is the most difficult task of this study because of the few published comprehensive case studies of geothermal exploration programs. Ward (1983) provides an excellent evaluation of the geophysical methods in the exploration of geothermal resources in the Basin and Range Province of the western US. Ward evaluated 14 methods in 13 high temperature sites (including Long Valley, Coso Hot Springs, Roosevelt Hot Springs and Raft River) and concluded that: a) none of the various geophysical methods were uniformly consistent in performance; b) none of the methods was ranked in the "good" category and only five methods were ranked in the good to fair category (MEQ, gravimetric, electrical resistivity, SP and heat flow/TG); c) the least effective methods are seismic noise, magnetic and MT; and d) no combination of any four methods was ranked as "good to fair" in success at more than one site. It is noteworthy that two of the least effective geophysical methods (magnetic and MT) were significantly utilized in the basin and range geologic setting worldwide according to Tables V and VI. Additional observations made by Ward are: a) quiet periods between MEQ swarms limits the use of the MEQ method in some areas; b) reflection and refraction seismology are not always applicable to reservoir delineation; c) the magnetic method is most useful for mapping zones of magnetite destruction; d) Schlumberger soundings and dipole-dipole profiling surveys are the best electrical resistivity methods; e) CSAMT and CSFEM methods have not been sufficiently tested yet; f) scalar AMT and tellurics should be limited to reconnaissance surveys; g) the SP method shows great promise but does not always produce a recognizable signature over geothermal systems; and h) shallow heat flow/TG is not always a reliable indicator of a high quality geothermal resource.

A variety of techniques, including CSMAT, VES, SP, gravimetric, magnetic, CSFEM, dipole-dipole, bipole-dipole, heat flow and remote sensing methods, were employed in the Puhimau thermal area of the Kilauea Volcano, Hawaii. The SP and VLF tilt angle and resistivity results delineated an area associated with high surface temperatures and a Schlumberger sounding was used to determine a minimum depth to the top of the subsurface conductive dike (Anderson, 1984). Comparable results were obtained using the CSAMT method (Bartel, 1984).

Most of the geothermal areas around the world are characterized by subsurface resistivities that are less than 10 ohm-m; regardless of the host rock resistviity. Consequently, in many areas it is sufficient to map the surface manifestations (hydrothermal alterations) of a deeper reservoir using an electrical resistivity technique. This has been successfully done in the Broadlands Field, New Zealand (bipole-dipole), Dieng Plateau, Indonesia (bipole-dipole), Olkaria Field, Kenya (dipole-dipole) and Roosevelt Hot Springs, USA (dipole-dipole or CSAMT). The use of the airborne TEM techniques have taken advantage of the near surface conductive zones above geothermal reservoirs in order to delineate some anomalous areas in the western US for more detailed studies.

In the Olkaria area, shallow TG and dipole-dipole profiling provided the most useful information and the bipole-dipole method the least useful information (Noble and Ojiambo, 1975). On the Island of San Miguel, Portugal, a reconnaissance geophysical effort that utilized the bipole-dipole method to map anomalous areas in the rugged terrain followed by Schlumberger soundings and dipole-dipole profiling was used to locate a successful geothermal well. Later an AMT survey in the same area confirmed the results of the electrical resistivity survey, perhaps indicating a more cost effective method of exploring the rugged terrain of the island (Hoover et al., 1984). Both the electrical resistivity and AMT surveys showed little correlation with the results of an MT 5-EX survey in the same area.

The existence of low resistivities is not a guarantee of anomalous subsurface temperatures. Sanford et al. (1979) reported on a case study of the Elephant Butte prospect, south central New Mexico which involved the use of the bipole-dipole method, modified Schlumberger soundings and subsequent heat flow determinations. The electrical methods successfully mapped the basement structure and faulting and delineated several areas of anomalously low resistivity. However, the heat flow data did not indicate any anomalous subsurface temperatures in the area.

An area where the gravimetric method works very well is in the Imperial Valley of California, USA. The East Mesa and Heber geothermal areas are characterized by areas of high density that are associated with dense cap rocks that form as a result of the hydrothermal activity. Horizontal electrical profiling also delineated low resistivity zones associated with the areas of high flow (Meidav and Fergerson, 1972). Seismic noise and magnetic methods were of no value in these two areas; the cultural noise level in the area prevents the recording of any potential natural noise.

