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GEOTHERMAL ENERGY

AN OVERVIEW OF OCCURRENCE AND EXPLORATION

by

Phillip M. Wright
Director
Earth Science Laboratory
University of Utah Research Institute
Salt Lake City, Utah, USA

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INTRODUCTION

Development of geothermal resources if being aggressively pursued on a worldwide basis. Approximately 3 800 MW of electricity are currently being generated from geothermal energy, and about 10 000 thermal MW are being used for direct heat applications. While this may seem small compared to the estimated 8.4 x 10^6 MW of human use of fossil energy (Williams and Von Herzen, 1974), it nevertheless represents a savings in the consumption of about 77 million barrels of oil per year worldwide. It is very difficult to estimate the ultimate potential contribution of geothermal energy to mankind's needs for at least three reason: 1) long-range future energy costs, although generally predicted to be higher than today's levels, are uncertain, and a large number of lower-grade geothermal resources would become economic at higher energy prices; 2) only preliminary estimates of the worldwide resource base have been made, and; 3) technology for using energy in magma, hot rock and normal thermal-gradient resources, whose potential contributions are very large, is not yet available.

Geothermal energy is heat that originates within the earth. The earth is an active thermal engine, and 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 phenomena as motion of the earth's crustal plates, uplifting of mountain ranges, occurrence of earthquakes, eruption of volcanos and spouting of geysers all owe their origin to the transport of internal thermal energy.

Although the feasibility of use of geothermal energy has been known for many years, the total amount of application today is small compared with the potential for application. The present availability of less expensive energy from fossil fuels has suppressed use of geothermal energy at all but a few of the highest-grade resources. Research and development of new techniques and equipment is needed to decrease costs of exploration, drilling, reservoir evaluation and extraction to make the vastly more numerous lower-grade resources also economic.

The objective of this paper is to present an overview of the nature of and exploration for geothermal resources.

NATURE AND OCCURRENCE OF GEOTHERMAL RESOURCES

Although the distribution with depth in the earth of density, pressure and other related physical parameters is 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, 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. Using present technology and under favorable conditions, holes can be drilled to depths of about 10 km, where temperatures range upward from about 150°C in areas underlain by cooler rocks to perhaps 600°C in exceptional areas.

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 is 61×10^{-3} watts/m^2 for the continents and 92×10^{-3} watts/m^2 for the oceans, including effects of sea-floor spreading discussed below, and since the mean surface area of the earth is about 5.1×10^{14} m^2, the rate of heat loss is about 42×10^{12} watts (42 million megawatts), a very large amount indeed (Williams and Von Herzen, 1974). 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 82 milliwatts/m^2 is about 50,000 times smaller than the flux of heat from the sun (much of which is reflected or re-directed into space), 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 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. The relative contribution to the observed surface heat flow of these two mechanisms is not yet resolved. Some theoretical

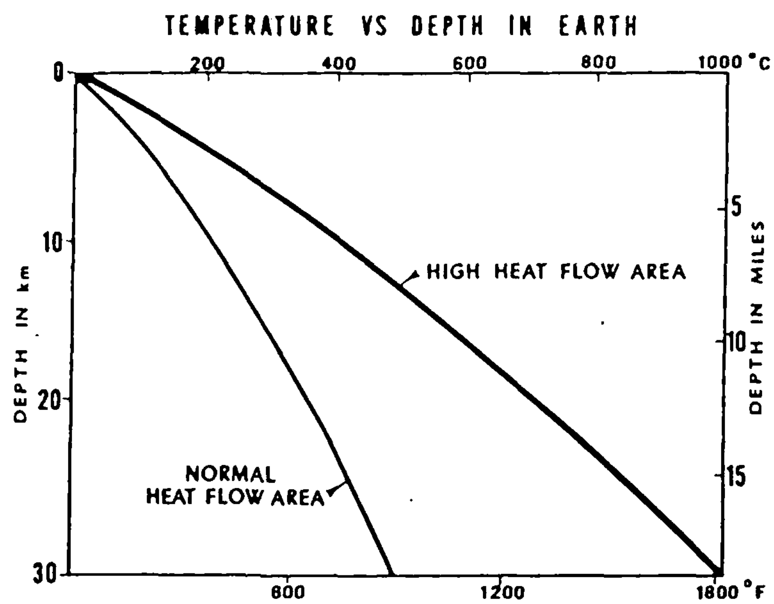
TEMPERATURE VS DEPTH IN EARTH
200
400
600
800
1000 °C
0
10
20
30
DEPTH IN km
5
10
15
DEPTH IN MILES
HIGH HEAT FLOW AREA
NORMAL
HEAT FLOW AREA
600
1200
1800 °F

FIGURE 1

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 4.7 billion years ago. We do know that the thermal conductivity of crustal rocks is low so that heat escapes from the surface very 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. White (1965) has estimated that the total heat stored above surface temperature in the earth to a depth of 10 km is about 1.3×10^{27} J, equivalent to the burning of about 2.3×10^{17} barrels of oil. It is apparent that if even a small part of this heat could be made available, its contribution would be significant.

Geological Processes

The fundamental cause of geothermal resources lies in the transport of heat near to the surface through one or more of a number of geological processes. We have seen that the ultimate source of that heat is in the interior of the earth where temperatures are much higher than they are at the surface. Geothermal resources commonly have three components:

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

We will now discuss some of the geological aspects of each of these factors.

Heat Source. Geothermal resource areas, or geothermal areas for short, are generally those in which higher temperatures are found at shallower depths than is normal. This condition usually results from either 1) intrusion of molten rock to high levels in the earth's crust, 2) higher-than-average flow of heat to the surface with an attendant high rate of increase of temperature with depth (geothermal gradient) as illustrated in Figure 1, often in broad areas where the earth's crust is thin, 3) heating of ground water that circulates to depths of 2 to 5 km with subsequent ascent of the thermal water to the surface, or 4) anomalous heating of a shallow rock body by decay of an unusually high content of radioactive elements. In many geothermal areas, heat is brought right to the surface by circulation of ground water. If temperature is high enough, steam may be produced, and geysers, fumaroles, and hot springs are common surface manifestations of underlying geothermal reservoirs.

The distribution of geothermal areas on the earth's surface is not random but instead is governed by geological processes of global, regional and local scale. 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, our learning rate is high.

Figure 2 shows the principal areas of known geothermal occurrences on a world map. Also indicated are areas of young volcanos 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 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 3. 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 next layer, the mantle, which lies below the crust. Mantle material below the lithosphere is less rigid 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 mixture of the same metals.

One unifying geological process that generates heat sources is known 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 convec-

CONCEPT OF PLATE TECTONICS

(NOT TO SCALE)

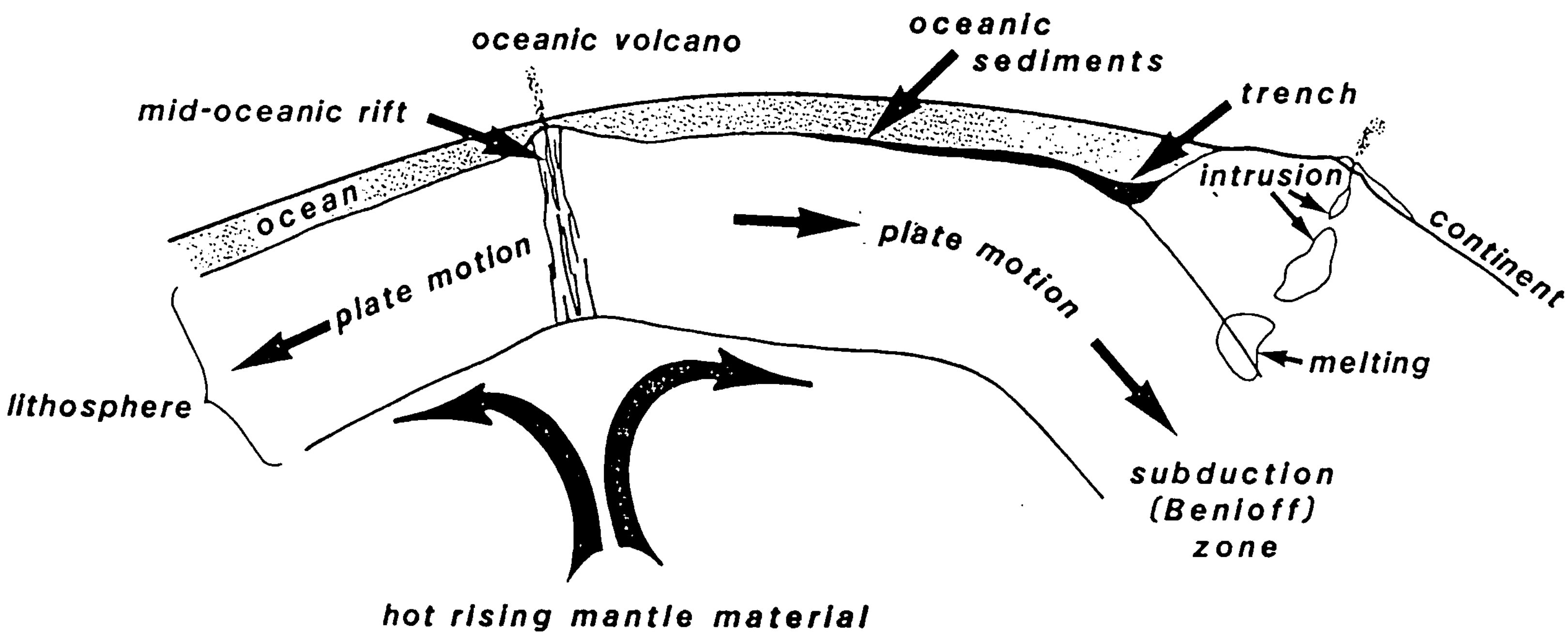

Figure 4

INTERIOR OF THE EARTH

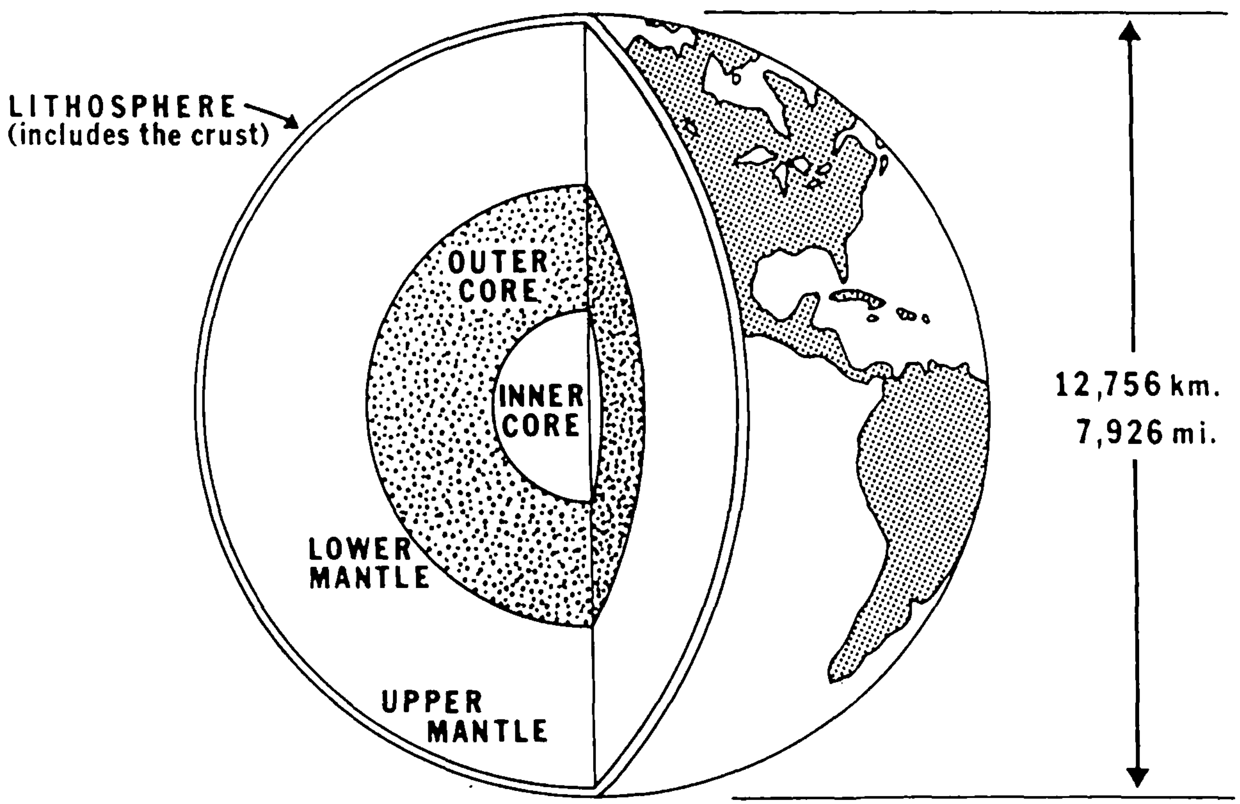

Figure 3

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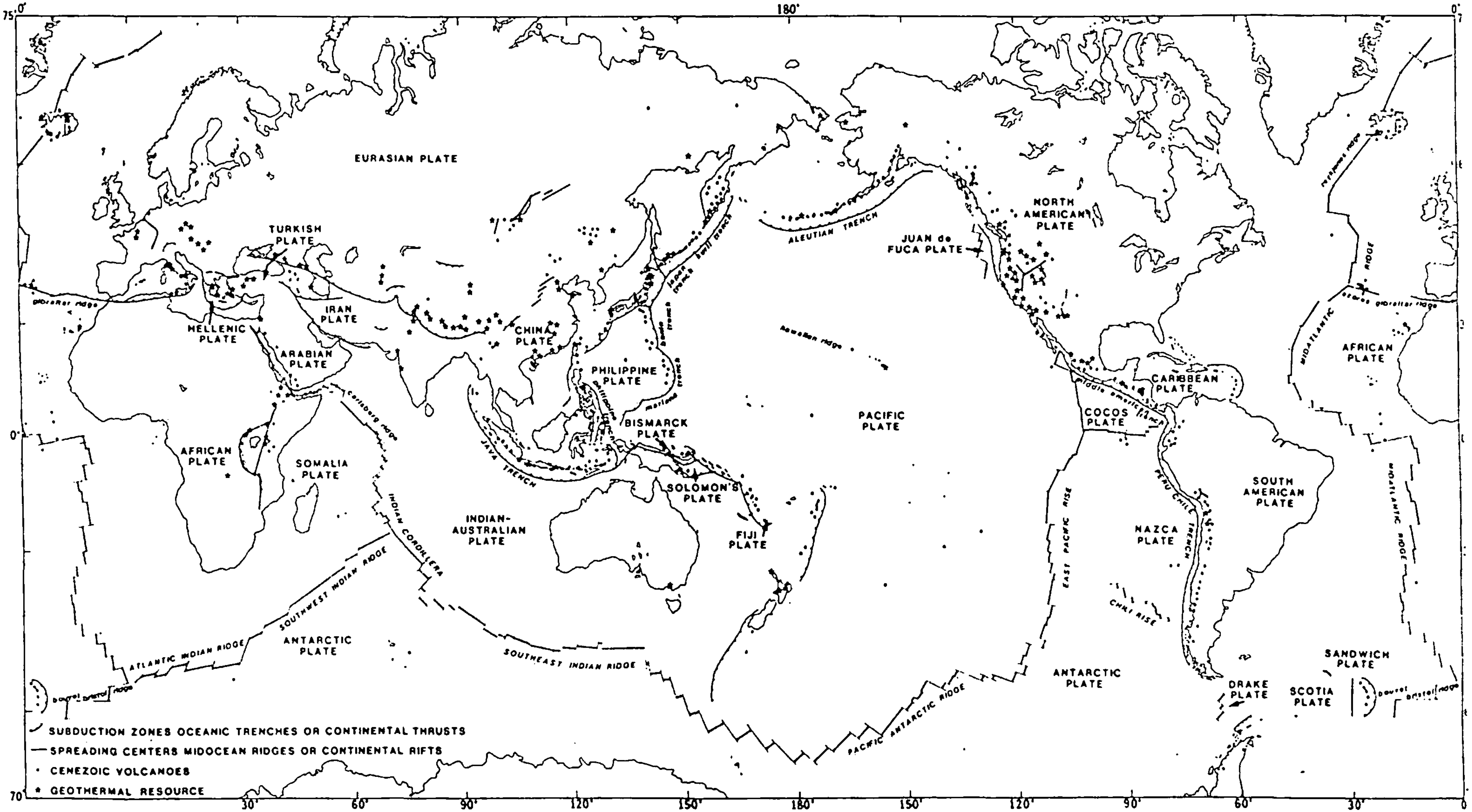

GEOTHERMAL RESOURCES AND PLATE TECTONIC FEATURES

Figure 2

tion 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 is dragged 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. 2 and 4), typified by the mid-Atlantic Ridge and the East Pacific Rise. 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. This intrusion of molten material brings large quantities of heat to shallow depths and is the heat source for the recently discovered oceanic hydrothermal systems. The laterally moving oceanic lithospheric plates impinge against adjacent plates, some of which 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 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. The upper layers of the descending plate often contain oceanic sediments rich in water, thus assisting in the melting process. The molten or partially molten rock bodies (magmas) ascend buoyantly through the crust (Figs. 4 and 5), probably along lines of structural weakness, and carry their contained heat to within 1.5 to 20 km of the surface. They may give rise to volcanos if part of the molten material escapes to the surface through faults and fractures in the upper crust. These shallow crustal intrusions occur on the landward side of oceanic trenches, usually 50 km to 200 km inland (Fig. 4). They are the cause of the volcanos in the Cascade Range of California, Oregon and Washington, for example, and of those of Central and South America. A number of these volcanic areas have geothermal systems associated with them.

Figure 2 shows where these processes of crustal spreading, formation of

CRUSTAL INTRUSION

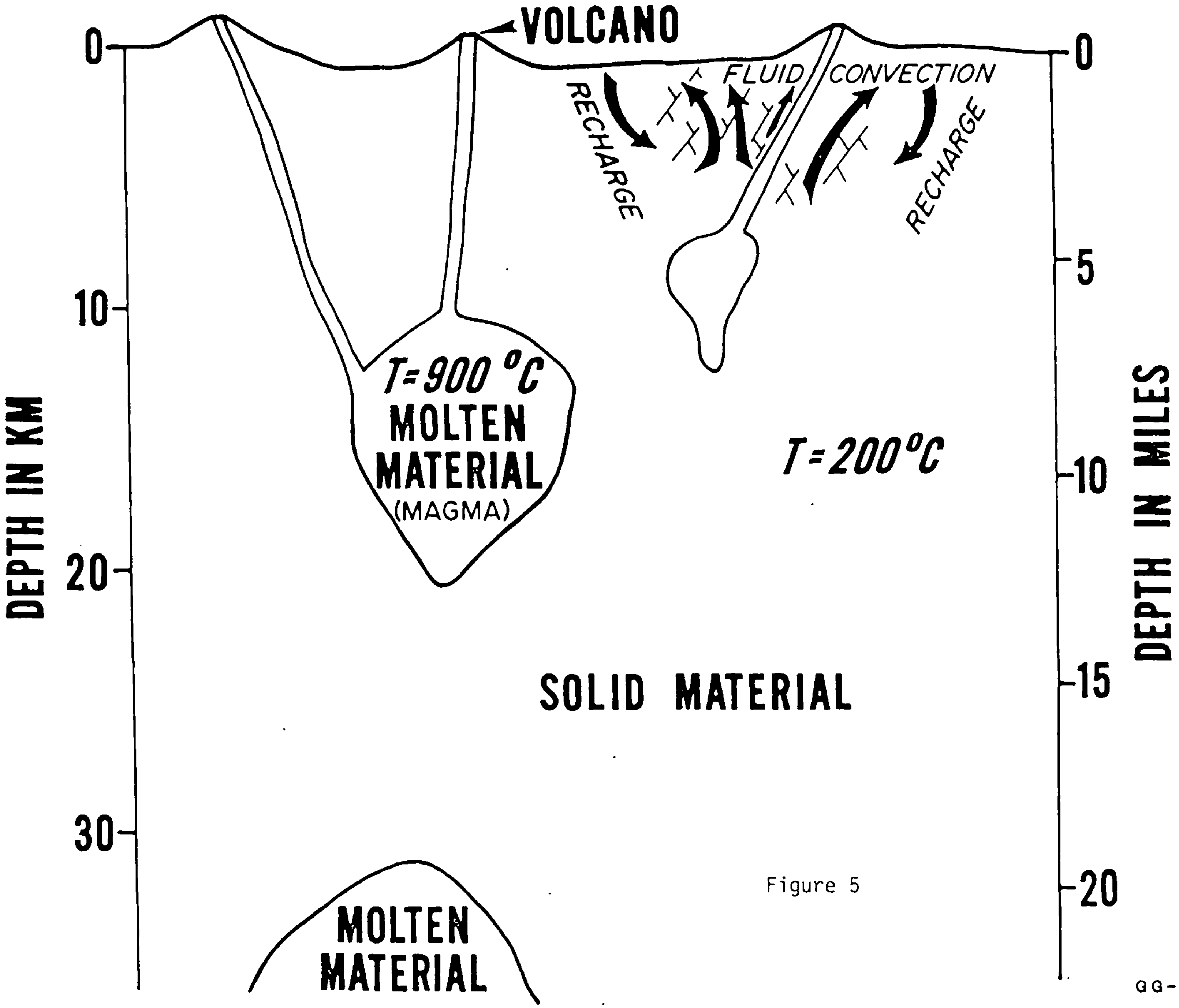

Figure 5

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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 consumed in the subduction zones, usually marked by trenches. 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 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 in somewhat more detail the development of crustal intrusions, 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. Intrusion of 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 (hundreds of square miles) 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 compositional 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 can be seen in the rocks that compose the Hawaiian Islands. The crystalline, plutonic variety of this rock that has cooled slowly and 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 also used for any crystalline igneous rock. Magmas that result in basalt or gabbro are termed "mafic" or "basic" whereas magmas that result in rhyolite or granite are termed "felsic" or "acidic".

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 seem to indicate a more or less direct pipeline for the magma from the upper mantle to the surface.

The origin of granites is a subject of some controversy. It can be shown that felsic magma may be derived by progressive segregation of the melt fraction of a basaltic magma as it cools and begins to crystallize. However, the chemical composition of granites is much like the average composition of the continental crust, and some granites probably also result from melting of crustal rocks by upwelling basaltic magmas. Basaltic magmas melt at a higher temperature and are less viscous (more fluid) than granitic magmas. Occurrence on the surface or in drill samples of felsic 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 they may indicate a large body of viscous magma at depth to provide a geothermal heat source. On the other hand, occurrence of young basaltic rocks 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 (Smith and Shaw, 1975). 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.

Another important source of volcanic rocks are the hypothesized point sources of heat in the mantle as contrasted with the rather large convection cells that drive plate motions. 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 volcanos is developed. Young volcanic rocks occur at one end of the chain with older ones at the other end. The Hawaiian Island chain is an example. The youngest volcanic rocks on the island of Kauai to the northwest end have been dated through radioactive means at about 4 million years, whereas the volcanos Mauna Loa and Mauna Kea on the island of Hawaii at the southeast end of the chain are forming today and are in almost continual eruptive activity. A new island southeast of Hawaii is being formed by suboceanic volcanic eruptions and at present is only a few hundred feet beneath the surface. To the northwest, the Hawaiian chain continues beyond Kauai for more than 2000 miles to Midway Island, where the last volcanic activity was about 16 million years ago. The trace of the island chain is consistent with the motions of the pacific plate postulated by geophysicists. Geologists also 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.

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 (40°C/km to 60°C/km). Geophysical and geological data indicate that the earth's crust is thinner than normal in the Basin and Range province and that the isotherms are warped upward beneath this area. Much of the western U. S. is geologically active, as manifested by earthquakes and active or recently active volcanos. Faulting and fracturing during earthquakes help to keep fluid pathways 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 the surface. Many of the hot springs and wells in the western United States and elsewhere owe their origin to such processes.

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 minerals themselves but rather it flows in pores between mineral grains and in open spaces created by fractures and faults. Porosity is the term given to the fraction or percent of void space in a volume of rock.

Permeability and porosity can be primary or secondary, i.e. formed with the rock or subsequently. Primary permeability in sedimentary rocks originates from intergranular porosity and it usually decreases with depth due to compaction and cementation. In volcanic sequences, primary intergranular porosity and permeability exist, but primary permeability also exists in open spaces at contacts between individual flows and within the flows themselves. Secondary permeability occurs in open fault zones, fractures and fracture intersections, along dikes and in breccia zones produced by hydraulic fracturing (Brace, 1968; Moore et al., 1985). Changes in permeability may also come about through mineral deposition by leaching by the thermal fluids.

Regarding exploration for hydrothermal systems, the key problem appears to be more in locating permeable zones than in locating high temperatures. Fractures sufficient to make a well a good producer need be only a few millimeters in width, but must be connected to the general fracture network in the rock in order to carry large fluid volumes. Grindly and Browne (1976) note that of 11 hydrothermal fields investigated in New Zealand, all of which have high temperatures (230°C to 300°C), five are non-productive chiefly because of low permeability. Three of the eleven fields are in production (Wairakei, Kawerau and Broadlands) and in each of these, permeability limits production more than temperature does.

Permeabilities in rocks range over 12 orders of magnitude. Permeabilities in pristine, unfractured crystalline rock are commonly on the order of 10^{-6} darcy (1μd) or less. However, in situ measurements at individual sites may vary by as much as 4 to 6 orders of magnitude, and zones of >100 md are commonly encountered. These higher permeabilities are due to increased fracture density. Fracture permeability may be inferred provided information is available on the spacing, continuity, aperture and orientation of fractures. Permeability in igneous rocks may decrease with depth, but its behavior is not systematic. Increased temperature has been shown experimentally both to increase and decrease permeability; these conflicting results arise from the complex interplay of thermal expansion, thermal stress cracking, and dissolution and redeposition of mineral phases.

Most geothermal systems are structurally controlled, i.e. the magmatic heat source has been emplaced along zones of structural weakness in the crust. Permeability has usually been increased in the vicinity of the intrusion from fracturing and faulting in response to stresses involved in the intrusion process itself and in response to regional stresses. Thus, an understanding of the geologic structure of a resource area can lead not only to evidence for the most likely location of a subsurface magma chamber, but also to inferences about areas of higher permeability at depth. Such areas would be prime geothermal exploration targets.

Heat Transfer Fluid. The purpose of the heat transfer fluid is to remove heat from the rocks at depth and bring it to the surface. This fluid is either water (usually saline) or steam. Water has a high heat capacity (amount of heat absorbed or released when the temperature increases or decreases by 1°C, 80 cal/gm) and a high latent heat of vaporization (amount of heat needed to convert water to steam or released when steam condenses, 540 cal/gm). Thus water, which pervades fractures and other open spaces in rocks and so is available in nature, is an ideal heat transfer fluid because a given quantity of water or steam can carry a relatively large amount of heat to the surface where it is easily removed.

The density and viscosity of water both decrease as temperature increases. Thus, water heated at depth in the earth is lighter than is cold water in surrounding rocks, and is therefore subjected to buoyant forces. If the heating is great enough that the buoyant forces overcome the resistance to flow imposed by the rock, the heated water will rise toward the earth's surface. As it does so, cooler water will move in to replace it, be heated and also rise. Because heated water will flow along paths of least resistance, it may also move laterally in places, but its net flow will be upward. In this way, natural convection can be set up in the groundwater above a source of heat such as an intrusion. This convective process can bring large quantities of heat near enough to the earth's surface to be reached by wells, and is thus responsible for the most economically important class of geothermal resources, as we shall discuss below.