Areas where the seismic noise method is effective are areas where active thermal manifestations occur at the surface such as the Norris Geyser Basin of Yellowstone National Park, USA (Oppenheimer and Iyer, 1979). In the basin, horizontal electrical profiling delineated the outlines of the near-surface hydrothermal alteration and Schlumberger soundings mapped a resistive layer

(75-130 ohm-m) overlain by a conductive layer (2-7 ohm-m) which represents the vapor-dominated and condensate-dominated portions of the geothermal reservoir, respectively (Zohdy et al., 1973). A similar geoelectric structure was determined for the Kawah Kamojang Field, Indonesia (Hochstein, 1975).

The VES and HEP methods have been successfully used in Reykjanes Peninsula of Iceland to locate thermal areas (Georgsson, 1981 and Georgsson, 1984). The three high temperature geothermal fields on the Peninsula all occur within areas of low subsurface resistivity (6 ohm-m) as opposed to a background of 10-12 ohm-m.

The dipole-dipole, VES and SP methods were effective in exploring the Cerro Prieto Field of Mexico. The electrical resistivity methods delineated a shallow zone of low resistivity (< 2 ohm-m) associated with high heat flow and located the trace of the producing fault (Garcia, 1975). The SP method was used to map a fault through the producing area (Corwin et al., 1980) and attenuation and velocity anomalies derived from the MEQ data were useful in delineating the geothermal field. A precision dipole-dipole survey was also used to map the intrusion of fresh water into the production zone (Wilt et al., 1983) and a seismic reflection survey determined a reflection attenuation within the production zone (Blakeslee, 1984).

KEY FINDINGS

The important findings of this study can be summarized as:

(1) The tabulated statistics on the utilization of geophysical methods for geothermal exploration are biased towards the geothermal methodology of the USA because the reference list is dominated by articles dealing with geothermal exploration within the USA (59% of the references fall within this category;

(2) A total of 88 resource areas/regions were reviewed and 562 instances of the use of geophysical methods for geothermal resources were recorded in table form;

(3) An average of 6 different geophysical methods were utilized in each of the 88 resource areas;

(4) An average of 5, 5 and 8 geophysical methods were used in low-, moderate- and high-temperature resource areas, respectively;

(5) The number of reported resource areas for a particular geologic setting increases with increasing resource temperature, except in the case of basin geologic settings;

(6) Most of the reported geothermal areas worldwide occur within the extrusive volcanic category;

(7) The VES, gravimetric and TG methods are used in over half of all resource areas and the heat flow, magnetic, MT, dipole-dipole reflection seismology, MEQ, remote sensing and bipole-dipole are used in 25% to 50% of all resource areas;

(8) The least-popular documented methods are the CSAMT, IP, pole-dipole and geomagnetic soundings which were used in less than 7% of the areas;

(9) The popoularity of geophysical methods varies with geologic setting

and temperature of the resource, but the VES and TG methods are clearly favored in the majority of areas and temperatures;

(10) The performance of the various geophysical techniques are very difficult to evaluate in the various geologic settings because of a lack of comprehensive case studies in the literature;

(11) Ward (1983) ranked the MEQ, gravimetric, electrical resistivity, SP and heat flow/TG methods as the most effective (good to fair) and the seismic noise, magnetic and MT methods as the least effective geophysical methods for the exploration of the Basin and Range Province of the western USA;

(12) In general the electrical resistivity methods appear to be the most effective reconnaissance method for the delineation of shallow geothermal reservoirs with near surface hydrothermal alteration zones; however, the specific geologic model determines which physical parameter has the best correlation with the geothermal reservoir and has the highest signal-to-noise ratio.