Water can also form steam in the earth, and the process can be quite complex. The transition of water to its gaseous form occurs at a temperature which is a function of pressure (Figure 6). This transition absorbs a considerable latent heat (540 cal/gm at 100°), and the latent heat appreciably diminishes when the temperature increases. The density of steam, like all

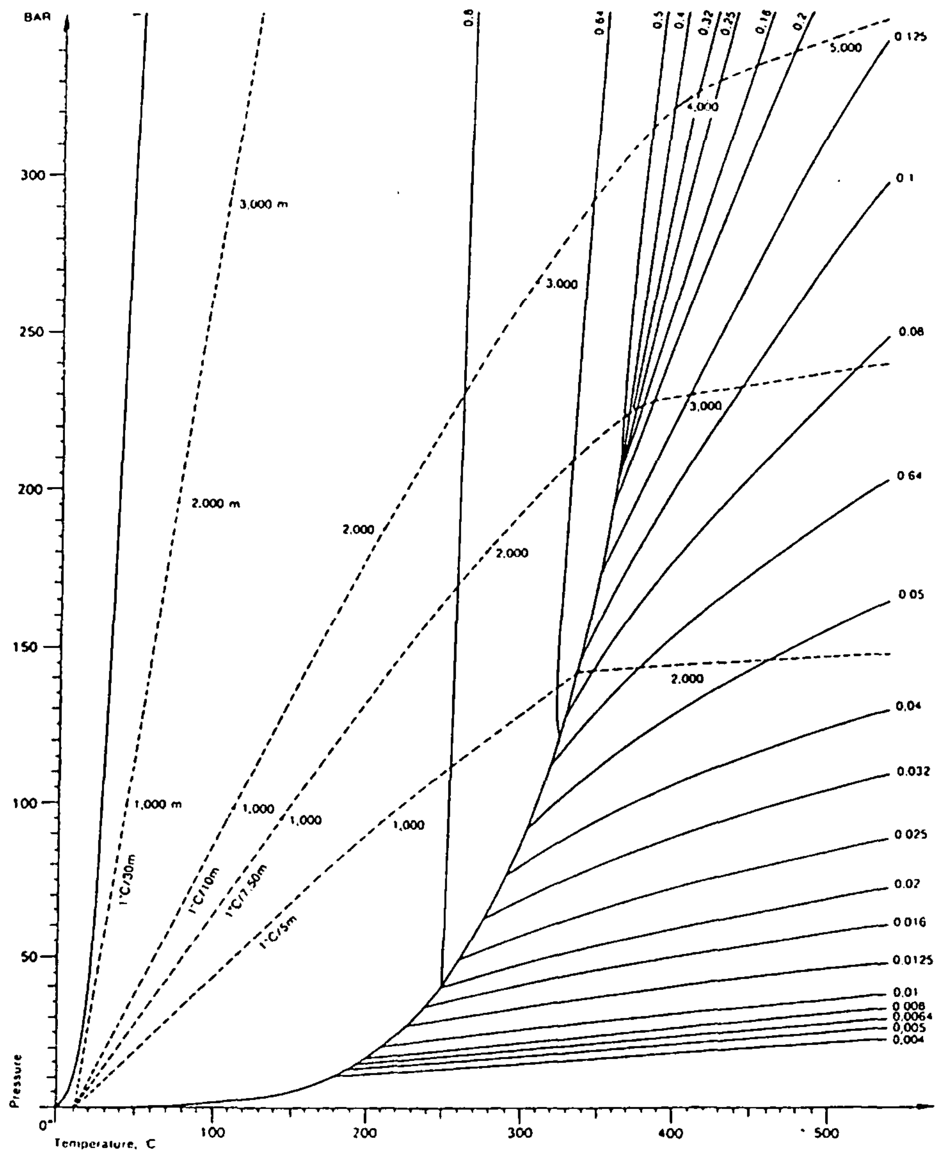

FIGURE 6. Graph of water density as a function of temperature and pressure. The dashed lines indicate the states as a function of the depth for a fluid phase in static equilibrium with a temperature of 11°C at the phreatic level and the different values of the thermal gradient, supposed uniform, taking account of the water density. Note that the critical state of water is reached for a gradient of about 1°C/7.5 m at a depth of 2,700 m. For a higher gradient, the hypothesis of a uniform thermal gradient becomes incompatible with the hypothesis of a static fluid equilibrium, this equilibrium being unstable.

gases, diminishes when the temperature increases and increases with pressure. At the critical point (T = 375°C, pressure = 221 bars), the specific volume of steam becomes equal to that of liquid water, the latent heat falls to zero, and there are no longer any differences between the two phases. For temperatures and pressures with higher values than the critical point, the supracritical domain, there exists only a single fluid phase whose density varies in a continuous manner, as shown in Figure 6. There are tables for water as well as for steam which furnish all the pertinent characteristics as functions of temperature and pressure. The viscosity values appear to be the least well known in the supracritical domain.

If the water contains dissolved gas, the gas will accumulate in the vapor phase, where the total pressure will be the sum of the partial pressures of the gases and the water vapor. On the other hand, dissolved salts are distributed very unequally between the two phases, nearly all being found in the liquid.

In some convective hydrothermal resources, the temperature never exceeds the boiling point for the pressure at any particular depth, and the system does not generate steam. However, in other systems the temperature can rise above the local boiling point, and steam is produced. The steam ascends and meets cooler rocks where it partially condenses while heating the rocks, and the pressure drop due to condensation brings up more steam. Cooler water descends alongside the steam cell and is heated by hotter rocks at depth until it finally vaporizes at the level of boiling. In this way, steam convection is set up. Several small steam plumes or convection cells may be present to start, but we predict the evolution toward a smaller number of important ascenting cells, each producing a substantial flow of vapor. If there is a sealed zone, a zone of low permeability, above the steam cells, steam will accumulate in the reservoir. The temperature and pressure in such a steam reservoir vary only slowly with depth. At Larderello, Italy, the reservoir temperature and pressure are 240°C and 35 bars, values that appear to be typical of other vapor-dominated systems.

With the foregoing material as background, we are now in a position to develop a classification for geothermal systems and to describe their workings in more detail.

Classification of Geothermal Resources

The classifications of geothermal resource types shown in Table 1 is modeled after one given by White and William (1975). Each resource type will be described briefly with emphasis on those that are presently nearest to commercial use in the U.S. In order to describe these resources, 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. In spite of disagreement over details, however, the models predicted below are generally acceptable and aid our thinking on the topic.

Geothermal resource temperatures range upward from the mean annual ambient temperature (usually 10-30°C) to well over 350°C. Figure 7 shows the span of temperatures of interest in geothermal work. It will be helpful for the reader to refer to this figure during the subsequent discussions.

Convective Hydrothermal Resources. Convective hydrothermal 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 (Fig. 5) that is 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 resource depends, among other less important variables, on temperature and pressure conditions at depth.

Figure 8 (after White et al., 1971) shows a simple conceptual model of a hydrothermal system where steam is the pressure-controlling fluid phase, a so-called vapor-dominated hydrothermal system. 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. This removes the latent heat of vaporization from this level, thereby cooling the rock and allowing more heat to rise from depth. Steam moves upward through fractures in the rock and is possibly superheated by the hot surrounding rock. At the top and sides of the system, heat is lost from the vapor to the

TABLE 1

GEOTHERMAL RESOURCE CLASSIFICATION
(Modified from White and Williams, 1975)

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

GEOTHERMAL TEMPERATURES

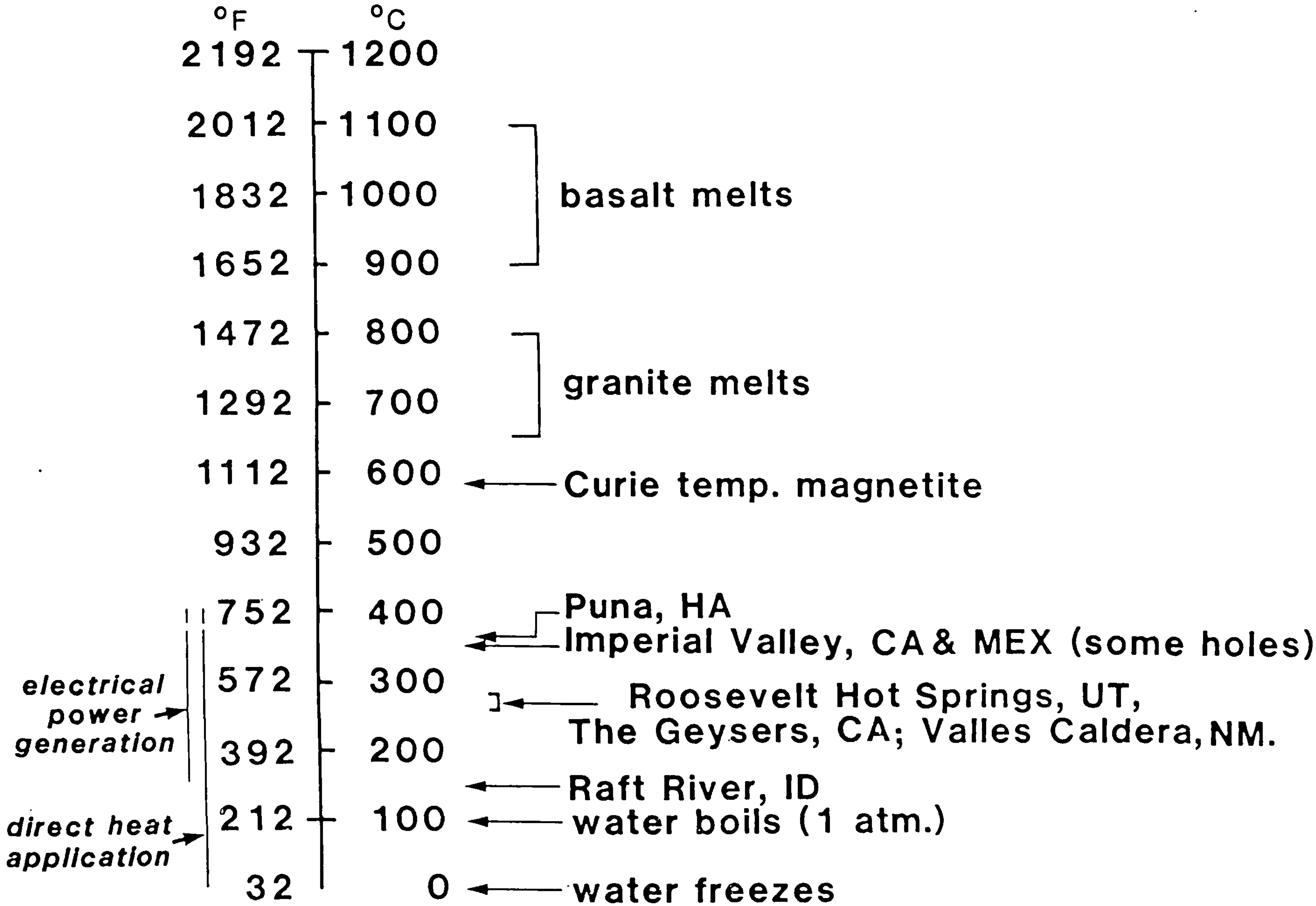

Figure 7

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VAPOR DOMINATED GEOTHERMAL RESERVOIR

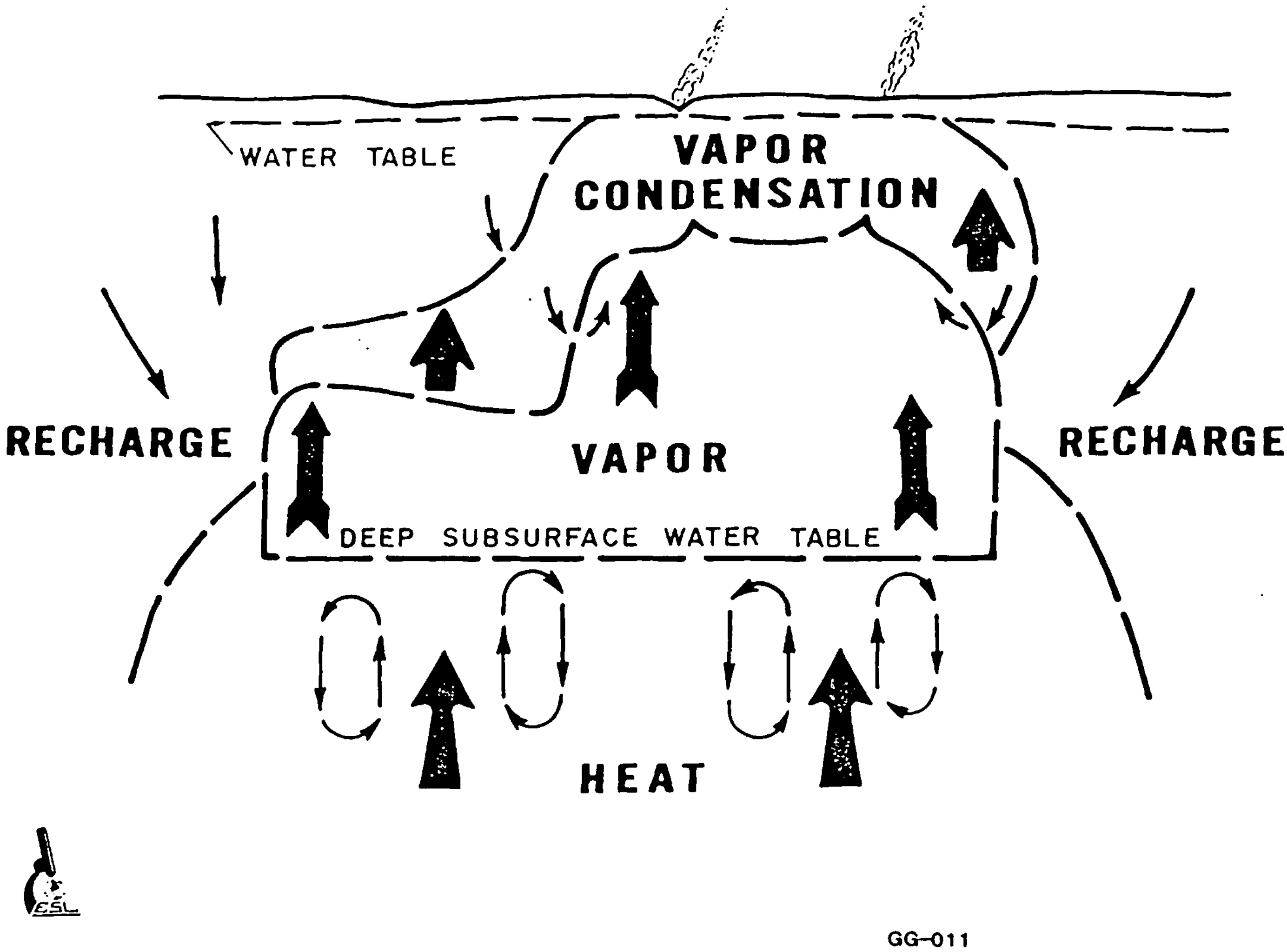


Figure 8

cooler 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. If an open fracture penetrates to the surface, steam may vent. Water lost to the system is 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 is controlled by the vapor phase and 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 steam 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 and above the reservoir is either naturally low or has been decreased by deposition in the fractures and pores of minerals from the hydrothermal fluid to form a sealed zone around the reservoir.

The Geysers geothermal area in California (see Figs. 20 and 21 and the discussion below) 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 about 1200 MWe (megawatts of electrical power, where 1 megawatt = 1 million watts). 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 vapor-dominated 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 by industry because they are generally easier and less expensive to develop than the more common water-dominated system discussed below.

Figure 9 schematically illustrates a high-temperature, hot-water-dominated hydrothermal system. Models for such systems have been discussed by White et al. (1971), Mahon et al. (1980), Henley and Ellis (1983), and Norton

HYDROTHERMAL SYSTEM IN VOLCANIC TERRANE

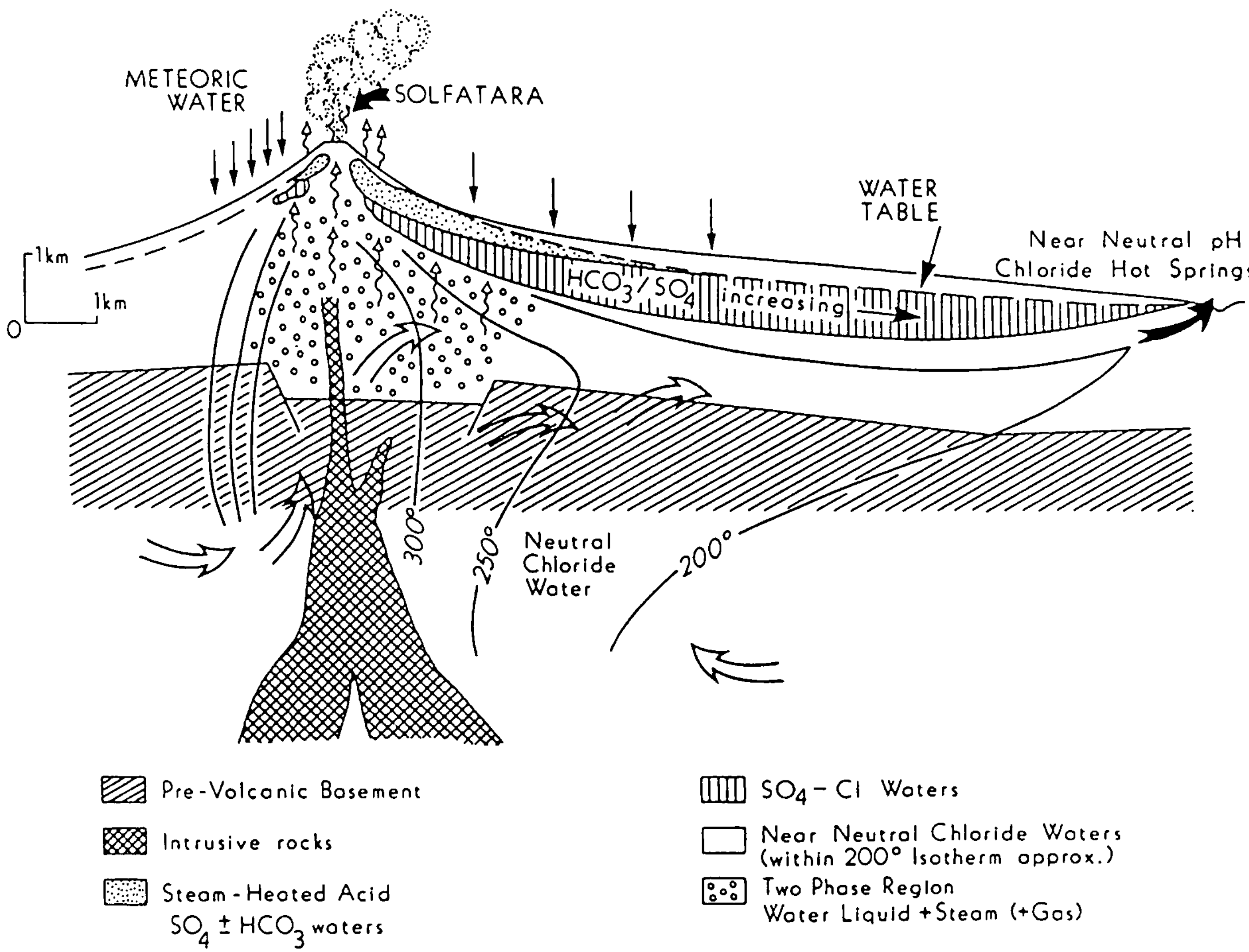


Figure 9

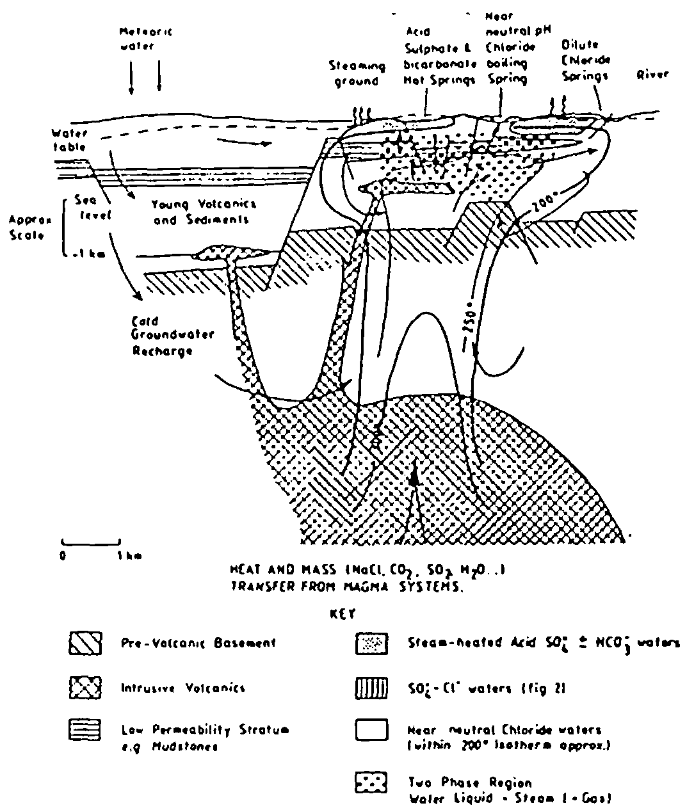
(1984), among others. The source of heat beneath many such systems is probably molten rock or rock that has solidified only in the last few tens of thousands of years, lying at a depth of perhaps 3 to 10 km. Normal ground water circulates in open fractures and removes heat from these deep, hot rocks by convection. Rapid convection produces uniform temperatures over large volumes of the reservoir. In some parts of the system, boiling may occur and a two-phase region (water and steam) may exist, but the pressure in the system is controlled by the water. Pressure therefore increases much more rapidly with depth than it does in a vapor-dominated system. 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 geothermal fluids of minerals in fractures and pore spaces. Surface manifestations of such geothermal systems 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 encounter tight, hot rocks with hot water inflow from the rock into the well bore mainly along open fractures. Areas where different fracture or fault 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.

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

The vertical pressure and temperature gradients in most high-temperature (i.e. > 200°C) hydrothermal convection systems lie near the curve of boiling point versus depth for saline water, and sporadic boiling occurs in many systems. Because boiling concentrates such acidic gases as CO_2 and H_2S in the steam, the oxygenated meteoric fluids overlying a boiling reservoir are heated and acidified. This process may lead to the deposition of clays and the formation of fluids having a distinct $NaHCO_3(-SO_4)$ chemical character.

The general structure of high-temperature systems associated with andesitic stratovolcanos (e.g., the Cascade Range, U.S.A.; Ahuachapan, El Salvador), silicic or bimodal volcanic regimes (e.g., Coso, California; Steamboat Hot Springs, Nevada; the Taupo volcanic zone, New Zealand) and sedimentary basins (e.g., the Imperial Valley, California, and Mexicali Valley, Mexico) are shown in Figures 9, 10, and 11, respectively. The mineral assemblages produced by the thermal fluids significantly alter the physical properties of the reservoir rocks. The six factors temperature, fluid composition, permeability, and to a lesser extent, pressure, rock type, and time each control the distribution and type of hydrothermal alteration (Browne, 1978). The alteration minerals are strongly zoned in most systems. Beneath the water table, clay minerals, quartz and carbonate are the dominant secondary minerals below temperatures of about 225°C. Chlorite, illite, epidote, quartz and potassium feldspar are important at higher temperatures. In the highest-temperature fields (above 250°C), metamorphism to the greenschist or higher facies may occur, resulting in significant densification of the reservoir rocks. Precipitation of silica may occur through cooling of the hot brine. The porosity and permeability of the silicified rocks are thereby considerably reduced, which can effectively seal the sodium chloride reservoir and prevent its expansion or appearance at the surface. However, steam and gas may be able to move through the sealed boundary and to interact with meteoric water above. The product of this interaction is usually a near-neutral pH sodium bicarbonate-sulfate water that forms a hot, secondary geothermal reservoir. Although the bicarbonate-sulfate waters may constitute an exploitable resource, it is the deep chloride water that is the prime hydrothermal resource.

Fumaroles may vent CO_2 and H_2S at the surface, which interact with meteoric water to produce highly acidic waters that cause advanced argillic alteration of near-surface rocks. Intense alteration of this type may extend



Meteoric water
Steaming ground
Acid Sulphate & bicarbonate Hot Springs
Near neutral pH Chloride boiling Spring
Dilute Chloride Springs
River
Water table
Sea level
Approx Scale
-1 km
Young Volcanics and Sediments
200°
250°
Cold Groundwater Recharge
0
1 km
HEAT AND MASS (NaCl, CO_2, SO_2, H_2O ..)
TRANSFER FROM MAGMA SYSTEMS.
KEY
Pre-Volcanic Basement
Intrusive Volcanics
Low Permeability Stratum e.g Mudstones
Steam-heated Acid $SO_4^=$ ± HCO_3^- waters
$SO_4^=$ - Cl^- waters (fig 2)
Near neutral Chloride waters (within 200° Isotherm approx.)
Two Phase Region Water Liquid + Steam (+ Gas)

Figure 10

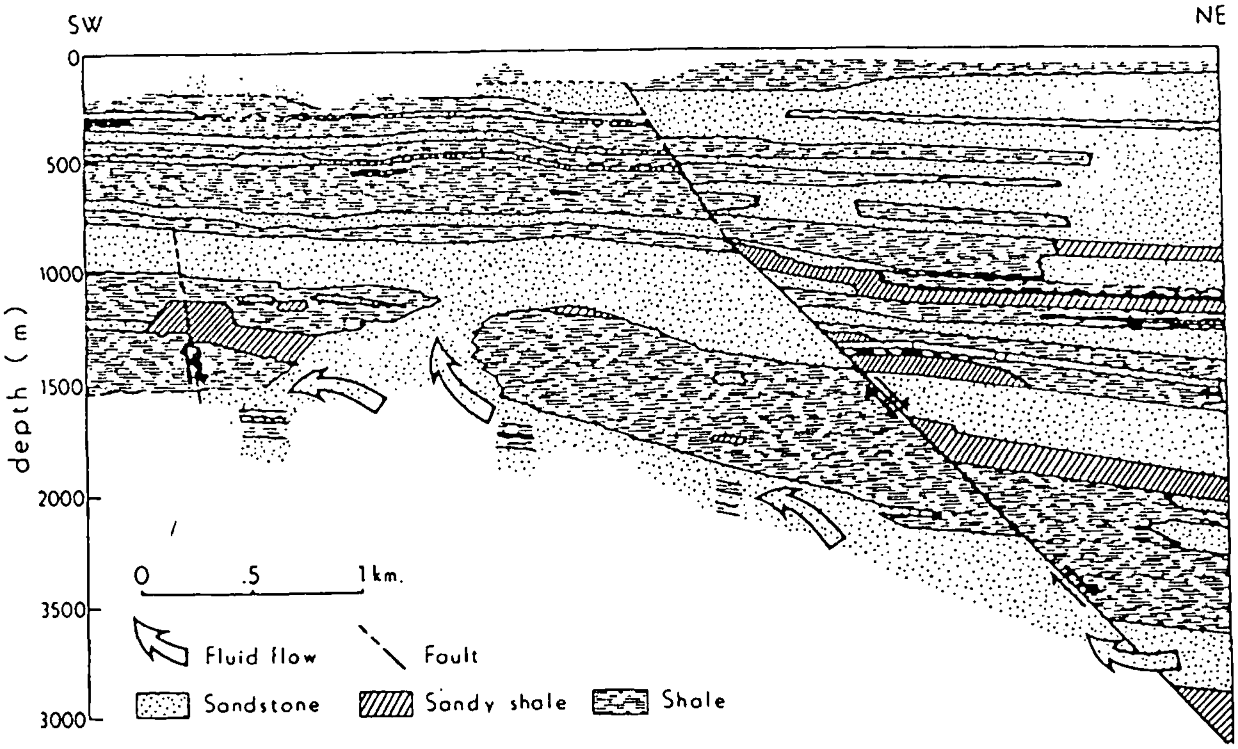


Figure 11. Fluid Flow Model of Cerro Prieto, Mexico.

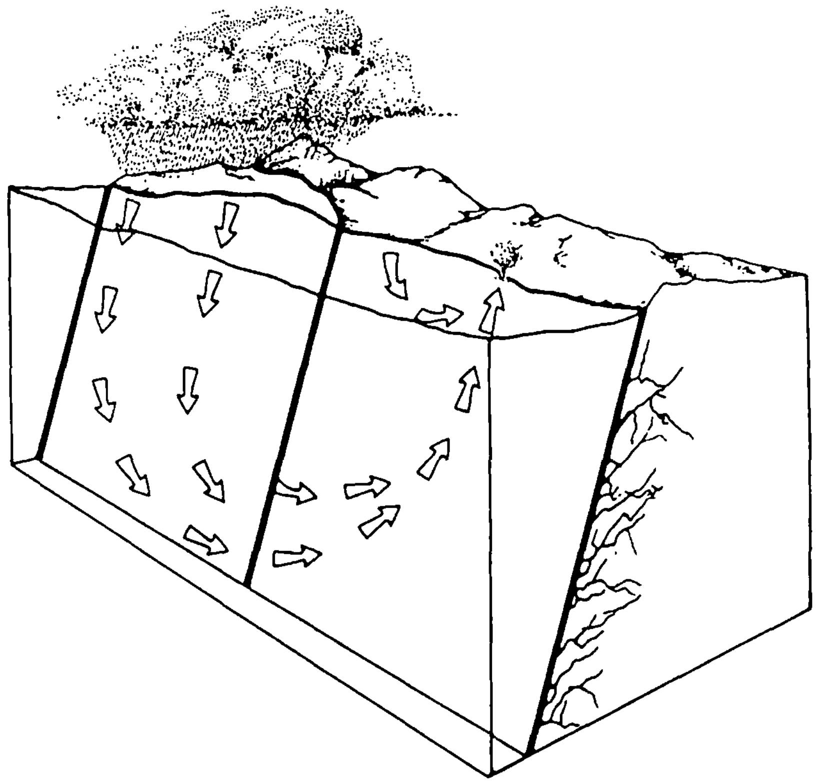
to depths of hundreds of meters below the surface in areas such as Cove Fort-Sulphurdale, Utah, where the water table is deep (Ross and Moore, 1985).