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LIST OF TABLES

TABLE I

PRINCIPAL LITERATURE SOURCES

Transactions and Proceedings of Technical Meetings:

Transactions of Geothermal Resources Council Meetings 1977 - 1984 (38) [202]

Abstracts of the Annual International SEG Meetings 1979 - 1984 (55) [53]

Proceedings of the Second UN Symposium on the Development and Utilization of Geothermal Resources, 1975 (49) [49]

Proceedings of the UN Symposium on the Development and Utilization of Geothermal Resources, 1970 (22) [22]

Proceedings of the International Congress on Geothermal Waters, Geothermal Energy and Volcanism of the Mediterranean Area: Geothermal Energy, 1976 (9([9]

Technical Journals:

Geophysics (25) [57]

Geothermics (14) [14]

Journal of Volcanology and Geothermal Research (8) [4]

Geoexploration (1) [1]

Geophysical Prospecting (1) [1]

Journal of Geophysical Research (12)

Publications:

The Role of Heat in the Development of Energy and Mineral Resources in the Northern Basin and Range

Geoelectric and Geothermal Studies (East Central Europe and Soviet Asia)

Note: values within the parentheses are the number of articles listed within the GEOREF bibliography; the values within the square brackets are the number of articles actually reviewed.

TABLE II

GEOPHYSICAL METHODS APPLIED TO GEOTHERMAL EXPLORATION

Abbreviation	Method	Geophysical Category
SN	seismic noise	passive seismic
MEQ	microearthquake	
T	teleseism	
RFL	reflection seismology	active seismic
RFR	refraction seismology	
GRAV	gravimetric	potential field
MAG	magnetic	
AMT	audiomagnetotelluric	passive EM
MT	magnetotelluric	
TEL	telluric	
GEOM	geomagnetic sounding	
CSAMT	controlled-source AMT	active EM
TEM	transient EM (time domain)	
CSFEM	controlled source frequency domain EM	
VES	vertical electrical soundings	electrical resistivity
HEP	horizontal electrical profiling	
DD	dipole-dipole (polar)	
BD	bipole-dipole	
PD	pole-dipole	
IP	induced polarization	
SP	self-potential	self-potential
RAD	radiometric	radiometric
HF	heat flow	temperature gradient
TG	temperature gradient	
STG	shallow temperature gradient	
BG	borehole geophysics	borehole geophysics
RS	remote sensing	remote sensing

TABLE III

WORLDWIDE USE OF GEOPHYSICAL METHODS FOR GEOTHERMAL EXPLORATION

LOCATION	RV	BR	VI	VI	BE	TEMP	SN	MEQ	T	RFL	RFR	GRAV	MAG	AMT	MT	TEL	GEOM	CSAMT	TEM	CSFEM	VES	HEP	DD	BD	PD	IP	SP	RAD	HF	TG	STG	BG	RS
RIFT VALLEYS																																	
EGYPT	X					+																							X	X			
ETHIOPIA	X					M						X												X									X
FRANCE RHINE GRABEN	X					+														+	X								X	X			
ISRAEL	X					+																							X				
KENYA	X					H		X				X									X		X	X							X		X
MEXICO CERRO PRIETO	X					H	X	X		X		X	X		X	X					X	X	X				X						X
REPUBLIC OF DJIBOUTI	X					+																					X						
UGANDA	X					M		X													X												
USA IMPERIAL VALLEY	X					H	X	X	X	X	X	X	X		X	X					X	X	X	X			X		X	X	X		
RIO GRANDE RIFT	X					L		X		X		X	X								X		X	X						X			
	X					M		X				X			X				X		X	X		X					X	X			
USSR BAIKAL RIFT	X					+									X														X				
W. GERMANY RHINE GRABEN	X					+							+		X																		