Outflow of the deep NaCl fluid may occur at a considerable distance from the hottest portion of a hydrothermal system. These chloride brines may emerge as boiling springs, frequently surrounded by silica deposits, or as a non-boiling mixture of local groundwater, with geothermal bicarbonate and chloride fluids. Because the solubility of calcite decreases with increasing temperature and decreases when CO_2 is released to the atmosphere, $CaCO_3$ in the form of travertine often precipitates where thermal waters mix with groundwaters and/or reach the surface.

Virtually all of industry's geothermal exploration effort in the United States is presently directed at locating vapor- or water-dominated hydrothermal systems of the types described above having temperatures above 200°C. A few of these resources are capable of commercial electrical power generation today. Current surface exploration techniques are generally conceded to be inadequate for discovery and assessment of these resources at a fast enough pace to satisfy the reliance the U.S. may ultimately put upon them for alternative energy sources. Development of better and more cost-effective techniques is badly needed.

The fringe areas of high-temperature vapor- and water-dominated hydrothermal systems often produce water of low and intermediate temperature. These lower-temperature fluids are suitable for direct-heat applications and may also be used for electrical power production with the newer binary technology typified by Ormat systems, for example. 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. 12). 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. This type of warm water resource is especially prevalent in the western U.S. where active faulting keeps conduits open to depth.

<u>Sedimentary Basins/Regional Aquifers</u>. Some basins are filled to depths



MODEL OF DEEP CIRCULATION HYDROTHERMAL RESOURCE

Figure 12

GG042a

of 10 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. The Madison group carbonate rock sequence of widespread occurrence in North and South Dakota, Wyoming, Montana, and northward into Canada contains warm waters that are currently being tapped by drill holes for space heating and agricultural purposes. In a similar application, 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 (Varet, 1982). Many other areas of occurrence of this resource type are known worldwide.

Geopressured Resources. Geopressured resources consist of deeply buried fluids 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 13 gives a few typical parameters for geopressured reservoirs and illustrates the origin of the above-normal fluid pressure. These geopressured fluids, found mainly in the Gulf Coast of the U.S. (Fig. 16), generally 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.

Industry has a great deal of interest in development of geopressured resources, although they are not yet economic. The U.S. Department of Energy, Geothermal Technology Division, is currently sponsoring development of exploitation technologies. This program is being managed by the Idaho National Engineering Laboratory of DOE in Idaho Falls, Idaho.

Radiogenic Resources. Research that could lead to development of radiogenic geothermal resources in the eastern U.S. has been done following ideas developed at Virginia Polytechnic Institute and State University (Costain et al., 1980). The eastern states coastal plain is blanketed by a layer of thermally insulating sediments. In places beneath these sediments, rocks are

GEOPRESSURED GEOTHERMAL RESOURCE

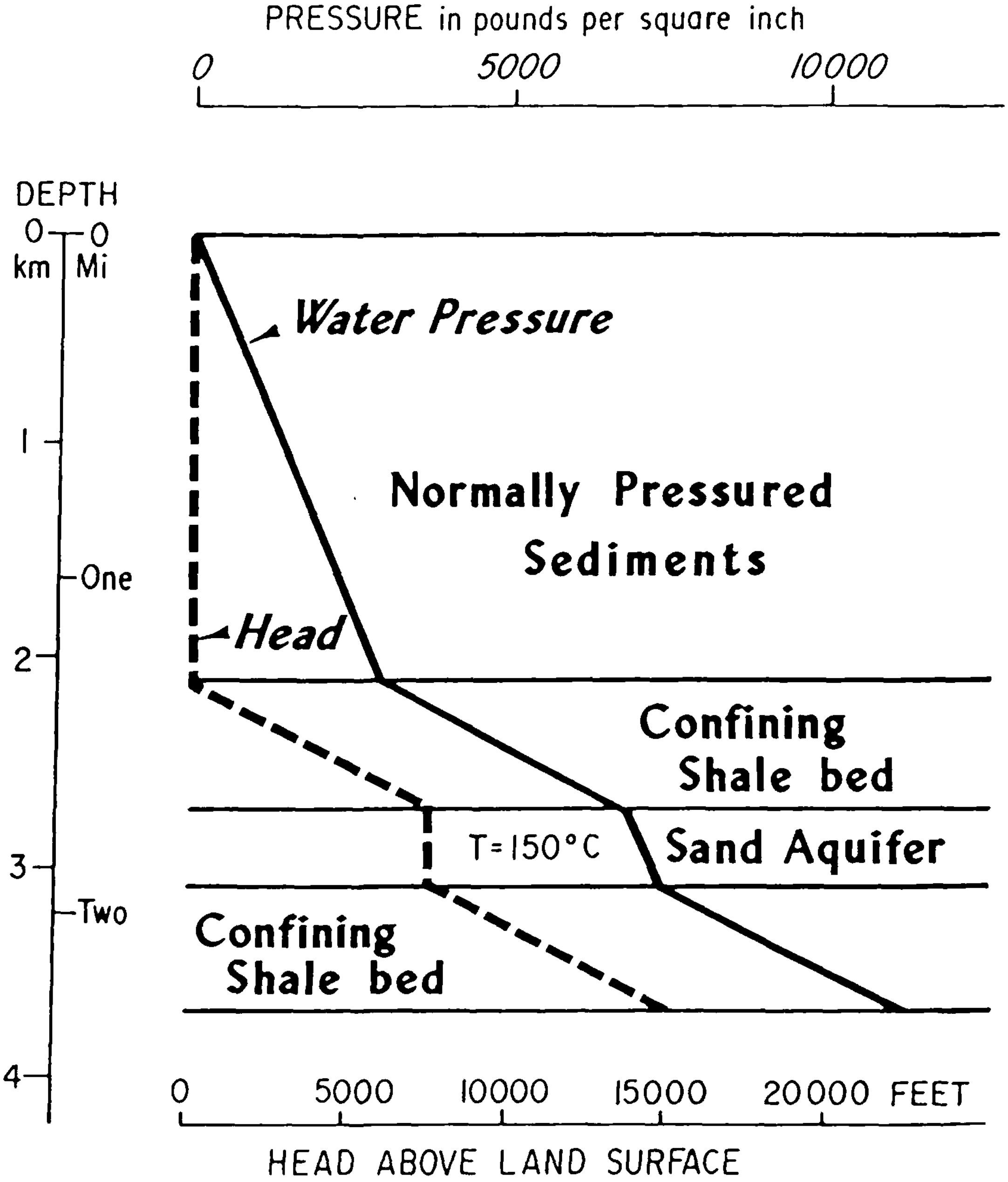


Figure 13

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RADIOGENIC GEOTHERMAL RESOURCE

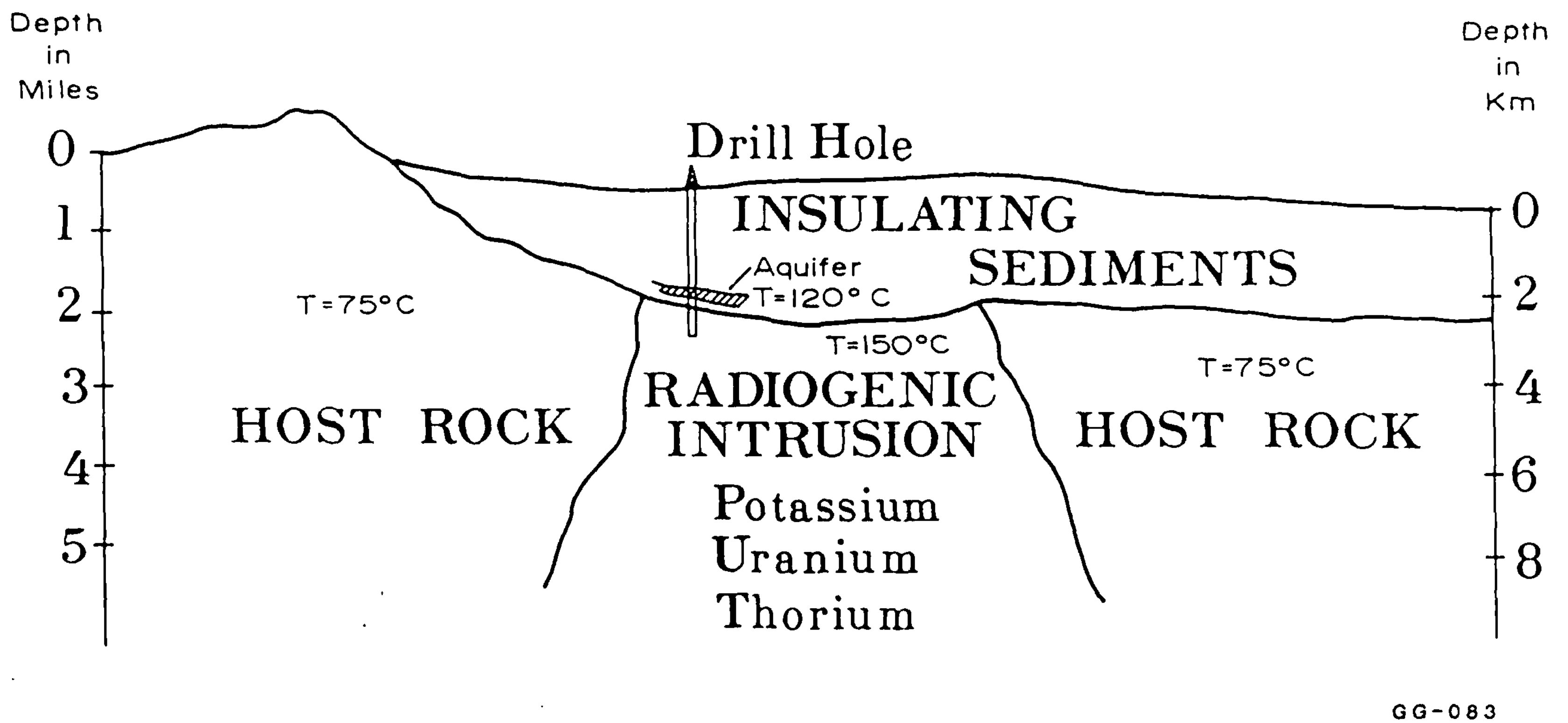
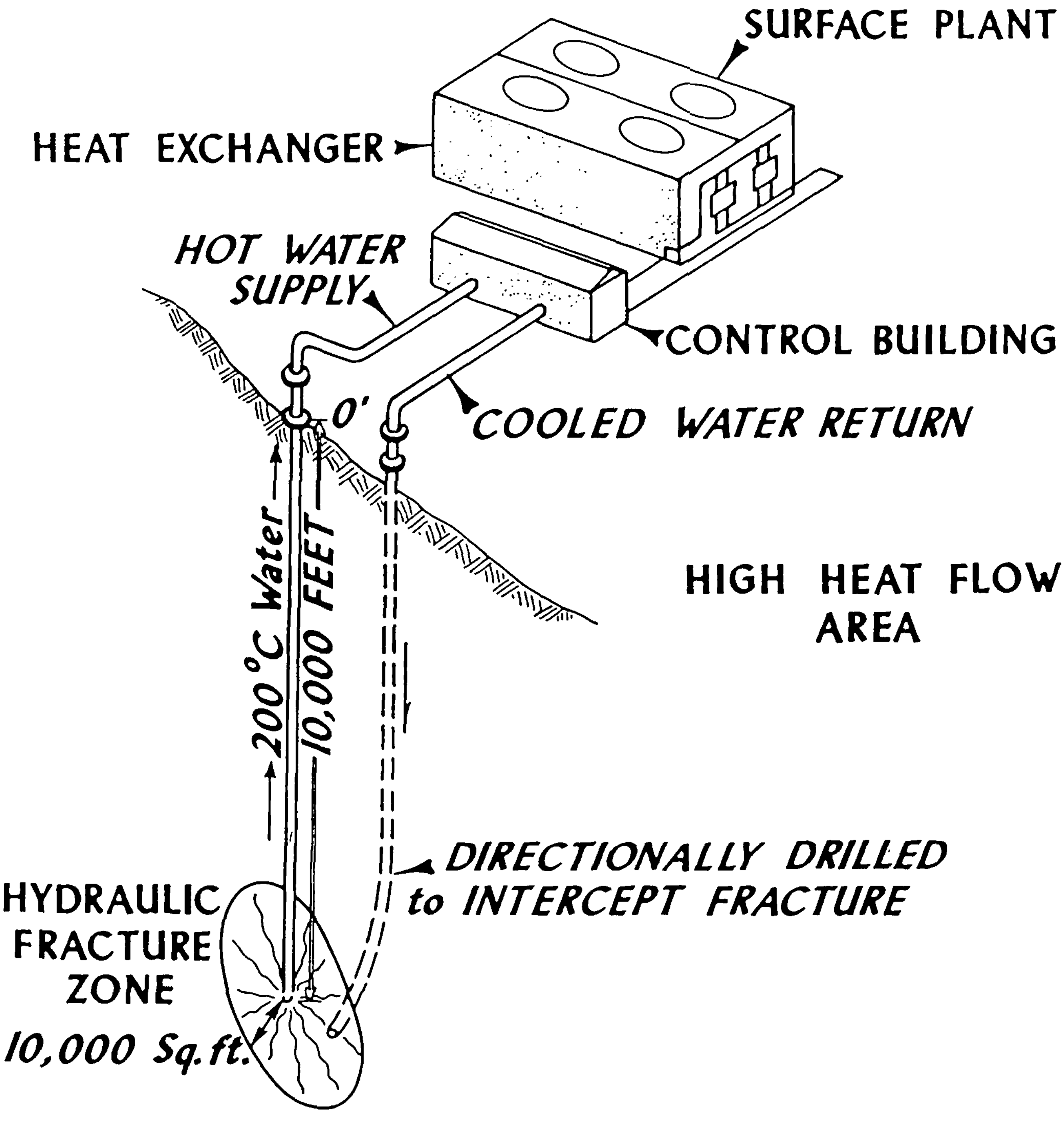


Figure 14



HOT DRY ROCK GEOTHERMAL RESOURCE

Figure 15

GG-081

believed to occur that have an anomalously high heat production due to high natural content of radioactive elements. These rocks represent old intrusions of once molten material that have long since cooled and crystallized. Geophysical and geological methods for locating such radiogenic rocks beneath the sedimentary cover have been partly developed, and very limited drill testing of the geothermal target concept (Fig. 14) has been completed under DOE funding. These resources may ultimately yield low- to intermediate-temperature geothermal water suitable for space heating and industrial processing. Availability of such a resource could mean a great deal to the eastern U.S. where energy consumption is high and where no shallow, high-temperature hydrothermal convection systems are known or expected to occur. Geophysical and geological data indicate that radiogenically heated rock bodies may be reasonably widespread.

Hot Dry Rock Resources. Hot dry rock resources 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 few pore spaces or fractures, and therefore contain little water and no interconnected permeability. The feasibility and economics of extraction of heat for electrical power generation and direct uses from hot dry rocks is presently the subject of a $150 million research program at the U.S. Department of Energy's Los Alamos National Laboratory in New Mexico (Smith and Ponder, 1982). Batchelor (1982) describes similar successful research conducted in England. Both projects indicate 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. During formation of the fracture system, its orientation and extent are mapped using passive seismic geophysical techniques. A second borehole is located such that it intersects the fracture system. Water can then be circulated down one hole, through the fracture system where it is heated, and up the second hole (Fig. 15). Fluids at temperatures of 150°C to more than 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. Experiments are underway at the U. S. Department of Energy's Sandia National Laboratories in Albuquerque, New Mexico to learn how to extract heat energy directly from molten rock. Techniques for locating a shallow, crustal magma body, drilling into it and implanting heat exchangers or possibly direct electrical converters remain to be developed (Carson and Allen, 1984).

Neither these experiments nor those of the hot, dry rock type described above are expected to result in economic energy production in the near future. In Iceland, however, where geothermal energy was first tapped for space heating in 1928, economic technology has been demonstrated for extraction of thermal energy from young lava flows (Björnsson, 1980). A heat exchanger constructed on the surface of the 1973 lava flow on Heimaey in the Westman Island group, recovers steam which results from downward percolation of water applied at the surface above hot portions of the flow. The space heating system which uses this energy has been operating successfully for over eight years.

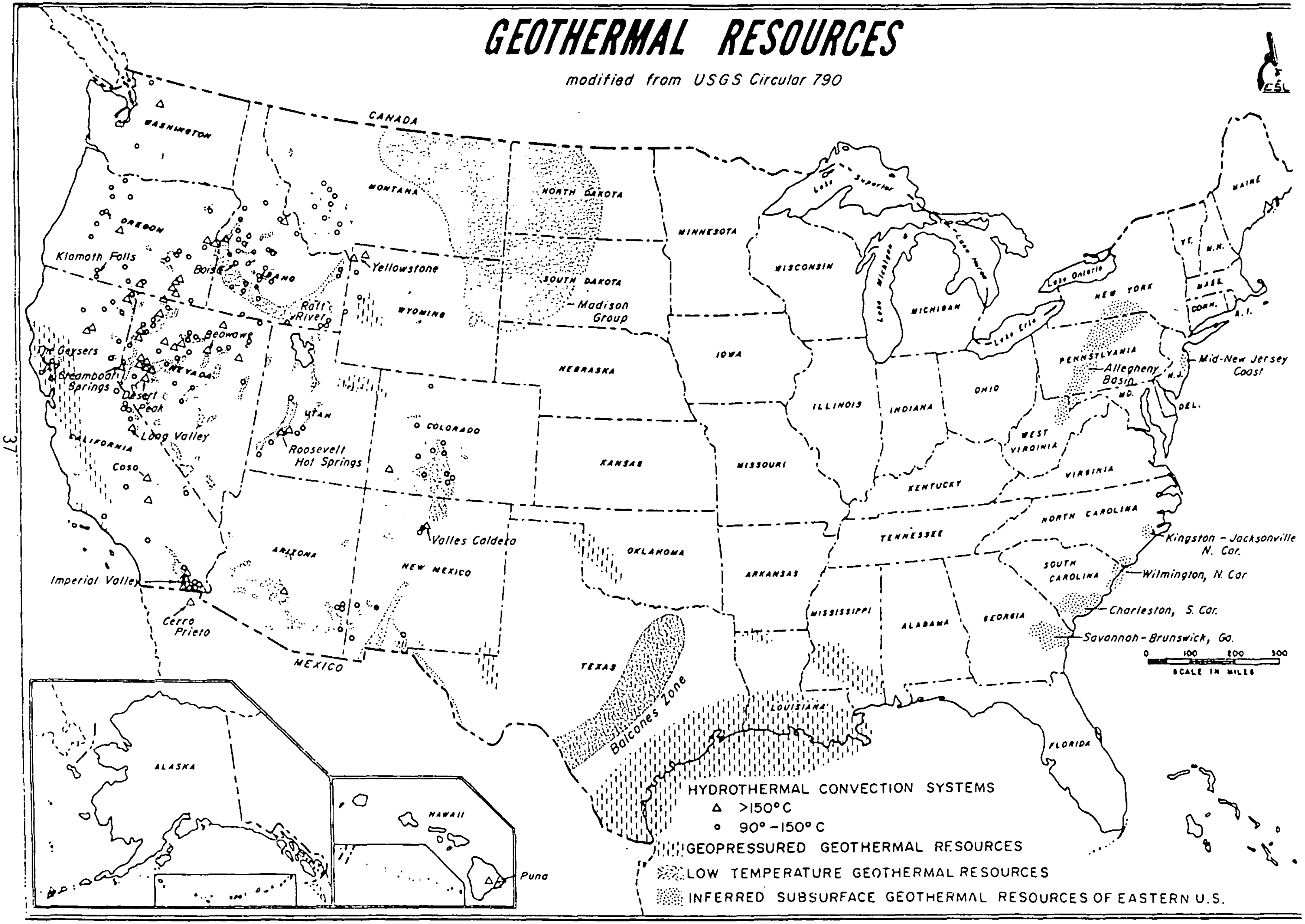
GEOTHERMAL RESOURCES IN THE CONTINENTAL UNITED STATES

Figure 16 displays the distribution of various resource types in the 48 contiguous states. Information for this figure was taken mainly from Muffler et al. (1978) and Reed (1982), where more detailed discussions and more detailed maps can be found. Not shown are locations of hot dry rock or magma resources because very little is known. In addition, it should be emphasized that the present state of knowledge of geothermal resources of all types is poor. Because of the very recent emergence of the geothermal industry, insufficient exploration has been done to define properly the resource base. Each year brings more resource discovery, so that Figure 16 will rapidly become outdated.

Figure 16 shows that most of the known hydrothermal resources and all of the presently known sites that are capable or believed to be capable of electric power generation from hydrothermal convection systems are in the western half of the U. S. The preponderance of thermal springs and other surface manifestations of underlying resources is also in the west. Large areas underlain by warm waters in sedimentary rocks exist in Montana, North and South Dakota, and Wyoming (the Madison Group of aquifers), but the extent and potential of these resources is poorly understood. Another important large area, much of which is underlain by low-temperature resources, is the northeast-trending Balcones zone in Texas. The geopressured resource areas of the Gulf Coast and surrounding states are also shown. Resource areas indicated in the eastern states are highly speculative because only one site has been drill tested to actually confirm their existence, which is only inferred at present.

Regarding the temperature distribution of geothermal resources, low- and intermediate-temperature resources are much more plentiful than are high-temperature resources. There are many, many thermal springs and wells that have water at temperature only slightly above the mean annual air temperature, which is the temperature of most non-geothermal shallow ground water. Resources having temperatures above 150°C are infrequent, but represent important occurrence. Muffler et al. (1978) show a statistical analysis of the temperature distribution of hydrothermal reosurces and conclude that the cumulative frequency of occurrence increases exponentially as reservoir temperature decreases (Fig. 17). This relationship is based only on data for known

Figure 16



Geo/RM-00

FREQUENCY OF OCCURRENCE VS TEMPERATURE FOR GEOTHERMAL RESOURCES

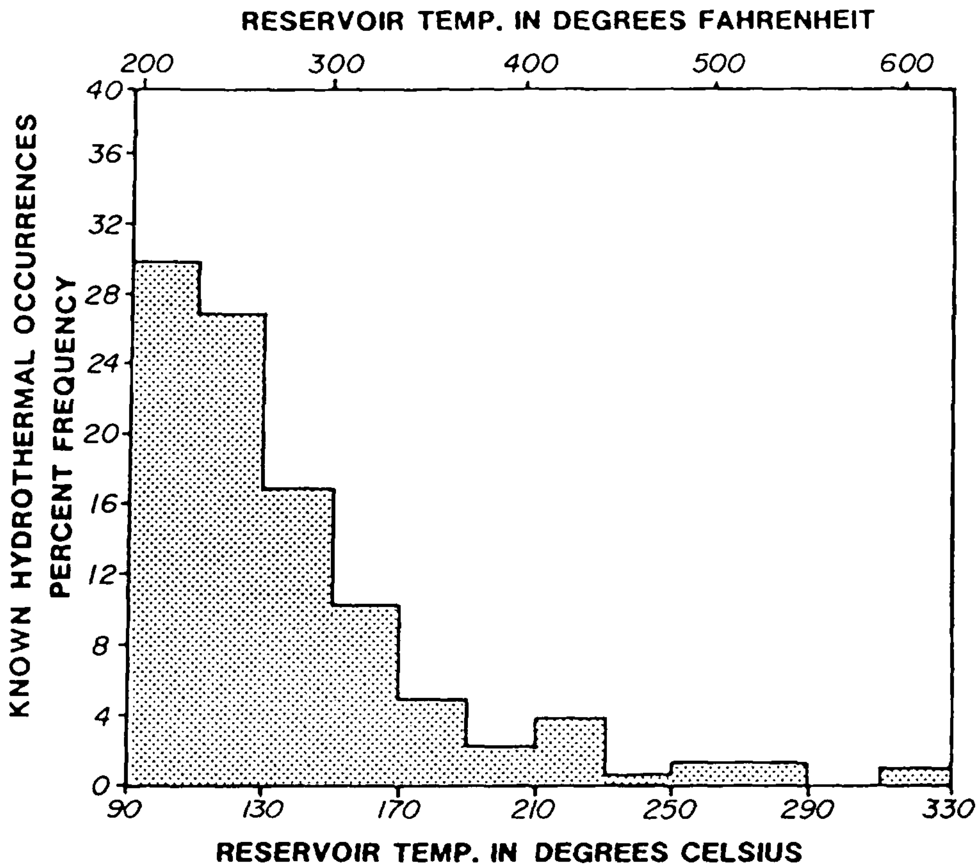


Figure 17

GG-007

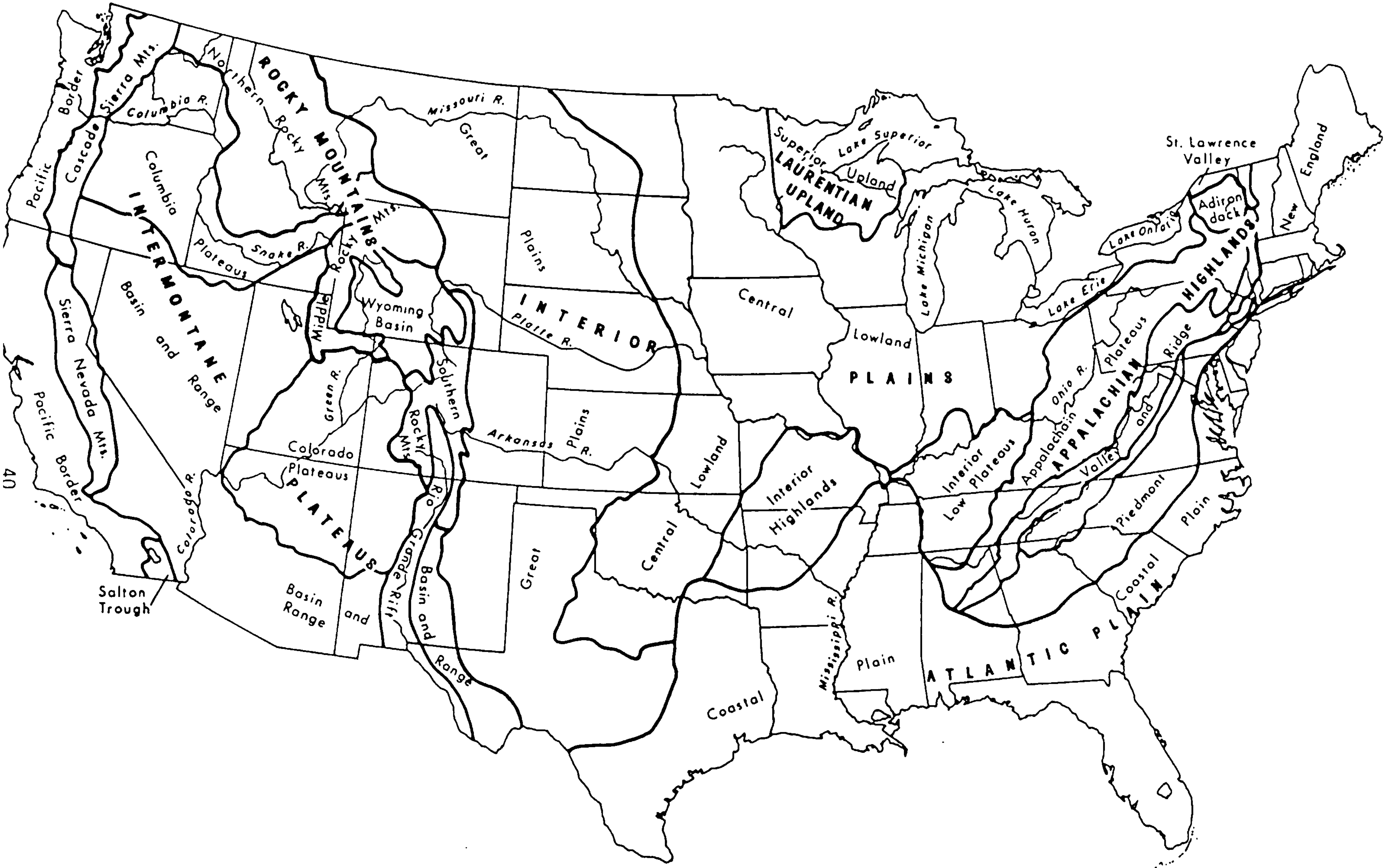
occurrences having temperatures 90°C or higher. It is firmly enough established, however, that we can have confidence in the existence of a very large low-temperature resource base, most of which is undiscovered.

Let us consider the known geothermal occurrences in a bit more detail, beginning in the Western U. S. The reader should refer to Figure 18 for locations of some of the geologic provinces discussed.

Salton Trough/Imperial Valley, CA

The Salton Trough is the name given an area along the landward extension of the Gulf of California. It is composed of the Imperial Valley in the U.S. and the Mexicali Valley in Mexico. The Salton Sea Trough is an area of complex, currently active plate tectonic geologic processes. As shown on Figure 2, the crest of the East Pacific Rise spreading center is offset repeatedly northward up the Gulf of California by transform faulting. Both the rise crest and the transform faults come onto the continent under the delta of the Colorado River (Fig. 19) and the structure of the Salton Trough suggests that they underlie the trough. The offsetting faults show right-lateral movement and trend northwestward, parallel to the strike of the well-known San Andreas fault. Elders (1979) and the contributing authors for his guidebook give summaries of the geothermal systems that occur in the Salton Trough.