BASIN AND RANGE

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VALLES         : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
CALDERA        : | |X|X| :H:X|X| : | :X|X: |X|X| : | |X:X| |X|X| | : : : |X| : :X:
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               : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
YELLOWSTONE    : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
NAT. PARK      : | |X|X| :H:X|X|X: | :X|X: |+| | : | | :X|X| | | |X:X: : | | : :X:
-----------------------------------------------------------------------------------
USSR           : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
KAMCHATKA      : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
               : | | |X| :X: | | :X|X:X|X: |X| | : | | : | | | | | : : : | | :X: :
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EMEN ARAB      : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
EPUBLIC        : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
               : | | |X| :+: | | : | : | : | | | : | | :X| | | | | : : : | | : : :
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               BASINS
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AST GERMANY    : | | | |X:+: | | : | : | : |X| |X: | | : | | | | | : : :X| | : : :
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URKEY          : | | | |X:M: | | : | :X| : | | | : | | :X| | | | | : : : |X| : : :
               : | | | |X:H: | | :X| :X|X: | | | : | | :X| | | | | : : : |X| : : :
-----------------------------------------------------------------------------------
SA             : | | | |X:+: | | :+| : | : | | | : | |X:+| |+| | | : : :X|X| : : :
COLORADO       : | | | |X:+: |X|X: | :X|X: |X|X| : | | :X|X| | | | : : : |X| : : :
PLATEAU        : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
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               : | | | |X:L: | | : | :X|X: | | | : | | : | | | | | : : :X|X| : : :
MID-           : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
CONTINENT      : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
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ROCKY MTNS.    : | | | |X:M: | | :X|X:X| :X|X|X| : |X|X:X| |X|X| | :X: :X| |X: : :
               : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
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               : | | | |X:+:X| | : | : | : | | | : | | : | | | | | : : :+|+| : : :
PAIN           : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
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SSR            : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
EAST           : | | | |X:+: | | :+| : | : |X| |X: | | : | | | | | : : :X| | : : :
CARPATHIAN     : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
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               : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
UKRAINE        : | | | |X:+: | | : | : | : | | | : | | : | | | | | : : :X| | : : :
               : | | | | : : | | : | : | : | | | : | | : | | | | | : : : | | : : :
-----------------------------------------------------------------------------------
               UNCLASSIFIED
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BRAZIL						+																							X	X			
DENMARK						+																							X				
IRAQ						+																									X		
ROMANIA						+															+												
SWITZERLAND						+																							X				
W. GERMANY URACH KGRA						+				X	X				X		X																
YUGOSLAVIA						+																							X				
LOCATION	R	B	V	V	B	T	S	M	T	R	R	G	M	A	M	T	G	C	T	F	V	H	D	B	P	I	S	R	H	T	S	B	R
	V	R	I	E		E	N	E		F	F	R	A	M	T	E	E	A	E	E	E	E	D	D	D	P	P	A	F	G	T	G	S
						M		Q		L	R	A	G	T		L	O	M	M	M	S	P						D			G		
						P						V					M	T															

TABLE IV

UTILIZATION OF GEOPHYSICAL METHODS BY GEOLOGIC OCCURRENCE AND RESOURCE TEMPERATURE

RESOURCE OCCURRENCE	TEMP		# RESOURCES	SN	MEQ	T	RFL	RFR	GRAV	MAG	AMT	MT	TEL	GEOM	CSAMT	TEM	CSFEM	VES	HEP	DD	BD	PD	IP	SP	RAD	HF	TG	STG	BG	RS	ROW TOTALS
RIFT VALLEY	LOW	1	0	1	0	1	0	1	1	0	0	0	0	0	0	0	1	0	1	1	0	0	0	0	0	1	0	0	0	8	
	MOD	7	0	2	0	0	0	2	1	0	2	0	0	0	1	1	3	1	0	2	0	0	1	0	3	3	0	0	1	23	
	HI	5	2	3	1	2	1	3	2	0	3	2	0	0	0	0	3	2	3	2	0	0	2	0	3	1	2	0	2	39	
	SUBTOTAL		13	2	6	1	3	1	6	4	0	5	2	0	0	1	1	7	3	4	5	0	0	3	0	6	5	2	0	3	70
BASIN AND RANGE	LOW	1	0	0	0	0	0	1	1	1	1	0	0	0	0	1	1	0	0	0	0	0	0	0	0	1	1	0	1	9	
	MOD	2	1	1	0	1	0	1	1	1	2	1	1	0	1	1	1	0	1	1	0	0	1	1	2	1	1	0	1	22	
	HI	3	3	3	2	2	2	3	2	2	2	2	0	1	1	2	3	1	3	3	0	1	2	2	2	2	1	1	1	49	
	SUBTOTAL		6	4	4	2	3	2	5	4	4	5	3	1	1	2	4	5	1	4	4	0	1	3	3	4	4	3	1	3	80
INTRUSIVE VOLCANIC	LOW	2	0	0	0	1	0	1	2	0	0	0	0	0	0	0	1	1	0	0	0	0	1	0	1	1	0	0	0	9	
	MOD	7	1	1	0	1	0	3	2	1	1	0	1	0	0	0	3	0	1	0	0	0	0	1	3	4	0	2	0	25	
	HI	3	0	1	1	1	1	2	2	2	2	0	0	0	1	2	3	0	3	2	1	0	1	1	1	3	0	0	2	32	
	SUBTOTAL		12	1	2	1	3	1	6	6	3	3	0	1	0	1	2	7	1	4	2	1	0	2	2	5	8	0	2	2	66
EXTRUSIVE VOLCANIC	LOW	4	0	0	0	2	1	1	0	0	0	0	0	0	0	1	2	1	1	0	0	0	0	0	1	2	0	0	1	13	
	MOD	19	2	3	1	6	5	8	4	6	4	3	1	0	3	0	11	4	5	3	1	0	7	1	9	7	1	1	8	104	
	HI	24	7	12	4	8	5	15	14	5	10	6	0	0	4	6	15	8	9	8	1	1	5	2	10	12	1	2	8	178	
	SUBTOTAL		47	9	15	5	16	11	24	18	11	14	9	1	0	7	7	28	13	15	11	2	1	12	3	20	21	2	3	17	295
BASINS	LOW	3	1	0	0	1	0	1	0	0	0	0	0	0	0	1	1	0	1	0	0	0	0	0	3	3	0	0	0	12	
	MOD	6	0	1	1	2	1	3	1	1	4	4	2	0	1	1	3	1	1	1	0	0	1	0	4	2	1	0	0	34	
	HI	1	0	0	0	1	0	1	1	0	0	0	0	0	0	0	1	0	0	0	0	0	0	0	0	1	0	0	0	5	
	SUBTOTAL		10	1	1	1	4	1	5	2	1	4	4	2	0	1	2	5	1	2	1	0	0	1	0	7	6	1	0	0	51
TOTAL	LOW	11	1	1	0	5	1	5	4	1	1	0	0	0	0	3	6	2	3	1	0	0	1	0	5	8	1	0	2	51	
	MOD	41	4	8	2	10	6	17	9	9	13	8	5	0	6	3	21	6	8	7	1	0	10	3	21	17	3	3	10	208	
	HI	36	12	19	8	14	9	24	21	9	17	10	0	1	6	10	25	11	18	15	2	2	10	5	16	19	4	3	13	303	
COLUMN TOTALS		88	17	28	10	29	16	46	34	19	31	18	5	1	12	16	52	19	29	23	3	2	21	8	42	44	8	6	25	562	