The Salton Trough has been an area of subsidence since Miocene times. During the ensuing years sedimentation in the trough has kept pace with subsidence, with shallow water sediments and debris from the Colorado River predominating. At present, 3 to 5 km of poorly-consolidated sediment overlie a basement of Mesozoic crystalline rocks that intruded Paleozoic and Precambrian sedimentary rocks. Detailed analysis of drilling data and of surface and downhole geophysics indicates that at least some of the known geothermal occurrences (Cerro Prieto, Brawley and the Salton Sea) are underlain by "pull-apart basins" apparently caused by crustal spreading above a local section of the East Pacific Rise crest (Elders, 1979). Very young volcanic activity has occurred at Cerro Prieto where a rhyodacite volcanic cone is known, and along the southern margin of the Salton Sea where rhyolite domes occur. The Salton Sea domes have an approximate age of 60,000 years (Muffler and White, 1969). The Cerro Prieto volcano has been difficult to date but may be about 10,000 years old (Wollenberg et al., 1980). Faulting is occurring at the present



PHYSIOGRAPHIC MAP of USA

(after Fenneman, 1928)

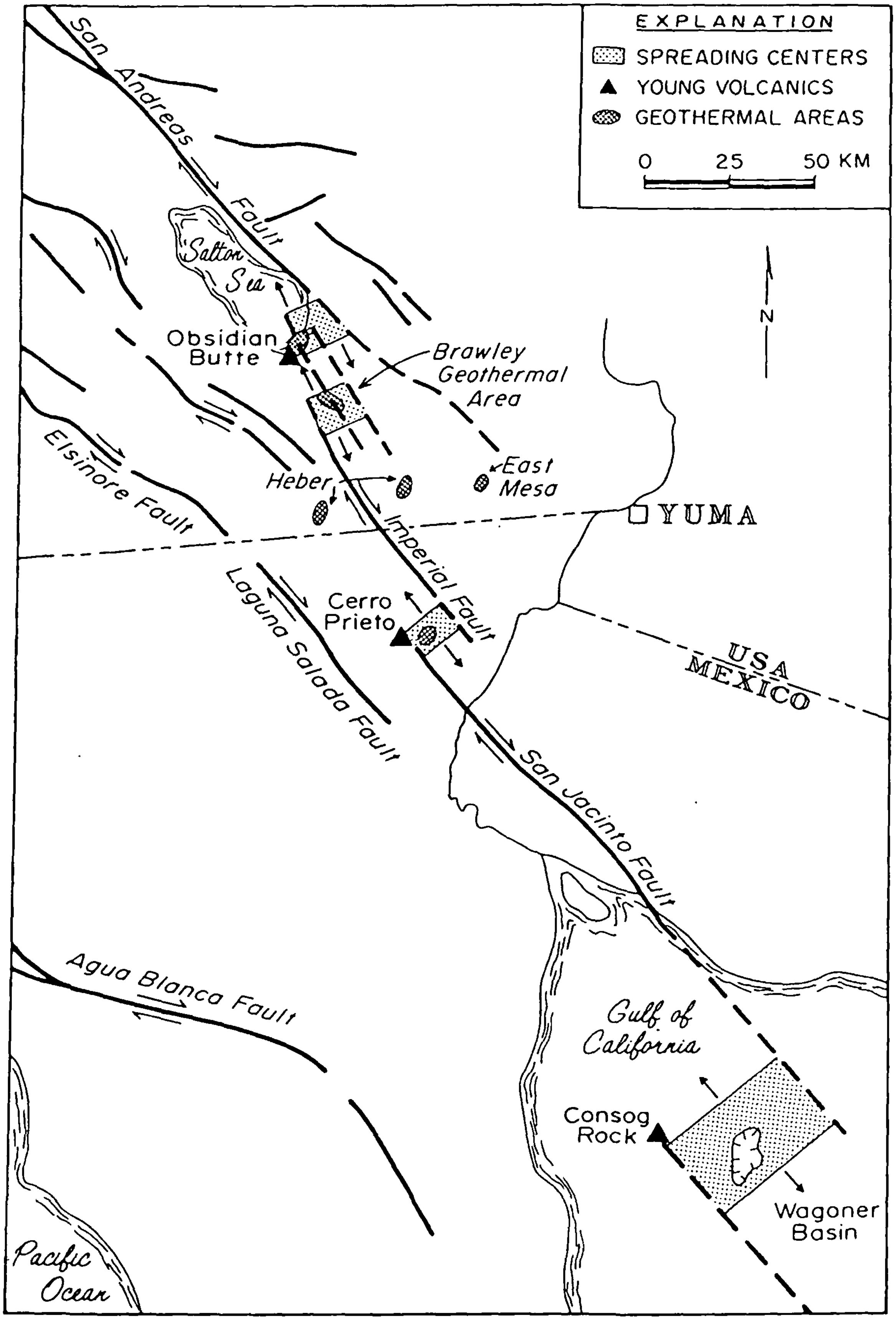
Figure 18

SCALE

0 100 200 300 400 MILES

0 200 400 600 KILOMETERS

USA-002



MAJOR STRUCTURES OF SALTON TROUGH

(after Palmer et al., 1975)

Figure 19

GG-086

time as evidenced by the many earthquakes and earthquake swarms recorded in the Salton Trough.

The Cerro Prieto field is the best understood geothermal occurrence in the Salton Trough because of the drilling done there and its history of production. We may take it has an example of a Salton Trough resource type (refer to Fig. 11). The field is water-dominated and the more than 60 wells produce from depths of 1.5 to over 3 km. Fluid temperatures range from about 200°C to over 350°C (Alanso et al., 1979). The rocks are composed of an upper layer of unconsolidated silts, sands and clays, and a layer of consolidated sandstones and shales overlying the crystalline basement (Puete Cruz and de la Pena, 1979). Two principal reservoir horizons occur in sandstones within the consolidated sequence and enhanced production has been noted in the vicinity of faults, indicating that fracture permeability is important, although intergranular permeability due to dissolution of minerals by the geothermal fluids is believed to be important also (Lyons and Van de Kamp, 1980). Reservoir recharge is apparently from the northeast and east and consists, at least partly, of Colorado River water (Truesdell et al., 1980).

The geothermal fluid from Cerro Prieto, after steam separation, contains about 25,000 ppm total dissolved solids. This figure is much lower than some of the other resources in the Salton Trough. For example, the Salton Sea area contains 20 to 30 percent by weight by solids (Palmer, 1975).

The heat source(s) for the several Salton Trough resources have not been found by drilling, although basalt dikes have been intersected in several areas, leading credence to the idea of upwelling basalt in pull-apart zones. Presumably, intrusion brings magma up into the depth range 5-10 km beneath the thermal anomalies.

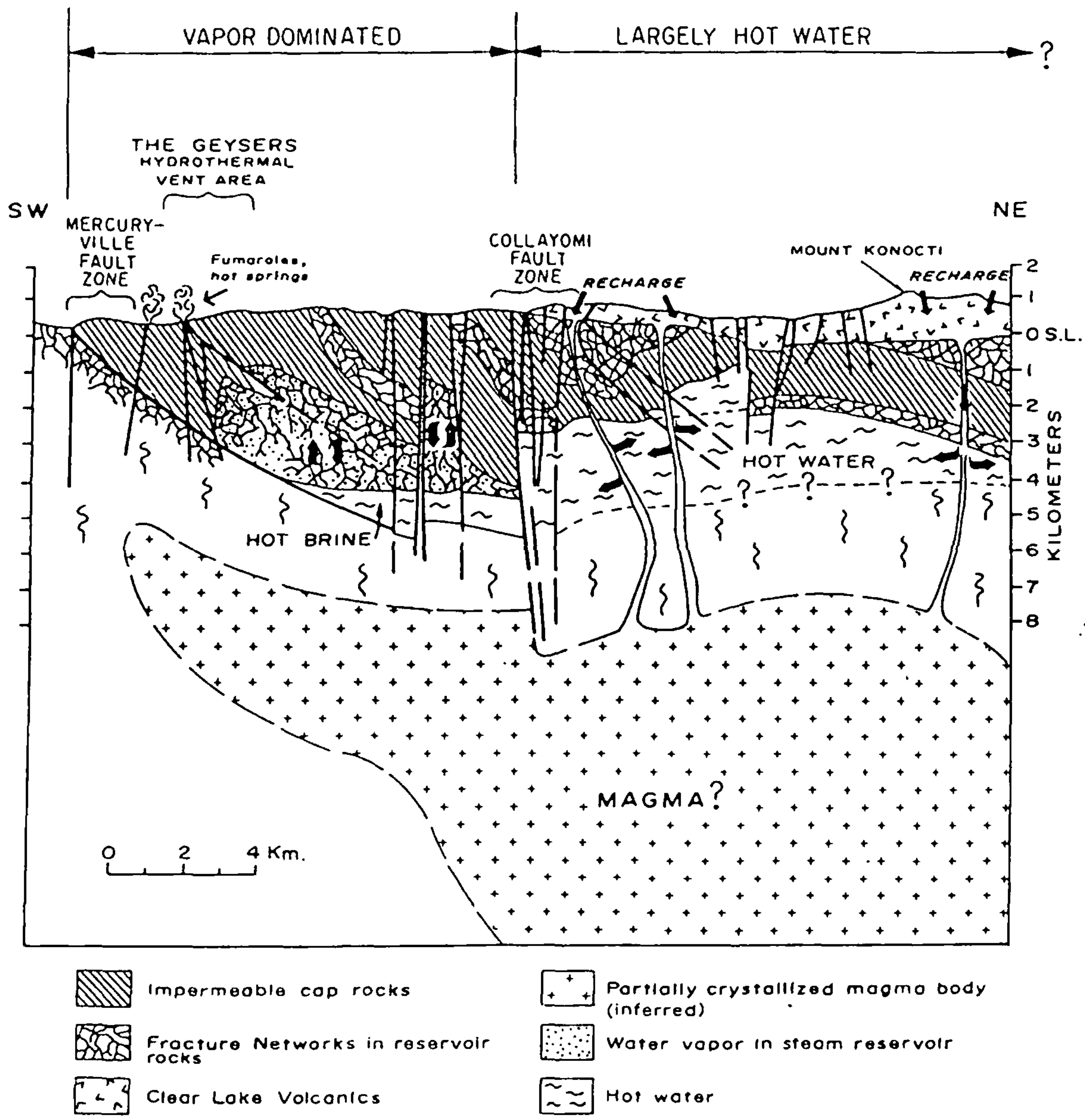
The Geysers, CA

The Geysers geothermal area is the world's largest producer of electricity from geothermal fluids with more than 1200 MWe on line and an additional several hundred scheduled. This area lies about 150 km north of San Francisco. The portion of the resource being exploited is a vapor-dominated field having a temperature of 240°C. The ultimate potential of the vapor-dominated system is presently believed to be around 2000 MWe. Associated with the vapor-dominated field are believed to be several unexploited hot water-dominated reservoirs whose volumes and temperatures are unknown (Fig. 20).

The geology of The Geysers area is complex, especially structurally. Reservoir rocks consist mainly of fractured greywackes, sandstone-like rocks consisting of poorly sorted fragments of quartzite, shale, granite, volcanic rocks and other rocks. The fracturing has created the permeability necessary for steam production in quantities large enough to be economically exploitable. Overlying the reservoir rocks, as shown in Figure 20, is a series of impermeable metamorphosed rocks (serpentinite, geenstone, melange and metagranite) that form a cap on the system. These rocks are all complexly folded and faulted. They are believed to have been closely associated with and perhaps included in subduction of the eastward-moving Pacific plate (Fig. 2) under the continent. This subduction apparently ended 2 to 3 million years ago.

As shown in Figure 21, the presently known steam field is confined between the Mercuryville fault zone on the southwest and the Collayomi fault zone on the northeast. The northwest and southeast margins are not definitely known. To the east and northeast lies the extensive Clear Lake volcanic field composed of dacite, rhyolite, andesite and basalt. The interval of eruption for these volcanics extends from 2 million years ago to 10,000 years ago, with ages progressively younger northward (Donnelly, 1977). The Clear Lake volcanics are very porous and soak up large quantities of surface water. It is believed that recharge of a deep, briny hot-water reservoir comes from water percolating through the Clear Lake volcanics, and that this deep reservoir may supply steam to the vapor-dominated system through boiling (Fig. 20) although these ideas are not universally supported by geologists and the deep water table has never been intersected by drilling.

The postulated water-dominated geothermal reservoirs do not occur everywhere in the Clear Lake volcanics. At several locations drill holes have found temperatures of 200°C at depths of only 2000 m, but the rocks are tight and impermeable (Goff, 1980). Fractured areas apparently host the water-dominated reservoirs at the Wilbur Springs district (Thompson, 1979), the Sulphur Bank Mine (White and Roberson, 1962) and other smaller occurrences. Potential in The Geysers area for discovery of additional exploitable resources is good.

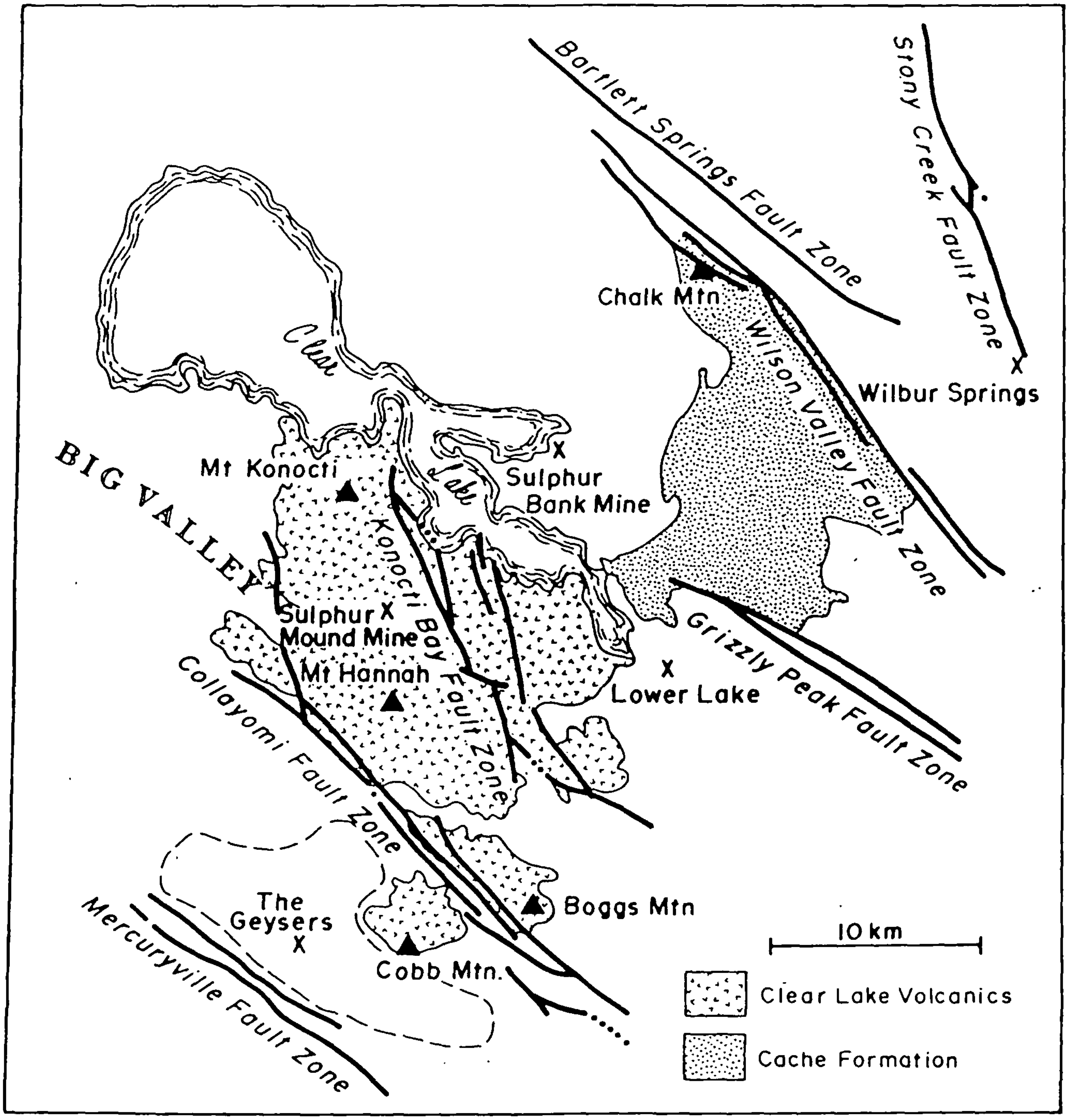


CRUSTAL MODEL FOR THE GEYSERS - CLEAR LAKE AREA, CA.

(after McLaughlin, 1977)

Figure 20

Ca/Ge-001



MAJOR STRUCTURES in THE GEYSERS-CLEAR LAKE AREA

(After Goff, 1980)

Ca/Ge-002

Figure 21

Basin and Range

The Basin and Range province extends from Mexico into southern Arizona, southwestern New Mexico and Texas on the south, through parts of California, Nevada and Utah, and becomes ill-defined beneath the covering volcanic flows of the Columbia Plateau on the north (Fig. 18). This area, especially the northern portion, contains abundant geothermal resources of all temperatures and has been one of the several areas of active exploration in the U. S. Resources along the eastern and western margins of the province appear to be both more abundant and of higher temperature. Electrical power is presently being generated from Roosevelt Hot Springs (20 MWe) and Cove Fort/Sulphurdale (3.2 MWe) in Utah. Candidate sites include Steamboat Springs, Dixie Valley, Desert Peak and Beowawe in Nevada and Coso, California. At both Desert Peak and Beowawe, plant construction is underway. Exploration is being or has been conducted at probably 20 or more sites in the Basin and Range, including, in addition to those named above, Tuscarora, McCoy, Baltazor, Leach Hot Springs, San Emidio, Soda Lake, Stillwater, and Humboldt House, Nevada; and Surprise Valley, and Long Valley Caldera, California. Direct application of geothermal energy for industrial process heating and space heating are currently operating n this area at several sites including Brady Hot Springs (vegetable drying), Reno (space heating) and Salt Lake City (greenhouse heating).

The reasons for the abundance of resources in the Basin and Range seem clear. This area, especially at its margins, is an active area geologically. Volcanism only a few hundred years old is known froms tens of areas, including parts of west central Utah on the east (Nash and Smith, 1977) and Long Valley caldera on the west (Rinehart and Huber, 1965). The area is also active seismically and faulting that causes the uplift of mountain ranges also serves to keep pathways open for deep fluid circulation at numerous locations. Rocks in the Basin and Range consist of Paleozoic and Mesozoic sandstones, limestones and shales that lie on Precambrian metamorphic and intrusive rocks. These rocks were deformed, complexly in some places, during the Nevadan and Laramide orogenies, and some base and precious metal deposits were formed. Beginning in mid-Tertiary times volcanic activity increased many fold with both basaltic and rhyolitic rocks being erupted. Extentional stresses also began to operate and a sequence of north-south mountain ranges were formed which separate valleys that have been filled with erosional debris from the mountains (Eardley, 1951). In some places more than 2 km offset has occurred along range-front faults, and the valleys may contain a hundred to as much as 3,000 m of unconsolidated erosional debris. This activity persists to the present time.

As an example of a Basin and Range hydrothermal system we will discuss Roosevelt Hot Springs, although it should not be supposed to be typical of all high-temperature occurrences in this province. This geothermal area has been studied in detail for the past six years (Nielson et al., 1978; Ward et al., 1978). The oldest rocks exposed (Figs. 22 and 23) are Precambrian sedimentary rocks that have been extensively metamorphosed. These rocks were intruded during Miocene time by granitic rocks (diorite, quartz monzonite, syenite and granite). Rhyolite volcanic flows and domes were emplaced during the interval 800,000 to 500,000 years ago. The area has been complexly faulted by north- to northwest-trending high-angle faults and by east-west high-angle faults. The Negro Mag fault is such an east-west fault that is an important controlling structure in the north portion of the field. The north-trending Opal Mound fault apparently forms the western limit of the system. The oldest fault system is a series of low-angle denudation faults (Fig. 23) along which the upper plate has moved west by about 600 m and has broken into a series of discrete blocks. Producing areas in the southern portion of the field are located in zones of intersection of the upper plate fault zone with the Opal Mound and other parallel faults. Producing zones in the northern part of the region are located at the intersection of north-south and east-west faults. The permeability is obviously fracture controlled.

Seven producing wells are shown on Figure 22 and more have been drilled recently. Fluid temperature is up to 260°C and the geothermal system is water-dominated. Average well production is perhaps 318,000 kg/hr (700,000 lbs/hr). Plans call for building up a base of experience with the 20 MWe power plant currently being operated there by Phillips Geothermal and Utah Power and Light, with two 50 MWe plants to be installed as knowledge of reservoir performance increases.

Cascade Range and Vicinity

The Cascade Range of northern California, Oregon, Washington and British Columbia is comprised of a series of volcanos, 12 of which have been active in

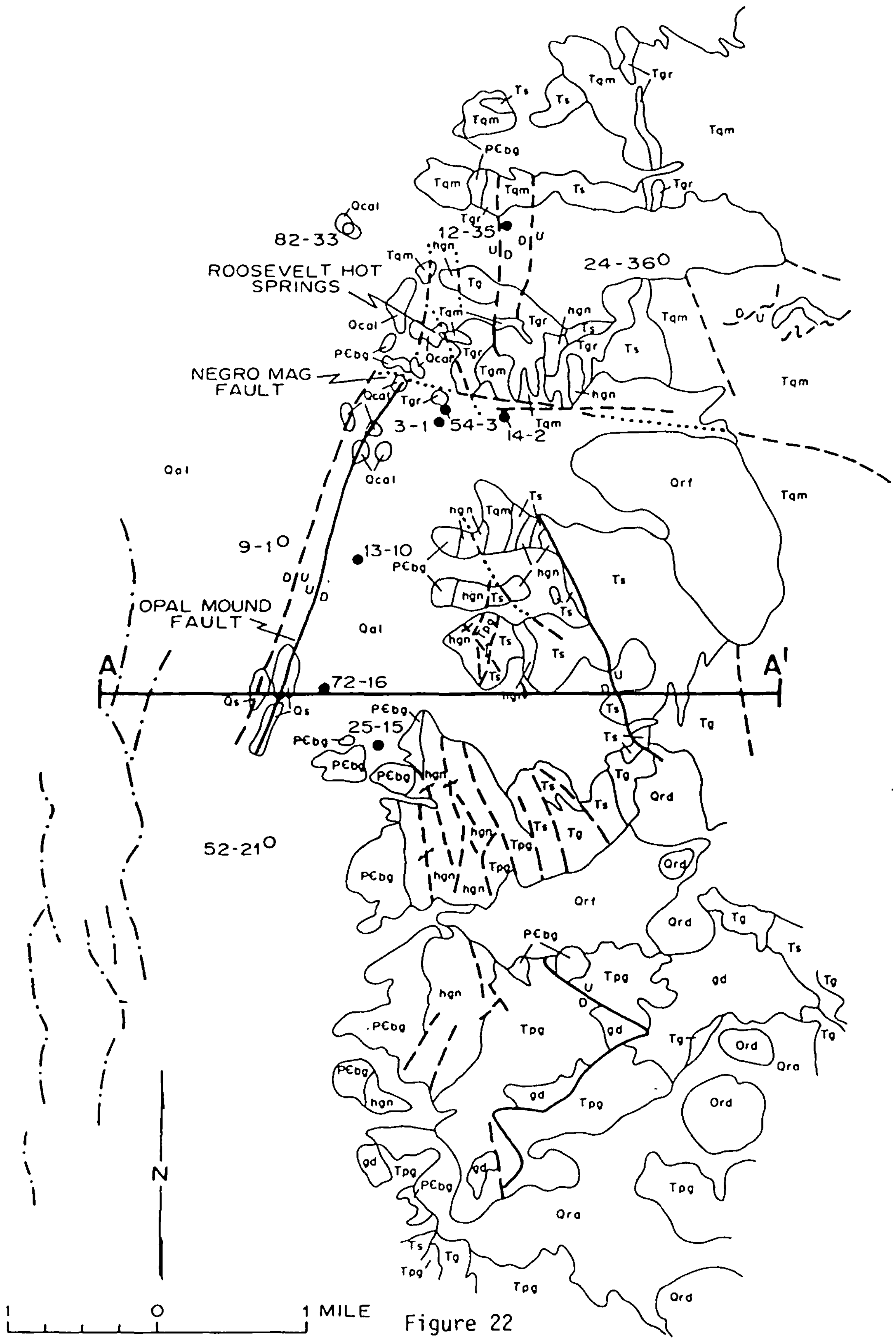


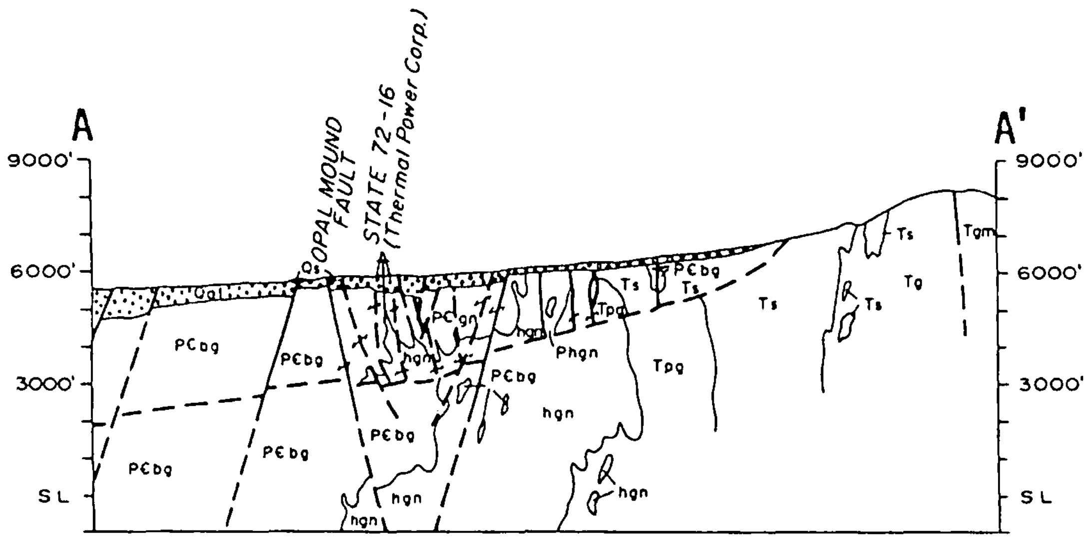
Figure 22

GEOLOGIC MAP
ROOSEVELT HOT SPRINGS, UTAH

(from Nielson et al., 1978)

GEOLOGIC CROSS SECTION
ROOSEVELT HOT SPRINGS, UTAH

(from Nielson et al., 1978)



EXPLANATION

Qal	- alluvium	Tg	- granite
Qcal	- silicified alluvium	Ts	- syenite
Qs	- siliceous sinter	Tpg	- porphyritic granite
Qrd	- rhyolite domes	Tqm	- quartz monzonite
Qra	- pyroclastic deposits	gd	- biotite diorite
Qrf	- rhyolite flows	hgn	- foliated hornblende granodiorite
Tgr	- fine-grained granite	PЄbg	- banded gneiss

Ut/R-005a

Figure 23

historic times. The May 18, 1980 eruption of Mount St. Helens attests to be the youth of volcanic activity here. The Cascade Range lies above the zone of subduction of the Juan de Fuca plate beneath the North American plate, (Fig. 2) and magma moving into the upper crust has transported large amounts of heat upward. In spite of the widespread, young volcanism, however, geothermal manifestations are not as plentiful as one would suppose they should be. The high rainfall and snowfall in the Cascades are believed to suppress surface geothermal manifestations through downward percolation of the cold surface waters in the highly permeable volcanic rocks. In the absence of surface manifestation, discovery becomes much more difficult.

No producible high-temperature hydrothermal systems have yet been located in the Cascades. Geological and geochemical evidence indicates that a vapor-dominated system is present at Lassen Peak in California, but it lies within a national park, and will not be developed. A hydrothermal system having temperatures greater than 200°C has been located at Newberry Caldera in Oregon through research drilling sponsored by the U. S. Geological Survey (Sammel, 1981), but the known the portion of the system lies within the caldera will not be exploited for environmental reasons.

Industry's exploration effort so far in the Cascades has been minimal, but has increased somewhat in the last several years as leases have been issued. The Department of Energy is currently sponsoring a cost-shared drilling program with industry to encourage more subsurface exploration in the Cascades. To date, one hole has been drilled in the south flank of Newberry volcano by GeoOperator, and results of that drilling will be made public soon.

The use of geothermal energy for space heating at Klamath Falls, Oregon is well known (Lund, 1980), and numerous hot springs and wells occur in both Oregon and Washington. Potential for discovery of resources in all temperature categories is great. Priest (1983) evaluated the geothermal geology of the Oregon Cascades in a very useful publication.

Columbia Plateaus

The Columbia Plateaus area is an area of young volcanic rocks, mostly basalt flows, that cover much of eastern Washington and Oregon and continue in a curved pattern into Idaho, following the course of the Snake River.

There are no hydrothermal resources having temperatures greater than 90°C known through drilling in this area. However, there are numerous warm springs and wells that indicate the presence of geothermal resources potentially suitable for direct heat uses.

Snake River Plain

The basalt flows and other volcanic deposits of the Snake River Plain are an extension of the Columbia Plateau eastward across southern Idaho to the border with Wyoming. The plain is divided into a western part and an eastern part. Thermal waters occur in numerous wells and springs in the western portion, especially on or near the edges of the plain. Geochemically indicated resource temperatures exceed 150°C at Neal Hot Springs and Vale, Oregon and Crane Creek, Idaho, but indicated temperatures for most resources are lower. Younger volcanic rocks occur in the eastern part of the plain, but no high-temperature resources (T>150°C) are yet identified, although numerous areas have warm wells and springs. This part of the plain is underlain by a high-flow cold-water aquifer that is believed to mask surface geothermal indications.

Direct use of hydrothermal energy for space heating is famous at Boise, Idaho, where the Warm Springs district has been heating homes geothermally for almost 100 years (Mink et al., 1977). Also in this area is the Raft River site where the Idaho National Engineering Laboratory of DOE constructed and operated a 5 MWe binary demonstration plant on a hydrothermal resource whose temperature is 147°C. This project is currently inoperative and the plant has been sold.

Rio Grande Rift

The Rio Grande Rift is a north-trending tectonic feature that extends from Mexico through central New Mexico and ends in central Colorado. It is a down-dropped area that has been filled with volcanic rocks and erosional debris from the bordering plateaus and mountains. The rift began to form in late Oliogocene times, and volcanic and seismic activity have occurred subsequently to the present. Young volcanism, faulting and high heat flow characterize the area today.

There are several low- and intermediate-temperature hydrothermal convection systems in this area, but the only high-temperature system that has been drill tested to any significant extent and where production is proven is a hot water-dominated system in the Valles caldera (Dondanville, 1978; Nielson and Hulen, 1984). Surface manifestations at the Baca No. 1 location in the caldera include fumaroles, widely distributed hot springs and gas seeps. Hydrothermal alteration extends over 40 km^2. Deep drilling has encountered a hydrothermal convection system in fractured Tertiary volcanic, Paleozoic sedimentary and Precambrian granitic rocks at an average depth of 2 to 3 km.. Temperatures as high as 300°C have been recorded. An attempt by DOE, Union Geothermal and Public Service Company of New Mexico to build a demonstration plant at that location failed when the steam supply proved to be inadequate. Also located near the caldera is the site of Los Alamos National Laboratory's hot dry rock experiment at Fenton Hill. Both the hot dry rock site and the hydrothermal convection system(s) probably derive their heat from magma that has provided the material for the several episodes of volcanism that created the caldera structure.

Elsewhere in the Rio Grande Rift, there are numerous hot springs and wells. Discovery potential appears to be high, although there are no known sites where fluids in excess of 150 to 170°C is indicated by present data (Harder et al., 1980).

Madison and other Aquifers

Underlying a large area in western North and South Dakota, eastern Montana and northeastern Wyoming are a number of aquifers that contain thermal waters. These aquifers have developed in carbonates and sandstones of Paleozoic and Mesozoic age. The permeability is both intergranular and fracture controlled in the case of the sandstones (e.g. the Dakota Sandstone) and fracture and solution cavities in the carbonates (e.g. the Madison Limestone). At least some of the aquifers will produce under artesian pressure. Depths to production vary widely but average perhaps 2,000 ft. Temperatures are 30-80°C (Gries, 1977) in the Madison but are lower in other shallower aquifers such as the Dakota. Direct use of the thermal water is being made at a few locations today, and it is evident that the potential for further development is substantial.

Balcones Zone, Texas

Thermal waters at temperatures generally below 60°C occur in a zone that trends northeasterly across central Texas. Many of the large population centers are in or near this zone, and there appears to be significant potential for geothermal development in spite of the rather low temperatures.

An initial assessment of the geothermal potential has been documented by Woodruff and McBride (1979). The thermal waters occur in a band broadly delimited by the Balcones fault zone on the west and the Luling-Mexia-Talco fault zone on the east. In many locations the thermal waters are low enough in content of dissolved salts to be potable, and indeed many communities already tap the warm waters for their municipal water supplies.

The geothermal aquifers are mostly Cretaceous Sandstone units, although locally thermal waters are provided from Cretaceous limestones and Tertiary sandstones. The thermally anomalous zone coincides with an ancient zone of structural weakness dating back more than 200 million years. The zone has been a hinge line with uplift of mountain ranges to the north and west and downwarping to the south and east. Sediments have deposited in the area of downwarping, and the rate of sedimentation has kept pace with sinking, keeping this area close to sea level. Structural deformation of the sediments, including faulting and folding, and interfingering of diverse sedimentary units have resulted in the complex aquifer system of today.

The source of the anomalous heat is not known with certainty but several postulates are (Woodruff and McGride, 1979): 1) deep circulation of ground waters along faults; 2) upwelling of connate waters, originally trapped in sediments now deeply buried; 3) stagnation of deep ground waters owing to faults that retard circulation; 4) local hot spots such as radiogenic heat sources (intrusions) within the basement complex, or; 5) other loci of high heat flow.

A minor amount of direct use is being made of these waters at present, and potential for further development is good.

Eastern Half of U. S.

Hydrothermal resources in other areas of the continental U. S. besides those mentioned above are very poorly known. There is believed to be potential for thermal waters of about 100°C at a number of locations along the Atlantic Coastal plain associated with buried intrusions that are generating anomalous heat through radioactive decay of contained natural uranium, thorium and potassium. Examples of such areas are shown on Figure 16 at Savannah-Brunswick, Charleston, Wilmington, Kingston-Jacksonville and the mid-New Jersey Coast. One drill test of such an area (Delmarva Peninsula near Washington, D. C.) has been conducted by DOE with inconclusive results regarding amount of thermal water that could be produced. Less than a dozen warm springs and wells are known at present. The Allegheny Basin is outlined on Figure 16 because it has potential for thermal fluids in aquifers buried deeply enough to be heated in a normal earth's gradient. Parts of Ohio, Kansas, Nebraska and Oklahoma as well as other states are believed to have potential for low-temperature fluids. No geothermal drill tests have been conducted, however.

Hawaiian Islands

The chain of islands known as the Hawaiian archipelago stretches 2500 km in a northwest-southeast line across the Pacific Ocean from Kure and Midway Islands to the Big Island of Hawaii. Built of basaltic volcanic rocks, this island chain boasts the greatest volcanic masses on earth. The volcano Kilauea rises 9800 m above the floor of the ocean, the world's largest mountain in terms of elevation above its base. The Kilauea, Mauna Loa and other vents on the big island are in an almost continual state of activity, but by contrast volcanos on the other islands have shown little recent activity. Haleakala on the island of Maui is the only other volcano in the state that has erupted in the last few hundred years, and the last eruption there was in 1790 (MacDonald and Hubbard, 1975).

Several of the Hawaiian islands are believed to have geothermal potential. The only area where exploration has proceeded far enough to establish the existence of a hydrothermal reservoir is in the Puna district near Kapoho along the so-called "East Rift", a fault zone on the east flank of Kileaua. Here a well was completed to a depth of 1965 m (Helsley, 1977) with a bottom-hole temperature of 358°C. Little is known in detail of the reservoir(s) at present, but they are believed to be fracture-controlled and water-dominated. A 3 MWe generator is currently being operated at the site. Exploration is currently underway by several companies in areas adjacent to the operating plant.

Elsewhere on the islands potential for occurrence of low- to moderate-temperature resources has been established at a number of locations on Hawaii, Maui and Oahu, although little drilling to prove resources has been completed (Thomas et al., 1980).

Alaska

Little geothermal exploration work has been done in Alaska. A number of geothermal occurrences are located on the Alaska Peninsula and the Aleutian Islands and in central and southeast Alaska. The Aleutians and the Peninsula overlie a zone of active subduction (Fig. 2), and volcanos are numerous. A hydrothermal system was located at Makushin volcano on the island of Unalaska (Reeder et al., 1985) and the island of Adak is also believed to have good discovery potential.

Low- and moderate-temperature resources are indicated in a number of locations in Alaska by occurrence of hot springs (Muffler et al., 1978). One area that has been studied in more detail and has had limited drilling is Pilgrim Hot Springs (Turner et al., 1980). This site is 75 km north of Nome, Alaska. Initial drilling has confirmed the presence of a hot water reservoir about 1 km^2 in extent that has artesian flow rates of 200-400 gallons/minute of 90°C water. Geophysical data suggest that the reservoir is near the intersection of two inferred fault zones. Further exploration work will be required to determine the potential of this reservoir.

Potential for Geothermal Development

Muffler et al. (1978) have dealt with the problem of how much accessible resource exists in the U. S. both at known sites and those that are undiscovered. They conclude that the undiscovered resource base is on the order of 3 to 5 times greater than the resources known today. These figures do not include possible hot dry rock or other more speculative resources. Table 2 is a summary of the current estimate of the geothermal resource base as taken from Muffler et al. (1978). This table demonstrates our lack of resource knowledge through the ranges and relative amounts of undiscovered resources and through the many missing numbers. We can conclude, however, that the geothermal resource base is large in the U. S.

TABLE 2

GEOTHERMAL ENERGY OF THE UNITED STATES
After Muffler et al. (1978) Table 20

RESOURCE TYPE	ELECTRICITY (MWe for 30 yr)	BENEFICIAL HEAT (10^{18} joules)	RESOURCE (10^{18} joules)
Hydrothermal			
Identified	23,000	42	400
Undiscovered	72,000-127,000	184-310	2,000
Sedimentary Basins	?	?	?
Geopressured (N. Gulf of Mexico)			
Thermal			270-2800
Methane			160-1600
Radiogenic	?	?	?
Hot Rock	?	?	?

EXPLORATION AND RESOURCE EVALUATION

Geothermal exploration may be divided into two types:

1) Exploring for geothermal resource areas, that is, locating geothermal resource areas, and
2) Exploring within geothermal resource areas, that is, defining the lateral and vertical boundaries and the properties of the actual reservoir(s).

In each case, the central problem of the geoscientific work is to site wells that intersect the resource and learn as much as possible about it. The main difference between the two problems is one of scale since many of the techniques used are common to both.

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 geologies are 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 led 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 hundreds of wells is available, the success rate for production is only about 80 percent. For wildcat geothermal drilling in relatively unknown areas the success rate is much lower -- about 15 percent for the Basin and Range Province of the western United States. The problem 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 system installation and maintenance. In many geothermal reservoirs, this means drilling into one or more fractures that are connected to the source area for the geothermal fluids. Although large blocks of rock in nature are 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 mostly indirect and provide only circumstantial evidence of the existence and location of the reservoir.

Geology

Collection of geologic data through surface geologic mapping and through logging of drill cuttings and core provides the basic data required for interpretation of all other exploration data. Often ignored or shortchanged in geothermal exploration, 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 (sedimentary rocks, plutonic rocks, volcanic rocks), (2) maps the structure within and among rock units (faults, fractures, folds, rock contacts), (3) studies the age relationships amoung rock units as shown by their mutual field relationships, (4) searches for evidence of geothermal activity, which evidence may range from obvious thermal springs, geysers and fumaroles to very subtle indications such as hydrothermal alteration of rocks, 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 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 1 million years old) volcanic rocks in the area that would indicate an underlying molten mass that could provide a source of heat?, (3) are there porous and permeable rock units or are there active faults or open rock contacts that could constitute a plumbing system?, and (4) is this a viable geothermal prospect area and if so what exploration techniques should be used next?

Stratigraphic Analysis. A thorough knowledge of the rock types in the prospecting area is fundamental. The geologist analyzes both surface outcrops and samples from drilling. He strives to identify rocks in the area that would make a good reservoir rock at depth, i.e. one that has adequate permeability or in which permeability may be developed. In a volcanic sequence, for example, sequences of young flows often are highly permeable whereas air fall or water-laid tuffs are easily altered to clay minerals and become impermeable. On volcanic islands, the portion of the rocks that were extruded in the atmosphere (subaerial) may be more permeable than those extruded under water (subaqueous). A volcanic island typically sinks during its formation, so the subaerial-subaqeous volcanic contact is found below sea level. The geologist will try to determine the affect that this change may have on permeability for a particular island and the expected depth of the contact. In areas like the Salton Trough, permeability is controlled by the type of rock (permeable sandstone or impermeable shale) and by its degree of metamorphism (high-temperature metamorphism causes the rocks to be brittle and to fracture whereas low-temperature metamorphism does not induce brittleness). It is obvious that an understanding of effects such as these is important to the success of a geothermal project.

Structural Analysis. A thorough knowledge of the structure of an area is important. Moore and Samberg (1979), for example, showed that at Cove Fort/Sulphurdale, Utah, much of the surface is covered by rock units that have slid into place from the east along an underlying nearly horizontal fault. Subsequent faulting has occurred along vertical faults, and the area now consists of separate fault blocks. One obvious implication from this discovery was that surface geology can be projected to depth only with great care. Faults can form zones of permeability if they fracture rock and create open spaces, or alternatively they can be filled with gouge, a rock flour that is quite impermeable. Gouge developed along faults can isolate the aquifers in individual fault blocks and decrease hydrologic communication across an area. Fluids trapped in sandstone aquifers between impermeable shale beds in isolated fault blocks can become highly pressured as the unit sinks through geologic time due to deposition of new sediments above. In such isolated blocks are found the geopressured resources of the Gulf Coast and elsewhere.

In places where faults intersect, permeability may be especially enhanced. It is important to determine the relative ages of faults and especially to be able to distinguish young faults and fractures from older ones. Older fualts are more likely to have had their open spaces filled by deposition of minerals. Relative ages of faults can sometimes be determined through detailed geologic mapping.

Age Dating. Certain minerals contain potassium, and a small percentage will be the naturally radioactive isotope K^{40}. This isotope decays to argon, $\nabla\theta^{40}$, with a half-life of about 1.2 billion years. By measuring the amount of A^{40} relative to the amount of K^{40} in a mineral, the time since A^{40} began to accumulate can be determined. In this way certain rocks can be dated. There are also other radioactive isotopes that can be used for dating. One must be careful about interpretation of the dates derived by these methods. In the case of K-Ar dating, for example, if the mineral being used for the dating has been heated sufficiently by a thermal event subsequent to its formation, the gaseous Ar may escape, thus resetting the radioactive clock to the date of the thermal event. Age dating has obvious use in geothermal exploration in terms of helping to locate young igneous rocks.

Geochemistry

A number of important exploration and reservoir production questions can be answered from studies of the chemistry of geothermal fluids and reservoir rocks, and so geochemistry plays a relatively important role in geothermal exploration (Ellis and Mahon, 1977). Geochemical reconnaissance involves sampling and analyzing waters and gases from hot springs and fumaroles in the area under investigation. The data obtained are then used to determine whether the geothermal system is hot-water or vapor-dominated, to estimate the minimum temperature expected at depth, to estimate the homogeneity of water supply, to infer the chemical character of the waters at depth, and to determine the source of recharge water. We will discuss some of the more important geochemical applications.

The processes causing many of today's high-temperature geothermal resources consist of convection of hot saline 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 (White, 1981).

Geothermal fluids contain a wide variety and concentration of dissolved constituents (Table 3). 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.

As geothermal fluids move through rocks, they interact chemically with the rocks, which themselves are usually chemically complex. Certain minerals in the reservoir rocks may be selectively dissolved by the fluids while other minerals may be precipitated from solution or certain chemical elements from the fluid may substitute for certain others within a mineral. These chemical/mineralogical changes in the reservoir rocks may or may not cause volume changes. Obviously, if the rock volume increases it must be at the expense of open space in the rock, which decreases 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 plugging. Thus, some hydrothermal systems form a sealed cap or a self-sealed zone above the reservoir and perhaps on the lateral boundaries also. In this self-sealing process, silica and calcium carbonate are the principal phases involved. The solubility of SiO_2 decreases with a decrease in temperature, with very little pressure effect. Calcite has a retrograde solubility, i.e., it is more soluble at low temperatures than at high temperatures. However, calcite solubility does increase rapidly with an increase in the partial pressure of carbon dioxide. Thus, as fluids which are saturated with calcium carbonate approach the surface, $CaCO_3$ is deposited as a result of the loss of CO_2. Other carbonate species such as dolomite ($MgCO_3$), as well as sulfates such as anhydrite ($CaSO_4$), show solubility relationships similar to those of calcite.

TABLE 3

REPRESENTATIVE ANALYSES OF GEOTHERMAL FLUIDS

Sample #	1	2	3	4	5	6	7	8	9	10	11	12
Temp °C	42	47	44.5	60		89	96		255	<260	292	316
pH		7.1	7.3	3.4	7.9	7.9	9.5		8.4			
SiO_2 (ppm)	52	11.3	157.3	136	289	293	373	400	690	563	705	400
Ca (ppm)	257	88.2	117.9	11.5	2.6	5.0	.8	10	17	8	592	28,000
Mg (ppm)	17	20.8	106	4.9	1.3	.8	.0	37	.03	<2	.6	54
Na (ppm)	578	11.3	228.9	5.7	247	653	230	117	1,320	2,320	6,382	50,400
K (ppm)		5.4	39.2	3.6	12.9	71	16	86	225	461	1,551	17,500
Li (ppm)	.5					.7	1.3		14.2	25.3	14.5	215
HCO_3 (ppm)		39.7	391		377	305	116	12.0		232	28	7,150
SO_4 (ppm)	932	57.3	748	126	340		89	414	36	72	<3.5	5
Cl (ppm)	625	7.4	110	7.4	9.6	865	30	10	2,260	3,860	11,918	155,000
F (ppm)	2.8	.25	.59	.5	.8	1.8	15	8	8.3	6.8		15
B (ppm)	2.6					4.9	2.0	24.1			13.4	390
As (ppm)						2.7			4.8	4.3		12

Sample Descriptions:

1. Hot spring; Monroe Hot Springs, Utah (Mundorff, 1970). Actively depositing travertine.
2. Hot spring; Yunotani geothermal field, Japan (Parmentier and Hayashi, 1981).
3. Hot spring; Yunotani geothermal field, Japan (Parmentier and Hayashi, 1981).
4. Acid sulfate water; Yunotani geothermal field, Japan (Parmenteir and Hayashi, 1981).
5. Water discharged from well; Yunotani geothermal field, Japan (Parmenteir and Hayashi, 1981).
6. Hot spring; Steamboat Hot Springs, Nevada (White et al., 1971).
7. Hot spring; Beowawe, Nevada, includes 149 CO_3 (Roberts et al., 1967).
8. Water discharged from well; The Geysers steam field (Frye; in Geothermal Resources Council Technical Session 5, 1980).
9. Well 44, Wairakei, New Zealand (Ellis and Mahon, 1977); pH measured at 20°C.
10. Brine discharged from well 54-3, Roosevelt Hot Springs, Utah (Capuano and Cole, 1981).
11. Analyses calculated from flashed brine, well M-26, Cerro Prieto (Fournier, 1981).
12. Brine discharged from well 11D, Salton Sea Geothermal Field (Palmer, T.D., 1975).

Other factors may also affect the deposition of carbonate and sulfate and other minerals, such as variations in pH, total pressure and partial pressure of oxygen. For example, subsurface boiling, accompanied by loss of CO_2, may cause the deposition of calcite, while the deposition of anhydrite may reflect the occurrence of locally oxidizing conditions produced when upwelling fluids contact aerated non-thermal groundwater.

This chemically complex hydrothermal system is dynamic through time, that is, for any given volume element in the reservoir, the fluid composition varies slowly with time, bringing about variation in the rock composition, porosity and permeability. However, because the rate of fluid circulation is perhaps only a few centimeters per year, in most hydrothermal systems a state of chemical equilibrium or near-equilibrium is observed to exist between reservoir fluid and reservoir rocks (Capuano and Cole, 1981; Helgeson, 1969). Lack of equilibrium could be evidence for rapid movement of fluid through the reservoir.

Chemical and Mineral Zoning. The hydrothermal mineral assemblages of active geothermal systems are dominated by clays or zeolites at relatively low temperatures, and by chlorite, illite, K-feldspar and epidote (or wairakite) at higher temperatures (Table 4). Quartz, calcite, pyrite and anhydrite are frequently associated with these minerals, and appear to form readily at both high and low temperatures. As expected, the distributions of the clay and silicate minerals is strongly temperature-dependent. At the lowest temperatures, below about 180°C, the stable assemblage consists of dolomite, kaolinite, montmorillonite and interlayered illite/montmorillonite. With increasing temperature and depth, montmorillonite, dolomite, kaolinite, and interlayered illite/montmorillonite disappear, and at temperatures above about 150°-180°C, the typical assemblage is illite, chlorite, potassium-feldspar and quartz. The calcium-aluminosilicates, wairakite and epidote appear only in rocks above 230-250°C. Prehnite, actinolite, diopside and biotite characterize the highest temperature assemblages associated with temperatures above about 300°C. One very important result of this mineral zoning is that the higher-temperature mineral assemblages cause the rocks to become brittle, and they fracture easily under the influence of tectonic movement and stress. This creates and renews fracture permeability in the higher-temperature parts of some Imperial Valley systems. In systems where base temperature is below

TABLE 4

SOME HYDROTHERMAL MINERALS IN SELECTED GEOTHERMAL FIELDS[1]

	Imperial Valley, California	Yellowstone, Wyoming	The Geysers, California	Pauzhetsk, Kamchatka	Matsukawa, Japan	Otake, Japan	Tongonan, Philippines	Kawah Kamojang, Java	N. Z. Volcanic Zone	El Tatio, Chile	Low temp Iceland	High temp Iceland	Larderello, Italy
Quartz	x	x	x	x	x	x	x	x	x	x	r?	x	x
Cristobalite		x		x	x	x	x	x	x	x			
Kaolin group	d	x	x	x	x	x	x	x	x	x			
Montmorillonite	d	x		x	x	x	x	x	x	x			
Interlayered illite-mont.	x			x	x	x	x	x	x	x	x	x	x
Illite	x	x	x	x	x	x	x	x	x	x			
Biotite	x			x					x				
Chlorite	x	x	x	x	x	x	x	x	x	x	?	x	x
Celadonite		x		x						x	x		
Alunite			x	x	x	x	x		x				
Anhydrite	x		x	x	x	x	x	x	x	x		x	x
Sulfur			x	x	x		x		x				
Siderite			x	x			x		x	x			
Ankerite	x			x								x	
Analcime		x		x					x		x	x	
Wairakite	x		x	x		x	x	x	x			x	x
Laumontite		x		x	x	x			x	x	x	x	
Heulandite		x		x		x	x		x		x	x	
Mordenite		x		x					x		x	x	
Prehnite	x			x					x		r?	x	
Amphibole	x			x	x	x			x			x	
Epidote	x			x		x	x	x	x		r	x	x
Sphene	x			x			x	x	x				
Adularia	x	x	x	x		x	x	x	x	x			x
Albite	x			x			x	x	x	x		x	
Rutile				x	x	x							
Leucoxene			x	x	x		x		x				
Magnetite									x				
Hematite		x		x			x	x	x	x	x		
Pyrite	x	x	x	x	x	x	x	x	x	x	x	x	x
Pyrrhotite	x						x		x				x
Base-metal sulfides	x	x		x		x							x
Fluorite		x						x					

(1) From Browne, 1978

Note: d = detrital, r = relict.
[a] includes Cerro Prieto, Baja, California, Mexico.
[b] deposited in discharge pipes and channels.

about 200°C, this development of enhanced permeability may not be possible.

Although the metal contents of many geothermal brines are significant, with the exception of pyrite and, in places, pyrrhotite, base metal sulfides are relatively uncommon at depth even in the deeper parts of the explored systems which deposit metal sulfides at the surface. The more commonly observed base metal sulfide minerals found at depth include sphalerite and galena, although chalcopyrite, arsenopyrite, nickel glaucodot, cobaltite and silver telluride also occur in rocks of the Broadlands field in New Zealand. In general the base metal sulfides in the Broadlands are present in rocks whose temperature range from 265°-300°C, whereas pyrrhotite is present above about 150°C (Browne and Ellis, 1970). The distribution of pyrite is not sensitive to temperature.

The interpretation of the mineral assemblages found in many thermal systems is complicated by the presence of minerals formed during earlier, frequently unrelated, hydrothermal events. The Roosevelt Hot Springs thermal system provides a situation where at least two distinct hydrothermal events can be recognized; an earlier event related to intrusion of the Tertiary Mineral Mountains pluton, and the present hydrothermal system (Nielson et al., 1978). Cross-cutting veins, identified in drill chips suggest that the depositional histories of these events was complex. The reservoir rocks consist of Tertiary granitic rocks and Precambrian gneiss and schist containing potassium feldspar, quartz, plagioclase, biotite and hornblende. The hydrothermal minerals include clays, illite, chlorite, calcite, pyrite, quartz, hematite, epidote and anhydrite.

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

The acid sulphate springs are typically a surficial feature produced by the oxidation of hydrogen sulfide to sulfuric acid. Altered ground surrounding the acid springs and fumaroles provides a striking example of reactivity of the waters. The altered areas are typically bleached and converted to a siliceous residue containing native sulfur, cinnabar, yellow sulfate minerals, and clay minerals including kaolinite and alunite. Similar acid alteration can, however, also be formed at depths where steam heating of groundwaters occur.

Chemical Geothermometry. 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 dissipated, and (2) fluids encountered during drilling may indicate that higher temperatures may be found elsewhere.

In the usual reconnaissance application, water samples are taken for analysis from springs and wells in the vicinity of the prospect. Proper sampling technique is very important. The samples must be filtered and properly acidified for preservation until analysis. At each sample location, pH and temperature are measured at the time of collection.

Both quantitative and qualitative geothermometers are available. The basic assumptions in application of quantitative geothermometers are that equilibrium has been reached in temperature-dependent chemical reactions between the reservoir rock and the fluid, and that no changes occur in the fluid after it leaves the reservoir and is sampled by a well or at the site of a natural spring. In this case, the chemistry of the fluid will reflect the chemical equilibrium at reservoir temperature, and analysis with subsequent interpretation yields an estimate of this temperature.

Several major-element geothermometers have been used successfully for estimating subsurface temperature, and a review of the geothermometers was given by Fournier (1981). For example in certain geothermal areas, the silica content of geothermal fluids appears to be limited above about 180°C by the solubility of quartz (SiO_2) and below 180°C by the solubility of amorphous silica, and both solubilities are temperature dependent. We thus can write:

$$T(^{0}C) = \frac{1309}{5.19 - \log_{10}C} - 273 \qquad 180^{0}C < T < 250^{0}C$$

(quartz geothermometer, no steam loss)

and

$$T(^{0}C) = \frac{731}{4.52 - \log_{10}C} - 273 \qquad 100^{0}C < T < 180^{0}C$$

(amorphous silica geothermometer)

where C = silica concentration in the geothermal fluid in mg/kg.

Other silica geothermometers are based upon equilibrium with chalcedony, α-cristobalite, or β-cristobalite, and it is obviously of importance to know which silica minerals exist in the reservoir rocks. If drill information is not available on this point, as it usually is not early in an exploration program, one must rely on the geologic mapping and inference to provide this information.

A second system of geothermometers is based upon the equilibrium reached among sodium (Na), potassium (K) and calcium (Ca) where reservoir rocks contain abundant quartz and feldspar (Fournier and Truesdell, 1973). One common geothermometer of this class is the sodium-potassium-calcium geothermometer:

$$T(^{0}C) = \frac{1647}{\log_{10}(Na/K) + \beta[\log_{10}(\sqrt{Ca}/Na) + 2.06] + 2.47} - 273$$

where $\beta = 4/3$ if < 100 and [1 cg ($\sqrt{Ca}$/Na) + 2.06] > 0, but if

T with $\beta = 4/3$ is > 100 or if [1 cg ($\sqrt{Ca}$/Na + 2.06] < 0, use $\beta = 4/3$.

Calculations are in mg/kg.

Different geothermometers frequently give different results when applied to the same fluid. Care must be taken in interpretation, and in this matter there is no good substitute for experience. Use of other data may help shed light on the relative reliability of the various geothermometers in specific geologic situations. For example, silica concentration can be affected by pH, and temperatures calculated from the sodium-potassium-calcium geothermometer may be in serious error if the CO_2 or magnesium concentrations are too high or if there has been addition of any of these elements through interaction of the fluid with sedimentary rocks or ion-exchanging minerals such as montmorillonite.

Mixing of the thermal reservoir waters with normal ground water can also change concentrations of the critical elements in a geothermometer, and can result in a calculated temperature that is either too high or is to low.