TABLE V

PERCENT UTILIZATION OF GEOPHYSICAL METHODS BY GEOLOGIC OCCURRENCE AND RESOURCE TEMPERATURE

RESOURCE OCCURRENCE	TEMP	# RESOURCES	SN	MEQ	T	RFL	RFR	GRAV	MAG	AMT	MT	TEL	GEOM	CSAMT	TEM	CSFEM	VES	HEP	DD	BD	PD	IP	SP	RAD	HF	TG	STG	BG	RS
RIFT	LOW	1	0	100	0	100	0	100	100	0	0	0	0	0	0	100	0	100	0	0	0	0	0	0	100	0	0	0	0
VALLEY	MOD	7	0	29	0	0	0	29	14	0	29	0	0	0	14	14	43	14	0	29	0	0	14	0	43	43	0	0	14
	HI	5	40	60	20	40	20	60	40	0	60	40	0	0	0	0	60	40	60	40	0	0	40	0	60	20	40	0	40
	ALL	13	15	46	8	23	8	46	31	0	38	15	0	0	8	8	54	23	31	38	0	0	23	0	46	38	15	0	23
BASIN AND	LOW	1	0	0	0	0	0	100	100	100	100	0	0	0	0	100	100	0	0	0	0	0	0	0	0	100	100	0	100
RANGE	MOD	2	50	50	0	50	0	50	50	50	100	50	50	0	50	50	50	0	50	50	0	0	50	50	100	50	50	0	50
	HI	3	100	100	67	67	67	100	167	67	67	67	0	33	33	67	100	33	100	100	0	33	67	67	67	67	33	33	33
	ALL	6	67	67	33	50	33	83	67	67	83	50	17	17	33	67	83	17	67	67	0	17	50	50	67	67	50	17	50
INTRUSIVE	LOW	2	0	0	0	50	0	50	100	0	0	0	0	0	0	0	50	50	0	0	0	0	50	0	50	50	0	0	0
VOLCANIC	MOD	7	14	14	0	14	0	43	29	14	14	0	14	0	0	0	43	0	14	0	0	0	0	14	43	57	0	29	0
	HI	3	0	33	33	33	33	67	67	67	67	0	0	0	33	67	100	0	100	67	33	0	33	33	33	100	0	0	67
	ALL	12	8	17	8	25	8	50	50	25	25	0	8	0	8	17	58	8	33	17	8	0	17	17	42	67	0	17	17
EXTRUSIVE	LOW	4	0	0	0	50	25	25	0	0	0	0	0	0	0	25	50	25	25	0	0	0	0	0	25	50	0	0	25
VOLCANIC	MOD	19	11	16	5	32	26	42	21	32	21	16	5	0	16	0	58	21	26	16	5	0	37	5	47	37	5	5	42
	HI	24	29	50	17	33	21	63	58	21	42	25	0	0	17	25	63	33	38	33	4	4	21	8	42	50	4	8	33
	ALL	47	19	32	11	34	23	51	38	23	30	19	2	0	15	15	60	6	32	23	4	2	26	6	43	45	4	6	36
	LOW	3	33	0	0	33	0	33	0	0	0	0	0	0	0	33	33	0	33	0	0	0	0	0	3	3	0	0	0
BASINS	MOD	6	0	17	17	33	17	50	33	33	67	67	33	0	17	17	50	17	17	17	0	0	17	0	67	33	17	0	0
	HI	1	0	0	0	100	0	100	100	0	0	0	0	0	0	0	100	0	0	0	0	0	0	0	0	100	0	0	0
	ALL	10	10	10	10	40	10	50	20	10	40	40	20	0	10	20	50	10	20	10	0	0	10	0	70	60	10	0	0
	LOW	11	9	9	0	45	9	45	36	9	9	0	0	0	0	27	55	18	27	0	0	0	9	0	45	73	9	0	18
SUBTOTAL	MOD	41	10	20	5	24	15	41	22	22	32	20	12	0	15	7	51	15	20	17	2	0	24	7	51	41	7	7	24
	HI	36	33	53	22	39	25	67	58	25	47	28	09	3	17	28	69	31	50	42	6	6	28	14	44	53	11	8	36
	TOTAL	38	99	32	11	33	18	52	39	22	35	20	6	11	14	18	59	22	33	26	3	2	24	9	48	50	9	7	28