Isotope Geochemistry. There is strong evidence, based on stable isotope analysis, that the fluids which form hydrothermal convection systems are largely meteoric in origin (Craig, 1963). The evidence is based on an observed positive δO^{18} shift of geothermal fluids relative to meteoric fluids in the vicinity of the system (Figure 24). The relative abundances of the isotopes of oxygen is well known. The shift in geothermal waters is due to the interaction of the fluids with the heavier δO^{18} characteristic of host rocks. Since rocks contain negligible hydrogen, there is little change in deuterium, δD. Reviews of this topic may be found in Ellis and Mahon (1977). The δO^{18} and δD values of magmatic fluids are also shown in Figure 24, and it can be seen that for some systems the fluid may contain some component of magmatic water, but this is generally not thought to be large.

Isotopic studies can help answer questions on reservoir recharge system permeability. Isotopes have also been used to attempt to quantify the age of geothermal systems. The most successful has been tritium (^3H) which has a half life of 12.26 years. Minor amounts of tritium are produced by cosmic radiation in the stratosphere. However, major amounts have been put into the atmosphere by tests of thermonuclear weapons. Tritium concentration is expressed in terms of the Tritium Unit (T.U.) which is equivalent to T/H of 1 x 10^{-18}. In continental climates in the temperate zone cosmic radiation produces about 10 T.U. Up to 10,000 T.U. were measured in 1963 following extensive atmospheric testing of nuclear weapons. This decreased until about 1968, and since then as remained fairly constant. The following generalizations can be made concerning the age of water in the absence of mixing. A T.U. of less than 3 indicates that no water younger than 25 years is present. Values of 3 to 20 T.U. suggest that a small amount of thermonuclear tritium is present, which suggests that the fluids entered the groundwater environment in the 1954-1961 time frame. If greater than 20 T.U. are found, the water is younger than 1963.

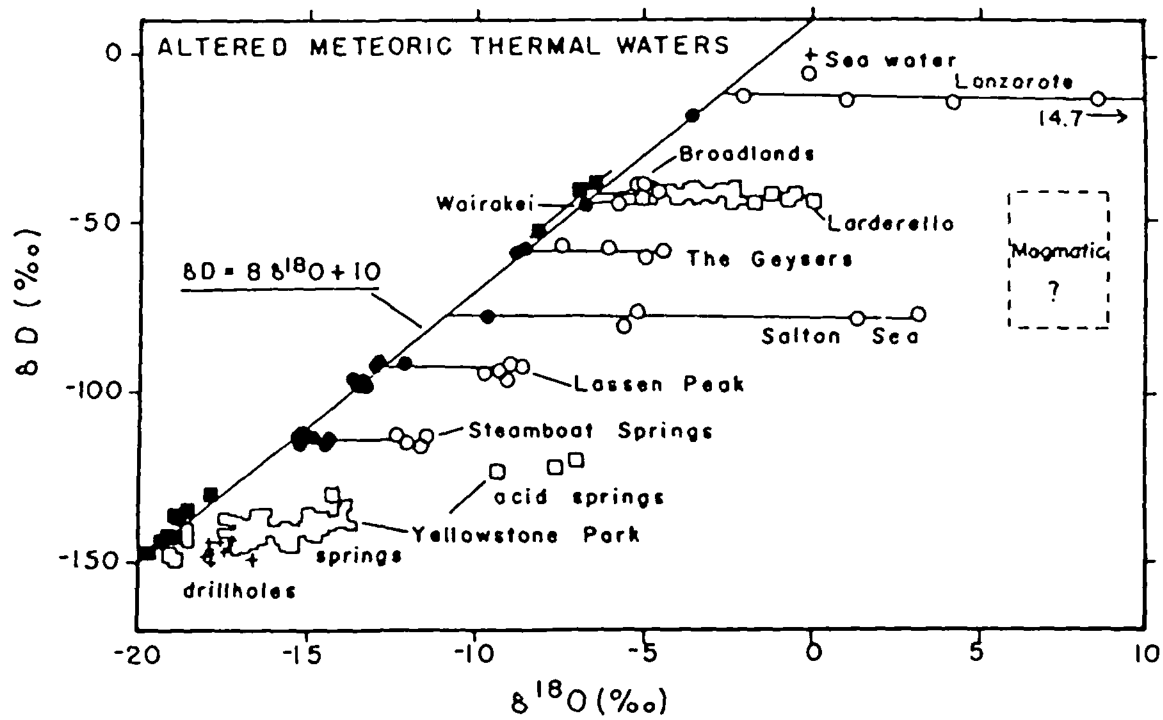


Figure 24. Oxygen-18 and Deuterium compositions of hot spring, fumarole, and drill hole thermal fluids derived from meteoric waters (o) and of meteoric waters local to each system (o).

Geophysics

Geophysical prospecting is the use of physical techniques and measurements either to detect directly the item sought or to provide indirect evidence of its existence and location. Such physical parameters as the distribution of temperature over the surface and at depth, the rate of heat flow to the surface, the electrical conductivity, magnetic susceptibility, density and elastic wave parameters of rocks can all respond in their own way to the presence of a geothermal resource (Ward, 1983; Wright et al., 1985) to an ore body, coal deposit or petroleum reservoir. We will discuss each of these methods briefly.

Thermal Methods. Under suitable circumstances, geothermal resources can be detected directly by application of one or more of the thermal methods. One basic parameter of interest is the heat flow, the rate at which heat flows upward toward the surface. We have seen previously that the outward flow of heat is a worldwide phenomenon (Sass et al., 1981), but in geothermal areas the heat flow is higher than this ubiquitous background amount, and so anomalous heat flow values may be clues to underlying geothermal resources.

The (vertical) heat flow is given as

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

where k = thermal conductivity (W/M-°C),
T = temperature (°C)
and z = the vertical coordinate in meters.

The quantity dT/dz is, of course, the vertical temperature gradient, the geothermal gradient, and in practice it is approximated by measuring temperature down a borehole and forming ratios $\Delta T/\Delta Z = (T_{z2} - T_{z1})/(z_2 - z_1)$ for various depth intervals. A typical value for the geothermal gradient is 30°C/km, or 0.03°C/m which is equivalent to 1.6°F/100 ft. 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. This need for samples of subsurface rocks exists in application of many geological, geochemical, geophysical and engineering techniques, and will be dealt with more fully below.

An often applied but dangerous shortcut to heat flow surveys is to forgo measurement of thermal conductivity, perhaps to save the cost of good sample collection and laboratory analyses, and to use only the thermal gradient data, $\Delta T/\Delta z$. Obviously, lateral as well as vertical variation of $\Delta T/\Delta z$ could be due to genuine changes in the heat flow field or simply to changes in lithology that are unrelated to any geothermal resource. And although it is often done, extrapolation of an observed temperature gradient to levels below the borehole is generally not justified and usually leads to disappointment.

Drilling can be expensive, and so the natural tendancy is to use thermal gradient or heat flow holes that are as shallow as possible. It is desirable to make the temperature measurements below the level affected by seasonal air temperature variations, and one is usually safe on this account with holes that are deeper than about 30 meters. Perhaps the biggest problem with shallow gradient holes, and deep holes in certain geologic environments, is movement of ground water. In some areas of sufficient topographic relief and abnormal precipitation, 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 pattern over the resource. It is imperative that one understands complications likely to be introduced into a heat flow or thermal gradient survey program before one embarks on use of this expensive technique.

Several workers have shown the utility in a few geothermal areas of the use of very shallow, say 2 meters deep, holes for heat flow studies or simply to measure anomalous near surface temperature. Such surveys at the Coso Hot Springs area in California show a +2C° anomaly over the reservoir in 2 m holes (LeShack and Lewis, 1983). Careful corrections must be applied for slope of the land, surface soil or rock type and vegetation (which affect reflectivity), surface hydrology, topography and other factors if these data are to be useful.

Existence of such shallow temperature anomalies implies that airborne or even satellite surveying in the thermal infrared region of the spectrum may be helpful. In practice, these methods have not been widely applied to date. Soil temperature fluctuations induced by sun angle variations, vegetation, ground slope and water table variations, to name a few variables, cause a high level of background "geologic" noise against which one must try to resolve the rare geothermal anomaly. Of course in specific areas, depending upon the geologic situation, infrared airborne surveying may be very helpful, but they would probably not constitute a first step in any exploraton program.

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 what they are associated can be detected.

With the exception of a few ore minerals, rock-forming minerals do not conduct electricity well. Electricity is conducted in the earth within waters that occupy the pore spaces in the rock. As we have seen, these waters invariably contain dissolved chemical species in the ionized state and the ions respond to an applied voltage difference between two points by moving through the water, thus sustaining a current. Two primary factors affect the conductivity of the groundwater: the concentraton of dissolved constituents and (2) the temperature. As we would expect, the higher the concentraton of dissolved species, the more ions in solution and the higher the rock conductivity (the lower the resistivity) will be. As temperature increases the activity of the ions increases, and so increasing temperature also causes increasing conductivity (decreasing resistivity). A third important cause of variation in rock conductivity is the amount, location and type of certain minerals that are capable of adsorbing ions from the solution or of ion exchange with the solution. If pore spaces or fractures are lined with minerals such as clays or zeolites, they may be holding loosely bound ions that are mobile enough to move along and within the layers of the clay structure in response to an electric field. Presence of clays and/or zeolites can greatly increase electrical conductivity (Moskowitz and Norton, 1977).

On the basis of the foregoing and from what we already know about the temperature, salinity and hydrothermal alteration within hydrothermal systems, one would expect such systems to be good electrical conductors. Indeed, low resistivity (high conductivity) has been discovered by surface surveys over many geothermal systems, and geophysical techniques that measure resistivity are in use worldwide in geothermal exploration (Ward and Sill, 1983). Most electrical techniques can be classified into one of those discussed below.

Galvanic Resistivity Surveys. In this technique two grounded electrodes are used to introduce a current flow in the earth, and voltage is measured between two separate grounded electrodes. There are several ways to deploy the electrodes, but perhaps the most useful configuration is the dipole-dipole array. Using this technique an effective depth of exploration of approximately two times the electrode separation can be achieved, and because the maximum practical value for separation is perhaps 500-750 meters, the resistivity method can detect low resistivity zones to depths of 7500 to 1500 meters. Volcanic areas often have high electrode contact resistance, causing low transmitted current and precluding deep exploration. Computer-aided interpretation methods are available and are easily applied. The method can be very useful for obtaining detail on a geothermal system.

Magnetotelluric (MT) Surveys. In this method, natural magnetic and electrical signals are used (Ward and Wannamaker, 1983). It can be shown that a measure of resistivity is given by the ratio of the electric field to the perpendicular magnetic field. Now, an electromagnetic field will penetrate into the electrically conducting earth to a depth dependant on its frequency. One can define a "skin depth" as that depth at which the electromagnetic field is attenuated in strength by the factor 1/e from its value at the surface, where e is the base of the natural logarithms, and equals approximately 2.72. Thus, lower frequency waves penetrate to deeper depths than do higher frequency waves, and by making simultaneous measurements of Ex and Hy for a range of frequencies, a depth sounding may be effected, the lower frequencies yielding information from deeper depths. Magnetotelluric surveys have been used with some success over a number of geothermal systems.

Magnetotelluric instrumentation incorporates the capability to make a tensor measurement, that is, to measure simultaneously both orthogonal electric field components (E_x, E_y) and all three orthogonal magnetic field components (H_x, H_y, H_z). This method is generally considered to be capable of exploration to depths of tens of kilometers, and to be capable of detecting magmas directly. Neither of these attributes is true in all cases. Although magma is conductive due to mineral semiconduction, the amount of contained water substantially affects the conductivity, dry magmas being much less conductive than wet ones (Lebedev and Khitarov, 1964; Wannamaker, in press). In geothermal exploration, it is possibly the wet magmas that we seek, however, because they have enough volatile content to produce the fracturing needed for hydrothermal convection. Depth of exploration depends to a certain extent on the near-surface resistivity structure. Also of great importance is the size and other characteristics of the magma body. Newman et al. (1985) have explored conditions under which crustal magma bodies can be detected. They conclude that if the body is isolated, i.e. has broken off from conductive magma at depth, it is more easily detected than if it maintains connective roots to the mantle.

The MT method has been used a great deal in geothermal exploration with generally disappointing results (Ward, 1983). By far the biggest problems appear to be misapplication and inadequate interpretation. Most MT data have been interpreted using one-dimensional inversion to a layered-earth resistivity structure. This method is totally inadequate in most geothermal exploration and usually produces misleading results. Full three-dimensional modeling is needed. The MT method has many subtleties, and must be applied with a great deal of care by geophysicists who are well experienced.

MT equipment can be considerably simplified if its range of operation is restricted to frequencies between about 1/10 Hz to 10,000 Hz, loosely called the audio range. This covers the depth range of usual interest in geothermal exploration. Therefore the AMT, or audiomagnetotelluric method, has seen some geothermal exploration. Most reported AMT surveys are scalar AMT, that is, only one component of electric field and one of magnetic field are measured at once. It can be demonstrated that in layered terrains this scheme is adequate for obtaining resistivity structure, but if resistivity also varies in either or both horizontal dimensions, as it invariably will in volcanic areas, scalar AMT is inadequate and is not recommended for exploration. For this task, tensor AMT measurement is needed. AMT has the advantage over conventional MT that it is less expensive both for surveying and interpretation.

CSEM. Controlled-source electromagnetic methods have been used as alternatives to galvanic resistivity or AMT surveying (Keller and Rapolla, 1974; Keller et al., 1982). A high-powered CSEM system has been developed and reported by workers at Lawrence Berkeley Laboratory (Wilt et al., 1981). The primary limitation of these techniques to date has been that interpretation methods have been limited to the one-dimensional case. Two- and three-dimensional algorithms are now becoming available, but further development is needed.

SP. Spontaneous-potential anomalies over convective hydrothermal systems arise from the electrokinetic and thermoelectric effects, which couple the generation of natural voltages with the flow of fluids and the flow of heat, respectively (Corwin and Hoover, 1979; Sill, 1983). SP surveys have been used successfully in certain volcanic terrains. On Hawaii, Zablocki (1976) found a large SP effect over the East Rift zone. Although these surveys are relatively inexpensive to run, they are also difficult to interpret in terms of the nature and location of the source area.

Seismic Methods

Elastic waves are transmitted through rocks and their measurement can be used to help determine the structure and properties of rock bodies. Two types of waves are most useful: (a) the compressional or primary (P) wave in which the particle motion is back and forth along the direction of travel of the wave, and (b) the shear secondary (S) wave in which the partial motion is perpendicular to the direction of travel of the noise. P-waves are ordinary sound-waves in rocks, and travel with typical velocities of 3-6 km/sec. S-waves have no analog in the air because fluids (liquids and gasses) do not support shear. In rock, S-waves travel at velocities about 70% of those of the P-wave.

Active Seismic Surveys. In this application, seismic waves are introduced into the earth, generally by explosion of a charge of dynamite or other explosive in a shallow borehole. Returns of seismic waves are measured at the surface. Coherent groups of waves are produced when a downgoing seismic wave reflects from a contact between two bodies of rock having different seismic velocities and/or densities and such reflections can be recognized in the surface seismic seconds and the depth to and attitude in space of the reflectors can be determined. In this way, much can be added to the knowledge of subsurface geology (Applegate et al., 1981). However, where the structure becomes complicated, diffraction of seismic waves occurs and makes the task of interpreting structure difficult. At Beowawe, Nevada, extensive and varied digital processing was ineffective in eliminating the ringing due to a complex near-surface intercalated volcanic-sediment section (Swift, 1979). This problem is typical in volcanic areas. 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 and for obtaining other structural-stratigraphic information. Because of access problems with large equipment, interpretational difficulties and the expense of reflection surveys, we see only limited use of this method in most volcanic exploration situations.

Passive Seismic Techniques. Seismic waves also occur naturally in the earth and such natural waves can be detected at the surface. There is limited evidence (e.g. Liaw and Suyenaga, 1982) that hydrothermal processes can generate seismic body waves in the frequency band 1 to 10 Hz. Noise also arises in such sources as traffic, trains, rivers, canals, wind, etc. Liaw and McEvilly (1978) have demonstrated that field and interpretive techniques for earth noise surveys require a great deal of understanding and care. These surveys can provide a guide to hydrothermal processes provided tha data quality is good and careful interpretation is done.

Microearthquakes frequently are closely related spatially to major geothermal systems. Accurate locations of these earthquakes can provide data on the locations of active faults that may channel hot water toward the surface. Microseismic activity is generally episodic rather than continuous, and this characteristic by provide a basic limitation to the technique in searching for or prioritizing geothermal prospect areas.

If a sufficiently distant earthquake is observed with a closely spaced array of seismographs, changes in P-wave traveltime from station to station can be taken to be due to velocity variations near the array. Traveltime residuals are computed as the observed arrival time minus that calculated for a standard earth. A magma chamber beneath a geothermal system would give rise to low P-wave velocities and hence to late observed traveltimes (Iyer and Stewart, 1977). While one can speculate that relative P-wave delays are caused by partial melts or magmas, they can also be caused by alluvium, alteration, compositional differences, lateral variations in temperature or locally fractured rock. Iyer et al. (1979) shows evidence for detection of magma in The Geysers using this technique.

Magnetic Methods

The earth has a main magnetic field that is similar in geometry to that of the classical bar magnetic and which is believed to arise from electrical currents flowing deep within the earth, in the electrically conducting, fluid core. This main field induces a magnetic response in certain magnetic minerals at and near the earth's surface and by detecting spacial variations in the total field the variations in distribution of magnetic minerals may be deduced and, through the process of interpretation, related to geology. The earth's field can be mapped on the ground, but it is much faster and cheaper to map this field from the air.

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 (Wright, 1981). Magnetic susceptibility often varies from 0 to 7000 μcgs units in these rock types and provides major magnetization changes which delineate geologic units. The scale of many geothermal systems is also similar to porphyry-type mineral occurrences.

Regional aeromagnetic data are often available as part of state or nationally sponsored surveys. These data often show major structural features and aid in forming a generalized geologic model for otherwise covered geology prospect areas. These regional data are generally too widely spaced and/or too high in altitude however, to warrant detailed quantitative model interpretation.

The locations of geologic structures (faults, fracture zones), intrusives, silicic domes and possibly major alteration areas (speculative) are often apparent on aeromagnetic data. Kane et al. (1976) show how gravity and magnetic data can be used to interpret structure at Long Valley, California. One other application of aeromagnetic surveying is in determining the depth to the so called Curie point isotherm. The Curie point of a rock is the temperature above which the rock ceases to be magnetic. Because of the increase in temperature with depth in the earth, the geothermal gradient, rocks below a certain depth will be non-magnetic because they will be above their Curie point, which for the mineral magnetite is about 585°C. In areas where the geothermal gradient is high, higher temperatures exist at shallower depths than elsewhere, and the Curie point isotherm is shallower. Through suitable interpretation techniques, usually computer based, the depth to the Curie point can be determined for an area from the aeromagnetic data from that area (Bhattacharyya and Leu, 1975). This gives one an idea of regions where the gradient is high, which would presumably be regions of greater geothermal potential. However, numerous limitations and assumptions apply to application of this interpretation technique, and such work must be under the direction of someone well experienced in it.

The Gravity Method

Minute variations in the earth's gravity field are caused by variations in density of subsurface rocks. In order to detect these variations very delicate instruments are required. The modern gravity meter measures 1 part in 10^9 of the earth's gravity field, and are among the most sensitive mechanical instruments yet made by man.

Gravity data are often acquired or compiled in the early stages of an exploration program. Regional data, with station densities of 1 station per sq km to 1 station per 25 sq km, may be available as the result of surveys by governments or universities. These data are generally 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, lithologic and other information. Isherwood (1976) interpreted the gravity survey of The Geysers area in terms of location of a subsurface magma chamber. In the Imperial Valley, California, gravity surveys have proved useful in locating areas where hydrothermal alteration and metamorphism have caused the sediments to become densified (Rex et al., 1971). Grannell (1980) showed how gravity could be used to monitor changes in the reservoir at Cerro Prieto due to production.

DRILLING AND LOGGING

Introduction

Geothermal drilling is generally done with somewhat modified, conventional drilling equipment. For wells that are expected to produce large quantities of hot water and/or steam for electric power production, large rotary drilling rigs of the type used for oil and gas wells are used, but if smaller quantities of lower temperature fluids for use in space heating are sought, a conventional water well rig might suffice. Because drilling is one of the most expensive steps in geothermal development and at the same time is typically fraught with problems and setbacks, it is important to choose the correct drilling contractor, equipment and techniques for the job at hand and not to compromise the drilling program for the sake of attempting to save money. One will usually come out ahead cost-wise by going at this job right in the first place rather than skimping and then running into expensive and time consuming difficulty. Even drilling contractors well experienced in other types of drilling will have trouble at first with geothermal drilling because there is not a direct transfer of oil field, water well or mining technology. The geothermal environment is characterized by temperatures above those in which most drilling equipment and muds were designed to operate. In addition, geothermal brines are especially corrosive due not only to their chemical composition but also to their elevated temperature. Special equipment and procedures are necessary to ensure the safety of the drilling crew and to minimize the possibility of blow-out. Besides all of these extra considerations, a good geothermal environment will have had a history of complex geological and geochemical processes that vastly increase the chances for difficulty during drilling attributable to zones of lost circulation, high angle faults or fractures that deflect the hole, or caving or sloughing of wall rock into the borehole.

During the drilling operation, the geologist, geochemist and geophysicist will be collecting exploration data important to their jobs. The geologist will log (record) information derived from examination of the drill chips or core. The geochemist will attempt to obtain uncontaminated samples of the subsurface fluids for analysis and the geophysicist will be concerned with geophysical well logging of the hole. We will consider these topics separately in what follows.

Lithologic and Mud Logging

Rotary drilling produces cuttings from the rock at the bottom of the hole which are brought to the surface by the drilling mud. These cuttings are collected at the shaker table on the rig, which separates them from the mud. It is the job of the geologist to specify sampling interval, sample size, collection method, washing procedures and packaging and labeling specifications. Grab samples from the shaker table at 3-meter intervals are usually recommended, and a minimum of 500 gm should be collected. When there is a great deal of lost circulation material in the mud, sample size should be increased to maintain the specified amount of rock chips. If the rock is hard, sample collection at the shaker table is no problem, but soft sands and especially clays may be lost through mixing with the mud. If high clay content is suspected, samples of the mud can be taken before the shaker table and analyzed. If the mud is oil-based, the cuttings must be washed immediately, otherwise washing is very difficult later. For water-based muds the geologist has options. Dried, unwashed samples are difficult to disarticulate, which argues for immediate washing. However, washing under laboratory conditions may produce better results.

As washed samples become available, the geologist logs these samples by examining them under a binocular microscope (Hulen and Sibbett, 1982). He identifies the rock type (lithology), notes evidence of hydrothermal alteration, looks for evidence of faults (development of fault gouge and slickensides), notes occurrence of vein minerals such as calcite or quartz and identifies material caved from the hole above the bottom. He records his observations on a standard log form.

If the drilling produces core, much more geologic information can be generated than from chips. The geologist estimates percentage core recovery by measuring the actual core and comparing to the footage drilled. The core is washed and a geologic log made by recording rock type, planar features, alteration, mineralization and other important features on a standard log form. The core may be photographed box by box. Some studies, such as core fluid analysis, require special handling and preservation. To preserve original fluids requires that the core be sealed in polyethylene tubes.

Fluid Sampling

It may be difficult and costly to obtain an uncontaminated sample of the subsurface fluids. Drilling muds are designed to hold back reservoir fluid pressure, and so there is usually a migration of mud filtrate into the rocks adjacent to the drill hole. In order to obtain an uncontaminated sample, enough fluid must be removed from the hole to eliminate contamination effects. There are several ways to do this, all of which require appropriate equipment and permits to handle large volumes of fluid at the surface. In the so-called "drill-stem test", the aquifer to be tested in packed off above and below and then flowed through the drill pipe until the chemistry of the produced fluid becomes stable. This is taken as an indication that uncontaminated reservoir fluid is being sampled. An alternative is to flow the entire uncased portion of the well, which may be cheaper because packers do not have to be set, but the procedure produces a mixed sample from several aquifers, and the interpretations that can be made from chemical analysis becomes more ambiguous and misleading. A third method of sampling is to case the entire hole, then perforate the casing at selected locations and obtain samples by flowing or bailing the hole.

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. Probes are lowered into the borehole to make these measurements. 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 understand fully the responses of various well logs in geothermal reservoirs and their typically fractured, altered, commonly igneous and metamorphic host rocks (Sanyal et al., 1980). In spite of the relative lack of knowledge of well-log reponse 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 a well (Hearst and Nelson, 1985).

A second, important problem in geothermal well logging is general lack of probes that will work in an environment where temperature exceed 200-250°C. Electric components and cables have generally not been available for temperatures this high and indeed have not generally been required by the petroleum industry, which makes the most use by far of well logging. Sandia National Laboratories has had an active and successful research program sponsored by the U.S. Department of Energy, to develop electronic components and logging tools for use in the geothermal environment, and so appropriate logging equipment is now becoming available.

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 may represent the only indication of permeable zones since production and injection tests can not, of course, be performed for cased intervals without perforation of the casing. 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 cross-calibration of logs that may be made with different instruments and different calibrations on successive logging runs. Few developers or drilling contrctors offer logging services themselves. Geophysycal logging of the well is almost alway done by a separate contractor.

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 herewith.

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

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

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 of detailed logging to detect zones of fluid flow or perform heat flow studies. 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. 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 induction log also measures the electrical resistivity or conductivity of surrounding rocks, but requires no electrode contact with the borehole wall, as the conventional resistivity logs do. Magnetic fields generated by coils in the probe are used to induce currents to flow in the rock and the response of the material surrounding the borehole is measured by other coils in the probe.

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 K^{40} to Ar^{40}. 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.

The gamma-gamma 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. Owing to lack of calibration, gamma density logging may not be as useful in igneous and metamorphic rocks as in sedimentary rocks. Densities of certain igneous and metamorphic rocks, for example, may exceed the calibration range of commercially available logging tools. Additionally, gamma-gamma 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. Hydrogen nuclei in the formation 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 and rock quality designation or intensity of fracturing. Fractures can be located by analyzing the full wave form of the incoming acoustic velocity signal.

The acoustic televiewer, also known as the borehole televiewer or seisviewer, provides, through complex instrumentation described by Hearst (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 geothermal system (Keys and Sullivan, 1979).

Well Logging Example. We have selected only one of numerous examples to illustrate the application of well logging to geothermal exploration. Murawato and Elders (1984) discuss well logs in the Salton Sea and Westmorland geothermal fields iN California. Figure 25 shows gamma ray, gamma-gamma density, spontaneous potential and resistivity logs for a hole that intersects sediments typical of the Salton Trough and basalt dikes. Note the expression of the dikes as opposed to the sediments on these data.

Cross plots of one type of borehole data vs. another can greatly facilitate data interpretation, particularly for boreholes in complex igneous and metamorphic terrain (Ritch, 1975; Glenn and Hulen, 1979). As an example of the utility of these plots, bulk density is plotted against neutron porosity in Figure 26 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, 1979). 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.

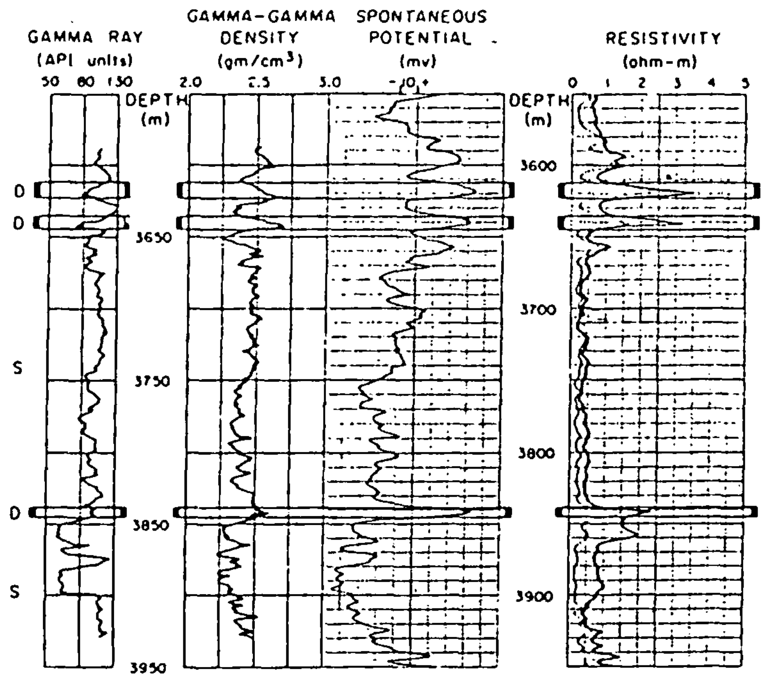
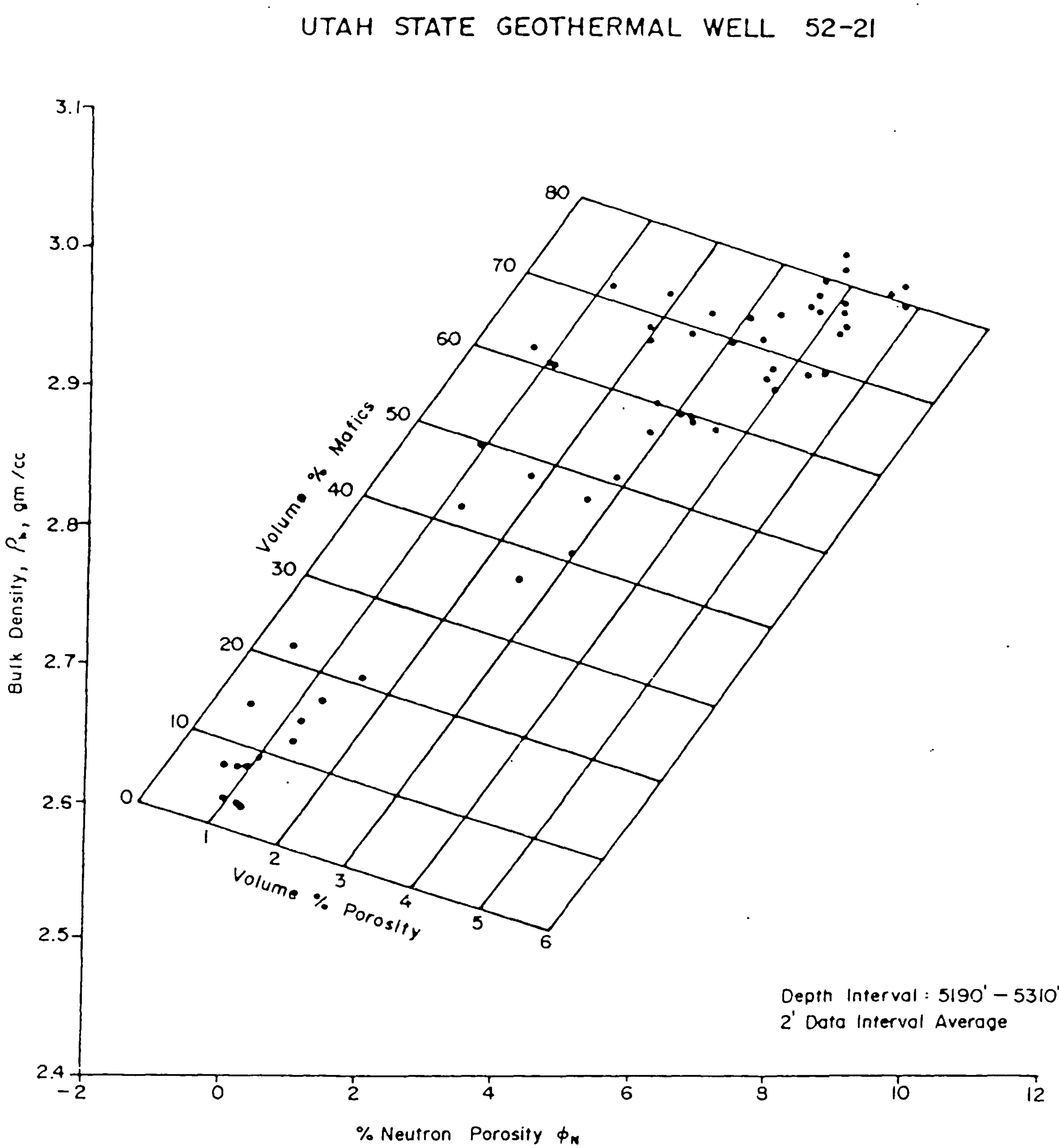


Figure 25. Geophysical well log from Magmamax 3, Imperial Valley, CA (from Muramoto and Elders, 1984).



UTAH STATE GEOTHERMAL WELL 52-21
Bulk Density, ρ_b, gm/cc
3.1
3.0
2.9
2.8
2.7
2.6
2.5
2.4
-2
0
2
4
6
8
10
12
% Neutron Porosity ϕ_N
Volume % Mafics
0
10
20
30
40
50
60
70
80
Volume % Porosity
1
2
3
4
5
6
Depth Interval : 5190' – 5310'
2' Data Interval Average

Figure 26

EXPLORATION STRATEGIES

Geothermal development is an interdisciplinary endeavor. Figure 27 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, there are simply no tools or techniques to solve a particular problem.

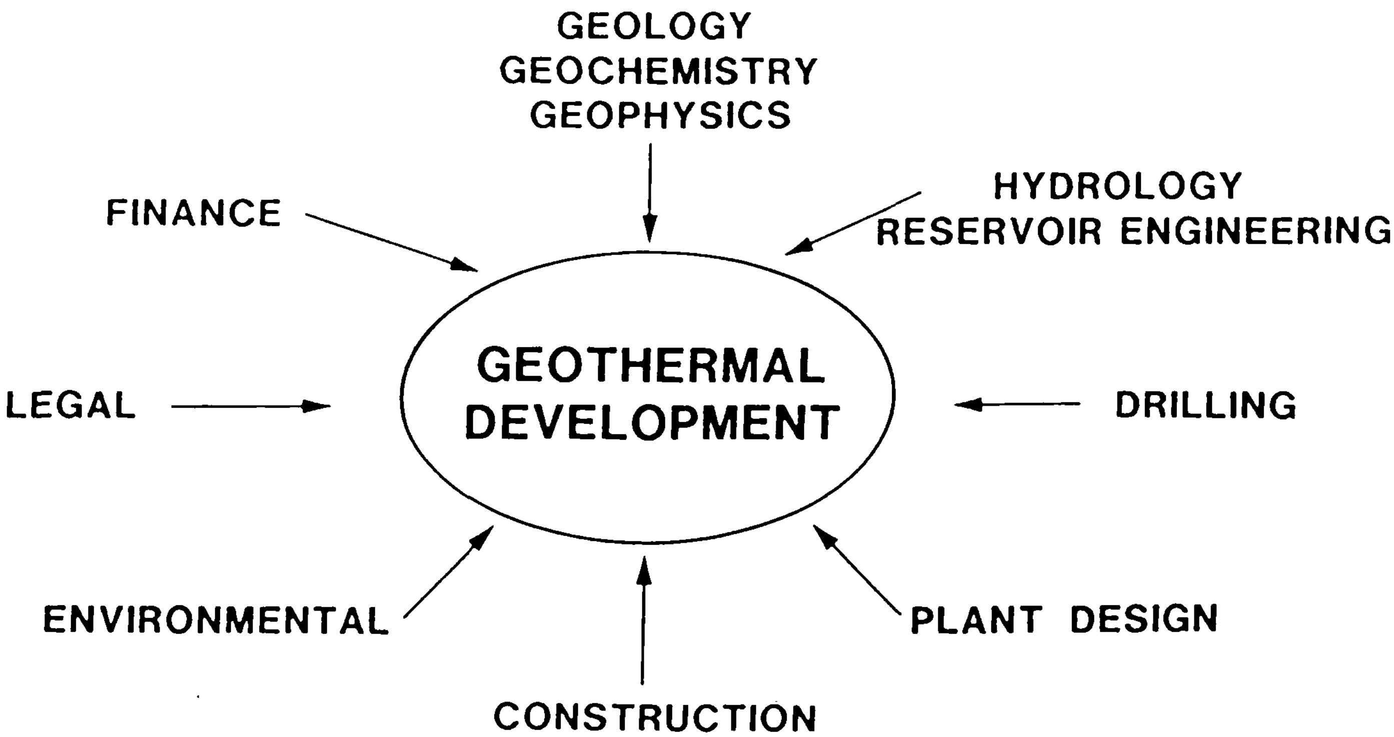
Geothermal Exploration

As previously mentioned, the geosciences have two primary applications in geothermal development: 1) exploration for geothermal systems, and 2) exploration within geothermal systems.

Figure 28 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.

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

GG-105

Figure 27

EXPLORATION AND EVALUATION SEQUENCE

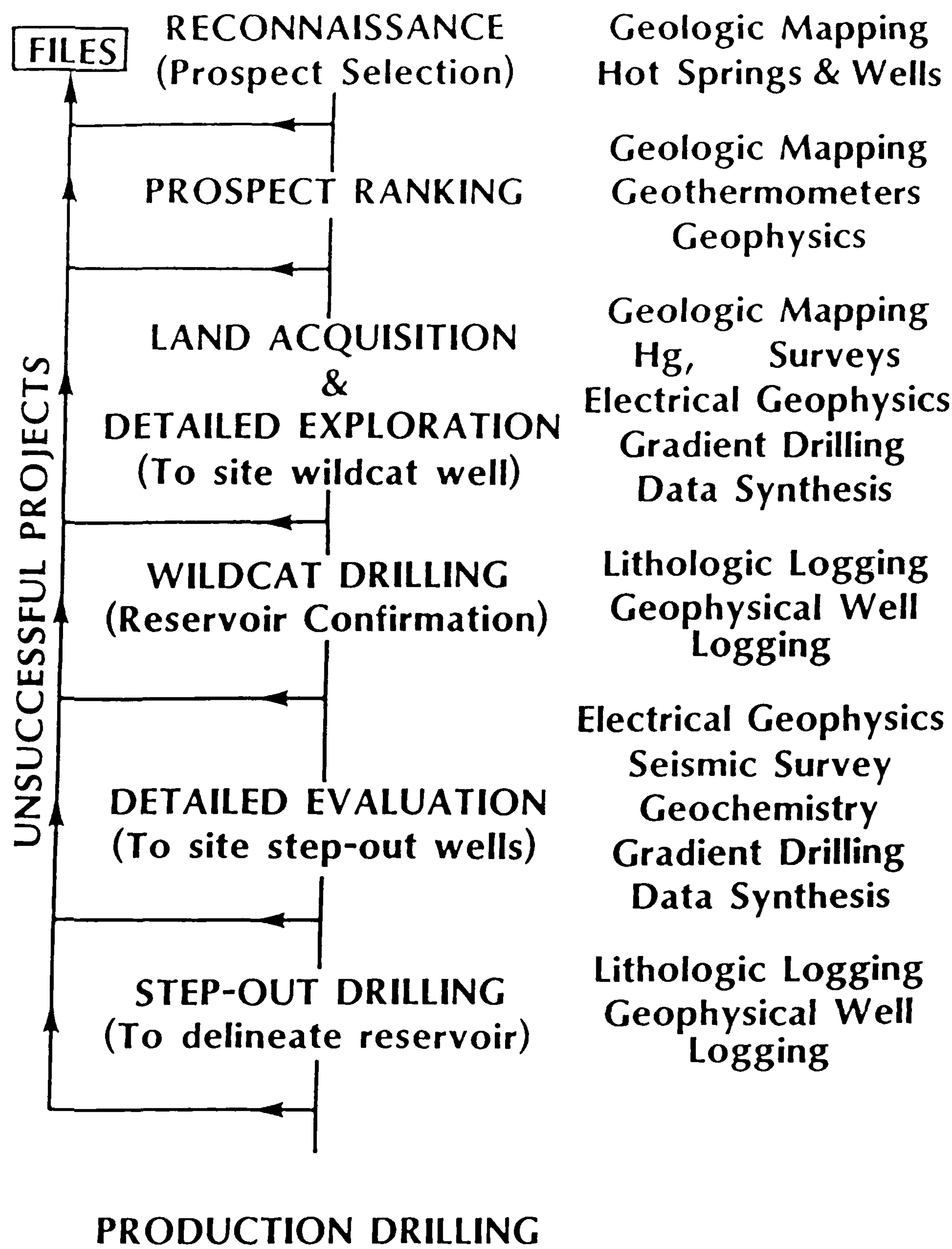


Figure 28

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 area must be designed specifically for application to that area by the geoscientists who are performing the work and interpreting the data.**

Exploration Strategy. Figure 29 is a diagram of a basic generic exploration strategy. Before such a strategy can become truly useful, much more detail must be added to each of the steps. Several aspects of Figure 29 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 assessing odds for success at each decision point, and then comparing the project to others or other uses of the money and manpower, optimum exploration will result and the risks and costs of exploration will be 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

Basic Generic Exploration Strategy

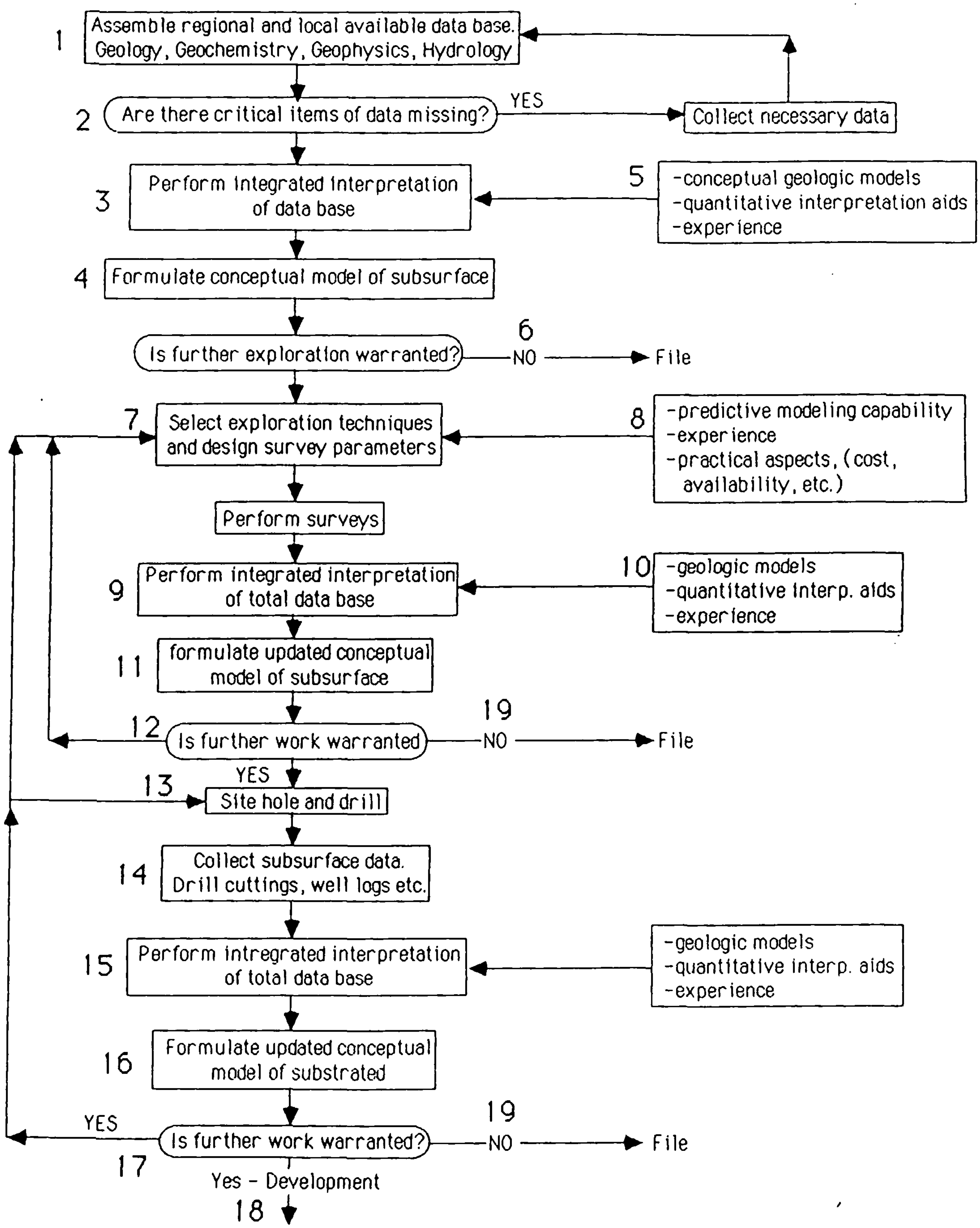


Figure 29

that money and human resources are deployed in the optimum way.

Available Data Base (1). All available regional and local geological, geochemical, geophysical and hydrological data should be assembled for the prospective exploration area and its surroundings. Once assembled, specialists in each of the earth science disciplines should assess the data in a preliminary fashion to determine its quality and to identify any obvious gaps (2). Often basic geologic data will be missing, 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 upon it and must be in agreement with it.

Integrated Interpretation (3). When the data base is judged to be sufficient, it should be interpreted by specialists. By "integrated interpretation" we mean to convey the necessity for the various specialists to work closely together in the data interpretation process. The objective of this integrated interpretation is to formulate a conceptual geologic model of the subsurface (4) in the exploration area that agrees with all of the available data. Of course, the model should concentrate on those features that are pertinent to the potential for occurrence of a geothermal resource in the area.

In order to perform this interpretation step, a number of ingredients must be available (5). These include (a) knowledge of geologic models of geothermal resources in other areas as a basis 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 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.

Exploration Techniques and Survey Design (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. 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. These will be used to help define lithology, petrography and hydrothermal alteration and for measurement of physical properties. Conventional geophysical well logs should be run in the hole, with a minimum logging suite probably being temperature, caliper, resistivity, gamma ray and acoustic logs. If the well is flowed or if there is a drill-stem formation test, samples of the fluids from the well should be carefully collected and preserved for analysis. Often a hydrothermal component of such fluid samples can be detected through chemical analyses, lending encouragement for further exploration. Chemical geothermometer calculations can be made from the analyses 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).

Basin and Range Exploration Strategy

Ward et al. (1981) reviewed a great deal of exploration data from the Basin and Range province and suggested the exploration strategy shown in Figure 30 for this area. I use it here as an example of the thought process that should be used to design strategies for other areas.

SUGGESTED HIGH TEMPERATURE HYDROTHERMAL EXPLORATION STRATEGY

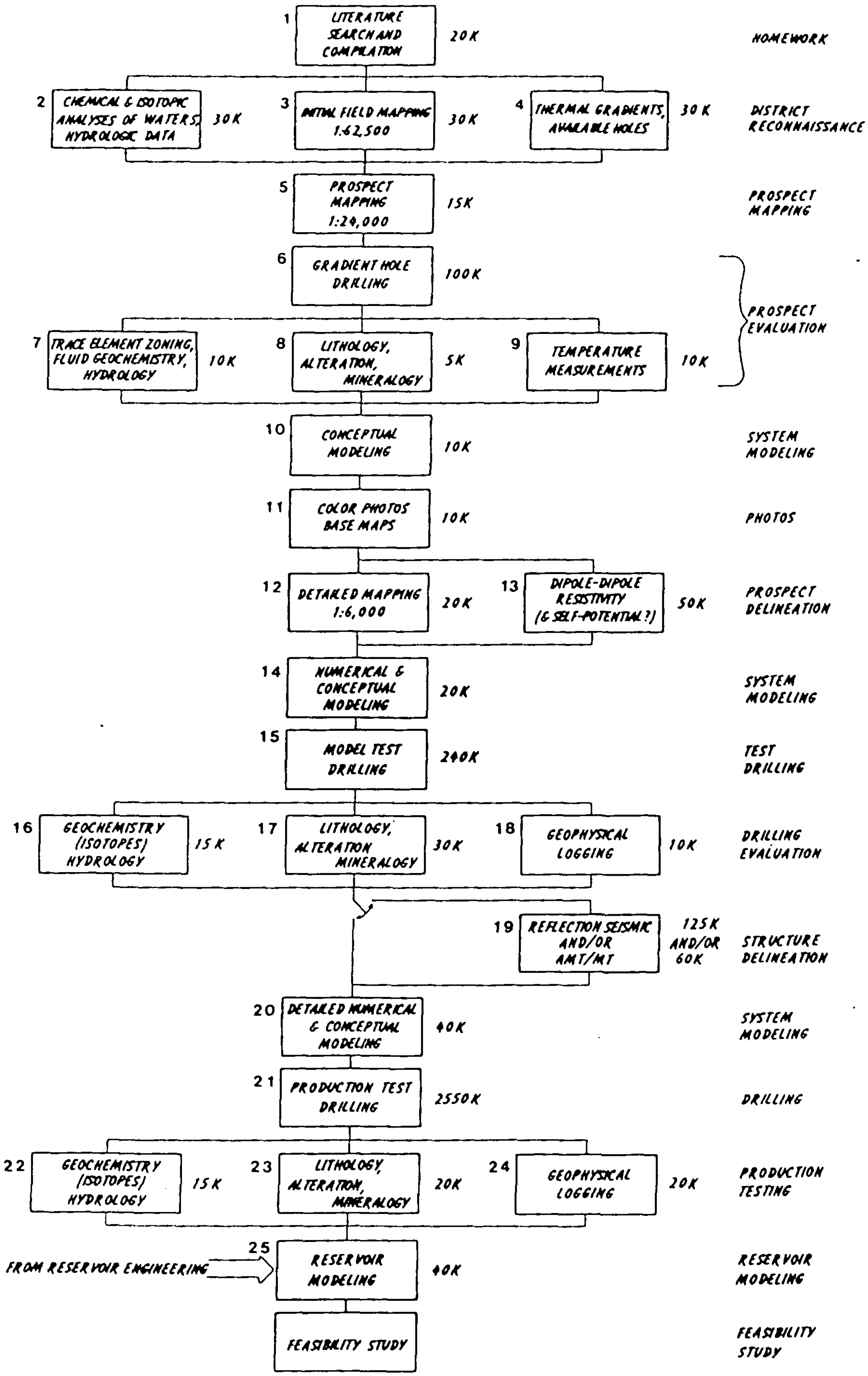


Figure 30

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EXPLORATION EXAMPLE -- COSO, CALIFORNIA

It seems appropriate in this overview to illustrate a few selected exploration data sets in the geothermal environment. Because of space limitations relative to the very large amount of data available, we have chosen just one area for which the geology is well known and where drilling has established the presence of a significant high-temperature convection system. The Coso geothermal system, Inyo County, southeastern California (Figure 16) provides an instructive example where both regional and detailed geophysical data contribute to an understanding of the geothermal resource.

Geologic Setting

The Coso geothermal area is located in the Coso Range of the western Basin and Range province, immediately east of the southern Sierra Nevada. Regional geologic mapping of the area was completed by Duffield et al. (1980), who expanded the results of several earlier workers. Northerly-trending fault-block mountains are formed of diverse lithologies which vary in age from Precambrian through Holocene. The oldest rocks are complexly folded Precambrian through Early Mesozoic marine sedimentary and volcanic rocks, many of which are regionally metamorphosed (Hulen, 1978). This older sequence is intruded by Jurassic-Late Cretaceous granitic stocks and plugs which appear to be portions of the southern Sierra Nevada batholith. Late Cenozoic volcanic rocks were erupted in two periods, 4.0-2.5 my and $\leq$ 1.1 my (Duffield et al., 1980), and formed domes, flows and pyroclastic deposits which covered much of the crystalline rocks in the Coso geothermal area. Hulen (1978) completed detailed geologic mapping and alteration studies of approximately 40 km^2 of the immediate Coso geothermal area in support of the U. S. Department of Energy drilling program at well CGEH-1. A generalization of his map, Figure 31. provides a useful reference base for our evaluation. Hulen (1978) and Duffield et al. (1980) describe hydrothermal alteration and active thermal phenomena (fumaroles, steaming boreholes, and "warm ground") which occur throughout an irregular 20 km^2 area along the eastern margin of the Coso rhyolite dome field. Drill hole CGEH-1 was drilled to a depth of 1 470 m in 1977 primarily in a mafic metamorphic sequence and a leucogranite which intruded the metamorphic rocks. This hole indicated temperatures in excess of 177°C and convective heat flow which appeared to be limited to an open

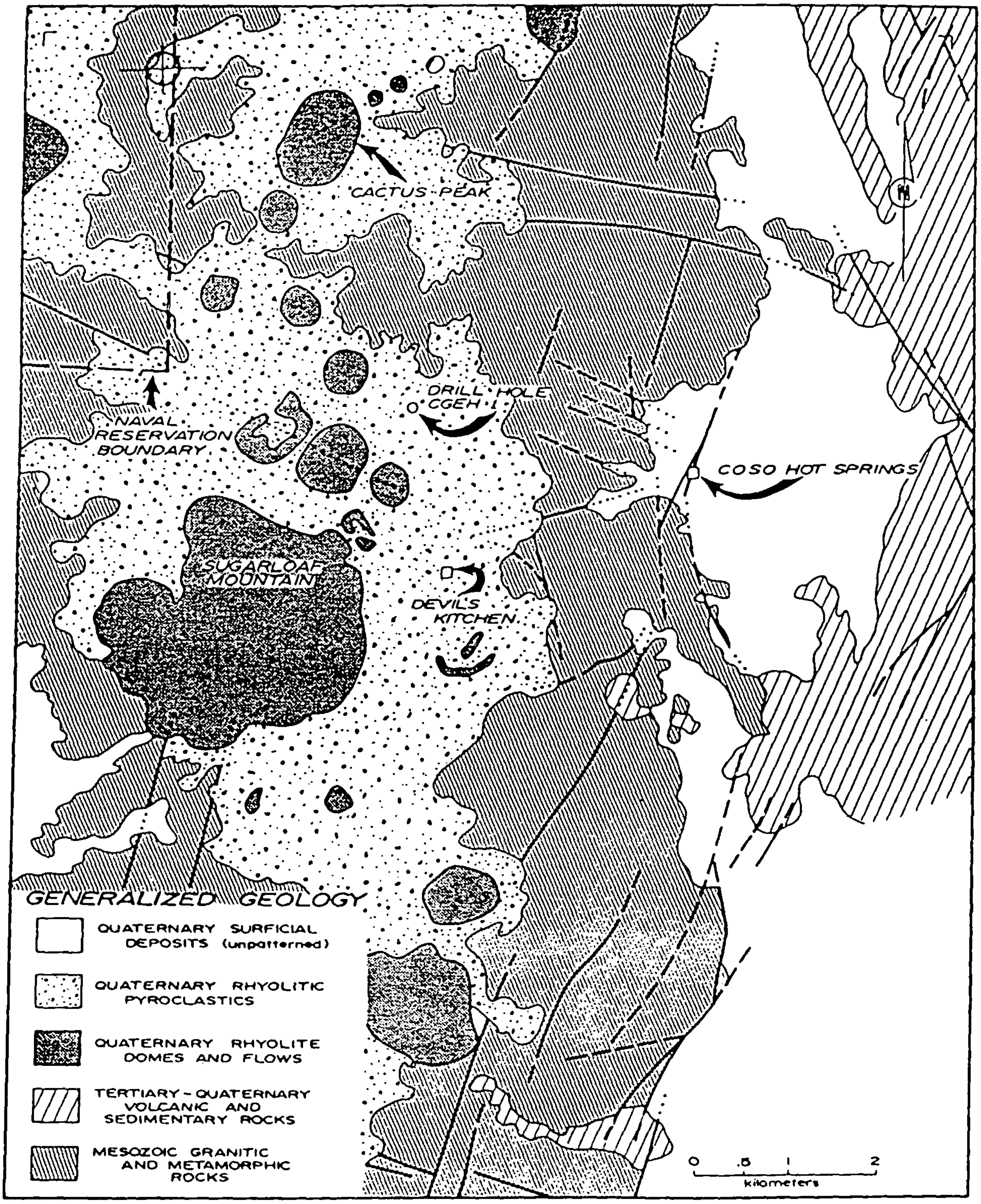

Figure 31. Geology of the Coso Geothermal Area, California.

fracture system between depths of 564 m and 846 m (Galbraith, 1978). Subsequently, several successful drill holes completed by California Energy Corporation have established the presence of a hydrothermal system.

Geophysical Studies

The Coso geothermal area is well expressed in quantitative thermal data. Combs (1980) completed a comprehensive study of the heat flow as determined in 24 shallow (35-110 m) and 2 deeper boreholes in an area of approximately 240 km^2 centered about the rhyolite dome field. He measured thermal gradients ranging from 25.3°C/km to 906°C/km, which he attributed to convecting hot water and former convective transport of heat by dikes that fed the domes and flows. Terrain-corrected heat-flow values ranged from 67 to 960 mW/M^2. The heat-flow anomaly is principally confined to the east-central portion of the rhyolite dome field and trends northeast to include Coso Hot Springs. LeSchack and Lewis (1983) describe shallow temperature surveys completed at Coso. The shallow (2 m) temperature measurements were made with a thermistor probe backfilled in a 2-m deep augered hole, after the thermistor equilibrated with surrounding earth temperatures. Temperatures of approximately 27.4 to 31.7°C form an anomaly pattern quite similar to the 400 mW/m^2 HFU contour of Combs (1980).

The University of Utah Research Institute completed a detailed dipole-dipole resistivity survey in September, 1977 as part of the U. S. Department of Energy resource assessment program which included the drilling of CGEH-1 (Fox, 1978a). A grid of three north-south lines and six east-west lines was surveyed to map the resistivity structure of a 41 km^2 area. An electrode spacing of 300 m was used for 41 line-km of survey, and a 150 m spacing for an additional 13 line-km. A $\leq$ 15 $\Omega \cdot m$ low-resistivity zone was observed in the survey in a background resistivity of ± 200 $\Omega \cdot m$, and the resistivity low includes Coso Hot Springs, Devil's Kitchen and much of the surrounding area.