TABLE VI

SUMMARY OF THE PERCENT UTILIZATION OF GEOPHYSICAL METHODS BY GEOLOGIC OCCURRENCE AND RESOURCE TEMPERATURE

RESOURCE OCCURRENCE	TEMP	# RESOURCES	SN	MEQ	T	RFL	RFR	GRAV	MAG	AMT	MT	TEL	GEOM	CSAMT	TEM	CSFEM	VES	HEP	DD	BD	PD	IP	SP	RAD	HF	TG	STG	BG	RS
RIFT VALLEY	LOW	1	D	A	D	A	D	A	A	D	D	D	D	D	D	D	A	D	A	A	D	D	D	D	D	A	D	D	D
	MOD	7	D	B	D	D	D	B	C	D	B	D	D	D	C	C	B	C	D	B	D	D	C	D	B	B	D	D	C
	HI	5	B	A	C	B	C	A	B	D	A	B	D	D	D	D	A	B	A	B	D	D	B	D	A	C	B	D	B
	ALL	13	C	B	C	C	C	B	B	D	B	C	D	D	C	C	A	C	B	B	D	D	C	D	B	B	C	D	C
BASIN AND RANGE	LOW	1	D	D	D	D	D	A	A	A	A	D	D	D	D	A	A	D	D	D	D	D	D	D	D	A	A	D	A
	MOD	2	A	A	D	A	D	A	A	A	A	A	A	D	A	A	A	D	A	A	D	D	A	A	A	A	A	D	A
	HI	3	A	A	A	A	A	A	A	A	A	A	D	B	B	A	A	B	A	A	D	B	A	A	A	A	B	B	B
	ALL	6	A	A	B	A	B	A	A	A	A	A	C	C	B	A	A	C	A	A	D	C	A	A	A	A	A	C	A
INTRUSIVE VOLCANIC	LOW	2	D	D	D	A	D	A	A	D	D	D	D	D	D	D	A	A	D	D	D	D	A	D	A	A	D	D	D
	MOD	7	C	C	D	C	D	B	B	C	C	D	C	D	D	D	B	D	C	D	D	D	D	C	B	A	D	B	D
	HI	3	D	B	B	B	B	A	A	A	A	D	D	D	B	A	A	D	A	A	B	D	B	B	B	A	D	D	A
	ALL	12	C	C	C	B	C	A	A	B	B	D	C	D	C	C	A	C	B	C	C	D	C	C	B	A	D	C	C
EXTRUSIVE VOLCANIC	LOW	4	D	D	A	A	B	B	D	D	D	D	D	D	D	B	A	B	B	D	D	D	D	D	B	A	D	D	B
	MOD	19	C	C	C	B	B	B	C	B	C	C	C	D	C	D	A	C	B	C	C	D	B	C	B	B	C	C	B
	HI	24	B	A	C	B	C	A	A	C	B	B	D	D	C	B	A	B	B	B	C	C	C	C	B	A	C	C	B
	ALL	47	C	B	C	B	C	A	B	C	B	C	C	D	C	C	A	C	B	B	C	C	B	C	B	B	C	C	B
BASINS	LOW	3	B	D	D	B	D	B	D	D	D	D	D	D	D	B	B	D	B	D	D	D	D	D	C	C	D	D	D
	MOD	6	D	C	C	B	C	A	B	B	A	A	B	D	C	C	A	C	C	C	D	D	C	D	A	B	C	D	D
	HI	1	D	D	A	A	D	A	A	D	D	D	D	D	D	D	A	D	D	D	D	D	D	D	D	A	D	D	D
	ALL	10	C	C	B	B	C	A	C	C	B	B	C	D	C	C	A	C	C	C	D	D	C	D	A	A	C	D	D
SUBTOTAL	LOW	11	C	C	D	B	C	B	B	C	C	D	D	D	D	B	A	C	B	C	D	D	C	D	B	A	C	D	C
	MOD	41	C	C	C	C	C	B	C	C	B	C	C	D	C	C	A	C	C	C	C	D	C	C	A	B	C	C	C
	HI	36	B	A	C	B	B	A	A	B	B	B	D	C	C	B	A	B	A	B	C	C	B	C	B	A	C	C	B
	TOTAL	88	C	B	C	B	C	A	B	C	B	C	C	C	C	C	A	C	B	B	C	C	C	C	B	A	C	C	B

A = utilization ≥ 50% B = 25% ≤ utilization < 50% C = 0% < utilization 25% D = 0%

TABLE VII

DISTRIBUTION OF ARTICLES AS A FUNCTION OF COUNTRY

LOCATION	
ALGERIA	1
AUSTRALIA	3
BRAZIL	1
CANADA	9
CARIBBEAN SEA	1
CHILE	3
CHINA	1
COSTA RICA	3
CZECHOSLAVAKIA	2
DENMARK	1
EAST GERMANY	2
EGYPT	2
EL SALVADOR	2
ENGLAND	1
ETHIOPIA	4
FIJI	0
FRANCE	10
GREECE	2
HUNGARY	3
ICELAND	13
INDIA	10
INDONESIA	9
IRAQ	0
ISRAEL	0
ITALY	30
JAPAN	21
KENYA	2
MEXICO	21
NEW GUINEA	1
NEW ZEALAND	12
NICARAGUA	2
PERU	1
PHILIPPINES	2
PORTUGAL	2
REPUBLIC OF DJIBOUTI	0
ROMANIA	0
SPAIN	1
SWITZERLAND	1
TAIWAN	7
THAILAND	1
TURKEY	6
UGANDA	1
USA	340
USSR	25
WEST GERMANY	4
YEMEN ARAB REPUBLIC	0
YUGOSLAVIA	1

TABLE VIII

THE SEVEN COUNTRIES REFERENCED THE MOST WITHIN THE AUGMENTED BIBLIOGRAPHY*

COUNTRY	NO. REFERENCES	PERCENTAGE OF TOTAL NO. OF REFERENCES
USA	340	59
ITALY	30	5
USSR	25	4
JAPAN	21	4
MEXICO	21	4
ICELAND	13	2
NEW ZEALAND	12	2
ELSEWHERE	113	20
TOTAL	575	100

* Based on the GEOREF bibliography and the articles from Geoelectric and Geothermal Studies (Adams, 1976).