A detailed low-altitude aeromagnetic survey of 927 line-km was completed over the Coso area by the University of Utah Research Institute for the U. S. Department of Energy in September 1977 (Fox, 1978b). The data were recorded on north-south flight lines with a 400 m line spacing at a mean terrain clearance of approximately 230 m. Basement lithologic and structural information are apparent in the magnetic data in the form of magnetic discontinuities which correspond in part to mapped faults and structural trends. Most significant is a broad magnetic low which covers about 26 km^2 in the southeast intersection of the two major trends. Rock magnetization measurements, geologic mapping and alteration studies indicate that the magnetic low is due in part to magnetite destruction resulting from hydrothermal alteration by the geothermal system, as well as to primary lithologic changes at depth.

The region that includes the Coso Range and the southern Sierra Nevada is one of the more active seismic areas in southern California as summarized by Walter and Weaver (1980), who established a 16-station seismographic network over an area approximately 40 km north-south by 30 km east-west in the Coso range as part of the U. S. Geological Survey studies to evaluate the geothermal resource potential. They recorded 4 216 local earthquakes ($0.5 < m < 3.9$) during the first 2 years of operation. Many of these events occured in a 520 km^2 area which included Coso Hot Springs (CHS), Devil's Kitchen and the rhyolite domes. In addition, Young and Ward (1980) presented a three-dimensional attenuation model for the Coso Hot Springs area as determined from teleseismic data. They determined that a shallow zone of high attenuation exists with the upper 5 km in the Coso Hot Springs-Devil's Kitchen-Sugarloaf Mountain area which they believed corresponds to a shallow vapor-liquid mixture or 'lossy' near surface lithology. No zone of significantly high attenuation was interpreted for the 5 to 12 km depth interval but high attenuation was noted below 12 km. Reasenberg et al. (1980) analyzed teleseismic P-wave residuals and mapped an area of approximately 0.2 s excess traveltime which they attributed to a low-velocity body between 5- and 20-km depth in the area of high heat flow and hydrothermal activity. They hypothesized that the low-velocity body could be caused by the presence of a partial melt in the middle crust.

Integrated Summary

Figure 32 summarizes the spatial overlap of the magnetic and resistivity lows, the 400 mW/m^2 heat flow anomaly and the anomalous ($\geq$ 26°C) ground temperatures at 2 m depth. The data are superposed on alteration and thermal features mapped by Hulen (1978). The prospect areas as indicated by the various data sets are generally in good agreement except perhaps for the

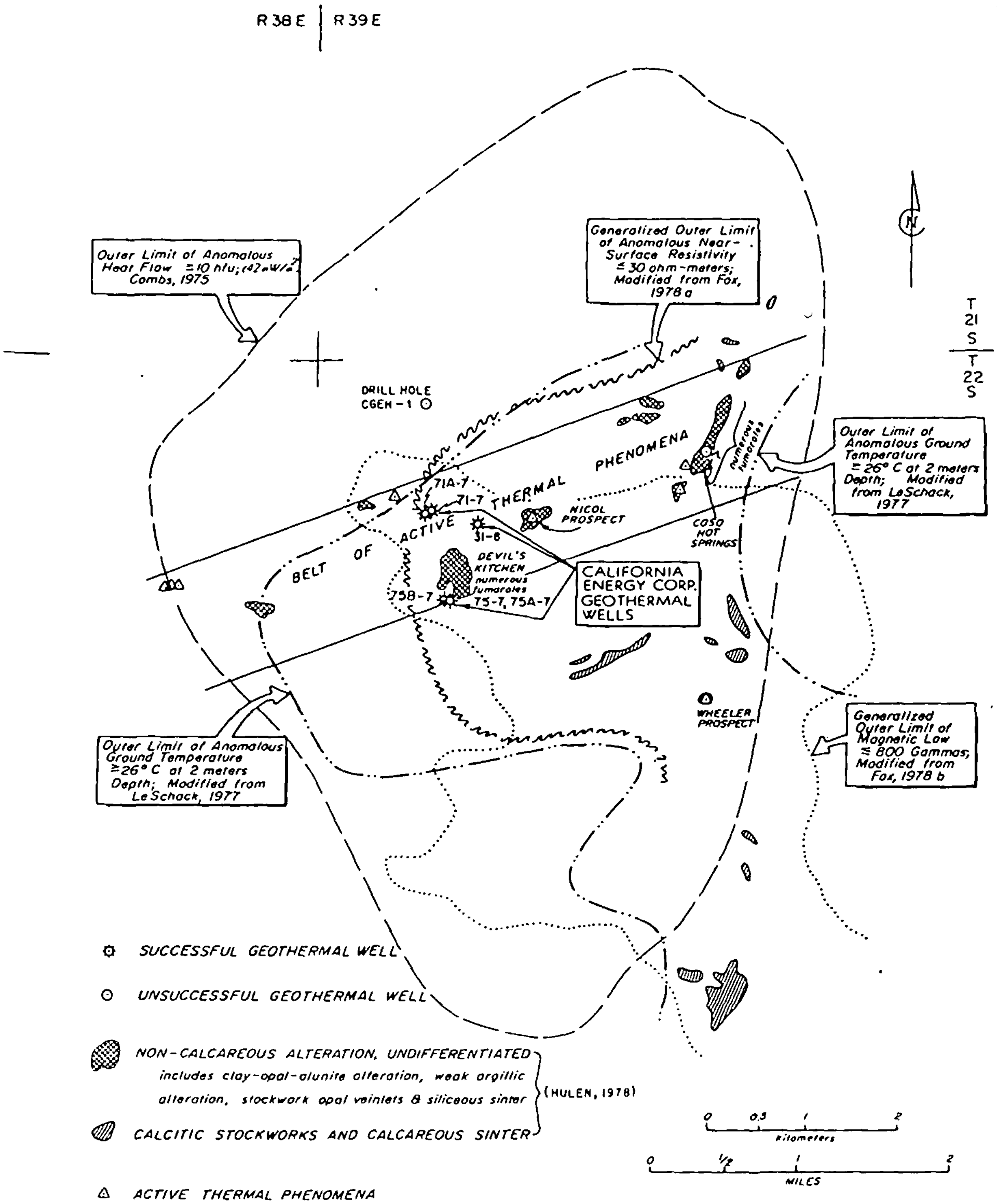

Figure 32. Geophysical Anomaly Summary for the Coso Geothermal Area.

extension of the heat flow high north of the belt of active thermal phenomena. The locations of several successful wells drilled by California Energy Corporation are also shown. This drilling has confirmed the presence of a high-temperature convective hydrothermal system in which the fluids are confined to major fracture zones within the crystalline rocks. Active exploration continues in the Coso area and future geoscientific studies and drilling will continue to improve our model of the hydrothermal system.

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THE NATURE AND GEOLOGIC CHARACTERISTICS OF GEOTHERMAL RESOURCES

by

Phillip M. Wright
University of Utah Research Institute

ABSTRACT

Geothermal resources occur where heat is concentrated in the upper levels of the earth's crust, ranging between the surface and about 5 km in depth. Such resources can be broadly divided into those with naturally occurring water, the so-called "hydrothermal" resources, and those without naturally occurring water, the so-called "hot rock" resources. At the present stage of geothermal resource development, we know by far the most about the hydrothermal resource type because this is the only type that can be economically developed today. Little deep drilling has been done in areas which would give us access to the hot rock resources.

Hydrothermal systems are found in many different geologic environments, and to a considerable extent, the characteristics of the environment determine those of the resource. In igneous and volcanic terranes, the hydrologic relationships are complex, and several reservoirs with different fluid compositions may be present. Fluid compositions depend on many factors, including temperature, rock type, origin of the fluids, residence time in the reservoir, boiling, mixing, cooling, fluid/rock interaction and mineral deposition. Heating of the fluids may be the result of nearby intrusive rock containing residual heat or may simply be the result of deep circulation and heating due to the earth's normal geothermal gradient. Permeability is usually controlled by faults, fractures and contacts between different rock types. These hydrothermal systems are usually volumetrically small and confined to a local area of heating and/or enhanced permeability. Typical areal sizes range from 0.1 sq km to 100 sq km.

The U.S. Geological Survey has assessed the geothermal resource base in the United States, and finds that the amount is large. Evaluation of volume and temperature data available in 1978 indicated that 1650 E18 joules of energy are present in 215 identified hydrothermal systems having temperatures greater than 90 deg C to depths of 3 km, excluding energy in National Parks. This is believed to be a minimum figure, but a more accurate estimate is not possible without more information. Electrical energy estimated to be producible from these resources is 23,000 megawatts for 30 years. The energy in the hot rock resources is very poorly known at the present time, but is probably at least two orders of magnitude more than the hydrothermal resource base. It is apparent that geothermal energy development can help replace the use of petroleum as that resource becomes more scarce and costly.

INTRODUCTION

Geothermal energy is heat that originates within the earth. At our current stage of technology, economic development of geothermal heat can be accomplished in a few areas where the heat is concentrated by geological processes. Approximately 4,733 megawatts of electricity (MWe) are currently being generated in 17 countries from geothermal energy, and about 10,000 thermal megawatts (MWt) are being used for direct heat applications. The United States produces 2,006 MWe of electrical power and uses 400 MWt in direct applications. While this is small compared to our use of an estimated 8.4 million MW of fossil energy (1), it nevertheless saves the consumption of 111 million barrels of oil per year worldwide and 35 million barrels per year in the U.S.

It is difficult to estimate the ultimate potential contribution of geothermal energy to mankind's needs for three reasons: 1) future energy costs are uncertain, and many lower-grade geothermal resources would become economic at higher energy prices; 2) only preliminary estimates of the worldwide resource base have been made, and; 3) technology is not yet available for using magma, hot rock, geopressured, radiogenic, and normal thermal-gradient resources, whose potential contributions are large.

The Earth's Internal Heat

Many large-scale geological processes are powered by redistribution of internal heat as it flows from inner, hotter regions to outer, cooler regions. Although the variations with depth in the earth of density, pressure and seismic velocity are well known, the temperature distribution is uncertain. We know that temperature within the earth increases with increasing depth (Figure 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 temperatures between 700 deg C and 1,200 deg 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 6,400 km deep, may as much as 6,000 deg C.

Because the earth is hot inside, heat flows steadily outward and is permanently lost by radiation into space. The mean value of surface heat flow is 82 E-3 watts/m2. Since the surface area of the earth is 5.1 E+14 m2, the rate of heat loss is about 42 million megawatts (1). White (2) estimates the total thermal energy above surface temperature to a depth of 10 km at 1.3 E+27 J, equivalent to burning 2.3 E+17 barrels of oil. The outward heat flux is about 5,000 times smaller than the flux of solar heat, and the earth's surface temperature is, thus, controlled by the sun and not by internal heat (3).

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

GEOLOGICAL PROCESSES

The genesis of geothermal resources lies in the geological transport of anomalous amounts of heat near enough to the surface for access. Thus, the distribution of geothermal areas is not random but is governed by geological processes of global, regional and local scale. Figure 2 shows the principal areas of known geothermal occurrences on a world map. Also indicated are areas of young volcanos and currently active geological structures. It is readily observed that geothermal resources occur in areas that have volcanic and other geological activity.

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

Heat Sources

In geothermal areas, higher temperatures are found at shallower depths than is normal. This condition usually results from either 1) intrusion of molten rock from great depth to high levels in the earth's crust, 2) higher-than-average surface heat flow, with an attendant high temperature gradient with depth (Figure 1), 3) ascent of ground water that has circulated to depths of 2 to 5 km, or 4) anomalous heating of shallow rock by decay of radioactive elements. Most high-temperature resources appear to be caused by the first mechanism.

A schematic cross section of the earth is shown in Figure 3. A solid layer, 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 lies below the crust. Mantle material below the lithosphere behaves plastically, flowing very slowly under sustained stress. The crust and mantle are composed of minerals whose chief building block is silica (SiO_2). The outer core is believed to be composed of a liquid iron-nickel-copper mixture while the inner core is a solid mixture of these metals.

Plate Tectonics. One geological process that generates shallow crustal heat sources in several different ways is known as plate tectonics (Figure 4). Outward heat flux from the deep interior is hypothesized to form convection cells in the mantle in which hotter material slowly rises, spreads out under the solid lithosphere, cools and descends again. The lithosphere cracks above areas of upwelling and is dragged apart along arcuate structures called "spreading centers", or "rift zones". These spreading plate boundaries are typically thousands of kilometers long, several hundred kilometers wide and coincide with the world's mid-oceanic mountain system (Figures 2 and 4). Crustal plates on each side of the rift separate a few centimeters per year, and molten mantle material rises in the crack, where it solidifies to form new crust. The upwelling of molten material brings large quantities of heat to shallow depths.

The laterally spreading plates press against adjacent plates, some of which contain the imbedded continental land masses, and in most locations the oceanic plates are thrust beneath the continental plates. These zones of under-thrusting, where crust is consumed, are called "subduction zones". They are marked by the world's deep ocean trenches, formed as the sea floor is dragged down by the subducted oceanic plate.

The subducted plate descends into the mantle and is warmed by the surrounding warmer material and by frictional heating. At the descending plate's upper boundary, temperatures become high enough in places to cause partial melting. The molten or partially molten rock bodies (magmas), ascend buoyantly through the crust (Figures 4 and 5) along zones of structural weakness, carrying their heat to within 1.5 to 20 km of the surface. They give rise to volcanos if part of the molten material escapes to the surface through fractures. Since the subducted plate descends at an angle of about 45 degrees, crustal intrusion and volcanos occur on the landward side of oceanic trenches 50 to 200 km inland. This is the process that causes the volcanos in the Cascade Range of California, Oregon and Washington, for example, and in many other parts of the globe as well.

Figure 2 shows where these processes of spreading, formation of new oceanic crust and subduction of oceanic plates are currently operating. Oceanic rises, where new crustal material is formed, occur in all major oceans. The East Pacific Rise, the Mid-Atlantic Ridge and the Indian Ridges are examples. In places, the ridge crest is offset by large faults that result from variations in the rate of spreading along the ridge. Such faults are called "transform faults".

Magmatic Intrusions and Intrusive Rocks. An ascending body of molten material may cease to rise at any level in the earth's crust and may or may not vent through erupting volcanos (Figure 5). Intrusion of magmas into the upper crust has occurred throughout geologic time. We see evidence for this in the occurrence of volcanic rocks of all ages and in the small to very large areas (hundreds of square miles) of crystalline, granitic rock, now exposed at the surface by erosion, that result when magmas cool slowly at depth.

Volcanic rocks extruded at the surface and crystalline rocks that have cooled at depth are known collectively as igneous rocks. They have a range of chemical and mineral compositions. At one end of the compositional 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 can be found in the rocks that compose the Hawaiian Islands. 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 black basalt and occurs mainly on land. The plutonic variety is granite. Magmas that result in basalt are termed "mafic" or "basic" whereas magmas that result in rhyolite or granite are termed "felsic" or "acidic".

The upper portions of the mantle are believed to be basaltic in composition. The great outpourings of basalt found on the ocean ridges and in places like the Hawaiian Islands seem to indicate a more or less direct pipeline from the upper mantle to the surface.

The origin of granites is a subject of controversy. Felsic magma can be derived by progressive segregation of the melt fraction from a basaltic magma as it cools and begins to crystallize. However, the chemical composition of granites is much like the average composition of the continental crust, and some granites also result from melting of crustal rocks due to heating by upwelling basaltic magmas. Basaltic magmas melt at a higher temperature and are more fluid than granitic magmas. Occurrence of felsic volcanic rocks of very young age (less than 1 million years and preferably less than 50,000 years) is a sign of good geothermal potential in an area because they may indicate a large body of viscous magma at depth to provide a strong heat source. On the other hand, occurrence of young basaltic rocks 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 (4).

Mantle Plumes. Another important source of volcanic rocks are point sources of heat in the mantle. It has been hypothesized that the upper mantle contains local areas of upwelling, hot material called "plumes". As crustal plates move over these hot spots, a linear or arcuate sequence of volcanos is developed. Young volcanic rocks occur at one end of the chain with older ones at the other end. The Hawaiian Island chain is an example. The youngest volcanic rocks on the island of Kauai on the northwest end have been dated through radioactive means at about 4 million years, whereas the volcanos Mauna Loa and Mauna Kea on the island of Hawaii at the southeast end of the chain are forming today and are in almost continual eruptive activity. To the northwest, the Hawaiian chain continues beyond Kauai for more than 2,000 miles to Midway Island, where the last volcanic activity was about 16 million years ago. The trace of the island chain is consistent with the motions of the Pacific plate as postulated by geophysicists from other data.

Thin Crust. 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 heat flow (100 E-3 watt/m2) and an anomalously high geothermal gradient (40 to 60 deg C/km). The typical geothermal gradient in the continental interior is 20 to 30 deg C/km. In the West, geologic evidence suggests that the crust is thinner than normal, accounting for upwarping of mantle isotherms and high measured geothermal gradients.

Fluids

For geothermal resources to be developed economically, an efficient means of bringing large quantities of heat to the surface is needed. Fortunately, nature provides water, which normally pervades fractures, pores and other open spaces in rocks. Water has a high heat capacity and a high latent heat of vaporization. Thus, it is an ideal heat-transfer fluid.

The density and viscosity of water both decrease as temperature increases. Water heated at depth is lighter than cold water in surrounding rocks, and is therefore subjected to buoyant forces. If heating is great enough for buoyancy to overcome the flow resistance of the rock, heated water will rise toward the earth's surface. As it rises, cooler water moves in to replace it. In this way, natural convection is set up in the groundwater around and above a source of heat such as an intrusion. Convection brings large quantities of heat within the reach of wells, and is, thus, responsible for the most economically important class of geothermal resources, the convective hydrothermal resources.

In some convective hydrothermal resources, the temperature never reaches the boiling point because of rapid water flux, and the system does not generate steam. However, in other systems pressure release (perhaps through sudden venting) causes the local boiling point to be reached, and steam is produced. The steam ascends and meets cooler rocks where it partially condenses while heating the rocks, and the pressure drop due to condensation brings up more steam. In this way, steam convection is set up. If venting exceeds recharge, the steam zone grows and steam will accumulate in the reservoir. The temperature and pressure in such a steam reservoir vary slowly with depth. At Larderello, Italy, the reservoir temperature and pressure are 240 deg C and 35 bars, values that appear to be typical of other vapor-dominated systems.

Permeability

Permeability is a measure of a rock's capacity to transmit fluid as a result of pressure differences. The flow takes place in pores between mineral grains and in open spaces created by fractures and faults. Porosity is the term given to the fraction of void space in a volume of rock. Interconnected porosity provides flowpaths for the fluids, and creates permeability, although there is no simple relationship between porosity and permeability.

Permeability and porosity can be primary or secondary, i.e. formed with the rock or subsequently. Primary permeability in sedimentary rocks originates from intergranular porosity and it usually decreases with depth due to compaction and cementation. In volcanic sequences, primary intergranular porosity and permeability exist, but primary permeability also exists in open spaces at contacts between individual flows and within the flows themselves. Secondary permeability occurs in open fault zones, fractures and fracture intersections, along dikes and in breccia zones produced by hydraulic fracturing (5) and (6). Permeabilities in rocks range over 12 orders of magnitude. Permeabilities in pristine, unfractured crystalline rock are commonly on the order of E-6 darcy or less. However, in-situ measurements at individual sites may vary by as much as 4 to 6 orders of magnitude, and zones of >100 millidarcy are commonly encountered. These higher permeabilities are due to increased fracture density.

Most geothermal systems are structurally controlled, i.e., the magmatic heat source has been emplaced along zones of structural weakness in the crust. Permeability may be increased around the intrusion from fracturing and faulting in response to stresses involved in the intrusion process itself and in response to regional stresses. Thus, an understanding of the geologic structure in a resource area can lead not only to evidence for the location of a subsurface magma chamber, but also to inferences about areas of higher permeability at depth. Such areas would be prime geothermal exploration targets. Regarding exploration for hydrothermal systems, the key problem appears to be more in locating permeable zones than in locating high temperatures. Fractures sufficient to make a well a good producer need be only a few millimeters in width, but must be connected to the general fracture network in the rock in order to sustain large fluid volumes.

CLASSIFICATION OF GEOTHERMAL RESOURCES

Geothermal resources can be classified as shown in Table 1, modeled after (7). To describe resources, we resort to simplified geologic models. A given model is often not acceptable in all details to all geologists. In spite of disagreement over details, however, the models presented below are generally acceptable and facilitate our thinking.

Geothermal resource temperatures range upward from the mean annual ambient temperature (10 to 30 deg C) to over 350 deg C (Figure 6). For convenience, geothermal temperatures are arbitrarily divided into high, intermediate or moderate, and low temperatures, corresponding to the ranges T > 150 deg C, 90 < T < 150 deg C, and T < 90 deg C, respectively.

Convective Hydrothermal Resources

Convective hydrothermal resources are geothermal resources in which the earth's heat is carried upward by convective circulation of naturally occurring hot water or steam. Underlying some high-temperature hydrothermal resources is presumably an intrusion of still-molten or recently solidified rock whose temperature ranges between 300 and 1,100 deg C. Other convective resources result from circulation of water down fractures to depths where the rock temperature is elevated even in the absence of an intrusion, with heating and buoyant transport of the water to the surface.

Vapor-Dominated Systems. Figure 7 (8) shows a conceptual model of a hydrothermal system where steam is the pressure-controlling fluid phase, a so-called "vapor-dominated" geothermal system. Convection of deep saline water brings heat upward to a level where boiling can take place. Boiling removes the latent heat of vaporization, thereby cooling the rock and water and allowing more heat to rise from depth. Steam moves upward through fractures and is possibly superheated by the hot surrounding rock. At the top and sides of the system, heat is lost from the vapor and condensation results, with the condensed water moving downward to be vaporized again. Within the vapor-filled part of the reservoir, temperature is nearly uniform due to rapid steam flux. If an open fracture penetrates to the surface, steam may vent or may heat the shallow ground water to boiling. Pressure within the reservoir is controlled by the vapor phase and increases slowly with depth. Because the surrounding rocks typically contain ground water under hydrostatic pressure, a large horizontal pressure differential exists between the steam in the reservoir and the water in adjacent rocks, and a significant question revolves around why the adjacent water does not move in and inundate the reservoir. We postulate that permeability at the boundaries of the reservoir is low either as a result of pre-existing geological features such as impermeable beds or faults, or that it has been decreased by deposition of minerals in the fractures and pores to form a sealed zone. The formation of a vapor-dominated system appears to require venting of steam at a rate in excess of water recharge to prevent flooding of the reservoir (8).

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

The Geysers geothermal area in California is an example of this type of resource. Other producing vapor-dominated resources occur at Lardarello and Monte Amiata, Italy, and at Matsukawa, Japan.

Water-Dominated Systems. Figure 8 (after Mahon and others, 1980) illustrates a high-temperature, hot-water dominated geothermal system. Models for such systems have been discussed by (8), (9), (10), and (11), among others. The heat source 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. Ground water circulates downward in open fractures and removes heat from these deep, hot rocks. Rapid convection produces uniform temperatures over large volumes of the reservoir. In some places, boiling may occur and a two-phase region may exist, but the pressure is controlled by water. Recharge takes place at the margins. Escape of hot fluids is often minimized by a near-surface sealed zone or cap-rock formed by precipitation of minerals in fractures and pore spaces. Surface manifestations include hot springs, fumaroles, geysers, travertine deposits, chemically altered rocks, or alternatively, no surface manifestation at all. If there are no surface manifestations, discovery is difficult and requires sophisticated geology, geophysics, geochemistry and hydrology.

Isotopic studies of hydrothermal fluids show that the bulk of the water and steam is derived from meteoric water (rain or snow), with the exception of those few systems where the fluids are derived from seawater or connate brines (12). Only a small percentage of the water comes from the intrusive rocks at depth. As the fluids move through the reservoir rocks, the compositions of both the fluids and the rocks are modified by the dissolution of primary minerals and the precipitation of secondary minerals. The entire hydrothermal convection system (rocks and fluids) is, in fact, a large-scale chemical reactor with interactions that are not completely understood today. The waters generally become enriched in NaCl and depleted in Mg. Salinities of high-temperature geothermal fluids range from less than 10,000 ppm total dissolved solids in some volcanic systems to over 250,000 ppm total dissolved solids in basin environments such as the Salton Sea, California (13) and (14). Table 2 shows some typical chemical analyses for hydrothermal fluids.

The pressure and temperature in most high-temperature hydrothermal convection systems lie near the curve of boiling point versus depth for saline water, and sporadic, local boiling occurs in many systems. Because boiling concentrates acidic gases (CO_2 and H_2S) in the steam, the oxygenated meteoric water overlying a boiling reservoir is heated and acidified. These acidic waters interact with the near-surface rocks to form certain hydrothermal minerals, typically clays, that can be used to help locate zones of subsurface boiling.

Hydrothermal Reservoirs. At this point, it is desirable to discuss the term "reservoir". The reservoir is the volume containing hydrothermal fluids at a useful temperature. The porosity of the reservoir rocks determines the total amount of fluid available, whereas the permeability determines the rate at which fluid can be produced. One must not envisage a large bathtub of hot water that can be tapped at any handy location, however. Both porosity and permeability vary over wide ranges at different points in the reservoir. A typical well encounters tight, hot rocks with steam or hot water inflow mainly along a few open fractures or over a restricted stratigraphic interval. Apertures of producing fractures may be as little as a few millimeters. Areas where different fracture or fault sets intersect or where fractures intersect favorable stratigraphic units may be especially favorable for production of large volumes of fluid. The longevity of a well depends upon how completely the producing zones are connected to the local and reservoir-wide network of porosity. If this inter-zone permeability is poor, the local open spaces are drained quickly and fluid production drops. However, if the well intersects a thoroughgoing geologic structure such as a major fault or fracture, the local producing volume around the well is recharged continuously, and fluid production can be maintained for many years.

Virtually all of industry's geothermal exploration effort in the United States is presently directed at locating vapor- or water-dominated hydrothermal systems having temperatures above 200 deg C. A few of the highest grade resources are capable of commercial electrical power generation today, and the majority of the growth in geothermal energy production is expected to come from hydrothermal resources until well into the next century.

Intermediate- and Low-Temperature Systems. The fringe areas of high-temperature vapor- and water-dominated hydrothermal systems often produce water of low and intermediate temperature. These lower-temperature fluids are suitable for direct-heat applications and may also be used for electrical power production as new binary conversion technology becomes available. 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.

Sedimentary Basins and Regional Aquifers

Some basins are filled to depths of 10 km or more with sedimentary rocks that have intergranular permeability. Such basins often contain accumulations of oil and gas. In some of the sedimentary units, circulation of ground water can be very deep. Vertical permeability is usually provided by faults. Water in deep rock units 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. Basin fluids range in chemical composition from relatively fresh water to highly saline. It is believed that many basin fluids were originally connate waters (trapped in the rocks at the time of formation) of seawater composition (15). Chemical interaction of these waters with rocks in the basin along their flowpaths leads to changes in the chemistry of the brine. Basins often contain evaporite beds of salts that dissolve easily in the basin fluids, bringing them to high salinities. An understanding of the chemistry of basin waters can sometimes lead to the identification of areas of upwelling fluids which may be thermally anomalous. Most basin waters are too low in temperature for the generation of electricity but may be used for direct applications such as space heating and greenhousing.

The Madison carbonate rock sequence of widespread occurrence in North and South Dakota, Wyoming, Montana, and northward into Canada contains warm waters that are currently being tapped by drill holes for space heating and agricultural purposes. In a similar application, space-heating systems installed in France use warm water contained in the Paris basin (16). Many other occurrences of this resource type are known worldwide.

Geopressured Resources

Geopressured resources also occur in basin environments. They consist of deeply buried fluids contained in permeable sedimentary rocks warmed in a normal or anomalous geothermal gradient by their great burial depth. The fluids are tightly confined by surrounding impermeable rock and bear pressure 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 9). A large amount of geopressured fluid is found in the Gulf Coast of the U.S. (Figure 12), where it generally contains dissolved methane. Therefore, three sources of energy are actually available from these resources: 1) heat, 2) mechanical energy due to the great pressure with which these waters exit the borehole, and 3) recoverable methane.

The U.S. Department of Energy, is currently sponsoring research to develop a better understanding of geopressured resources and exploitation technologies. Activities include the testing of geopressured wells to determine the nature and extent of the resource, its production characteristics and the potential environmental effects of long-term production. The research also includes the design and analysis of a total energy recovery system. These resources will probably contribute during the mid to late 1990s or the next century.

Radiogenic Resources

Research has been done that could lead to development of radiogenic geothermal resources in the eastern U.S. (17). The coastal plain of the East is blanketed by a layer of thermally insulating sediments. In places beneath these sediments, rocks occur that have an anomalously high rate of heat production due to decay of natural radioactive isotopes of uranium, thorium and potassium. These radioactive rocks represent old granitic intrusions, long since cooled. Methods for locating radiogenic rocks beneath sedimentary cover have been partly developed, and very limited drill testing of the geothermal target concept (Figure 11) has been completed under DOE funding, although no such research is being conducted by the federal government today.

Hot Dry Rock Resources

Hot dry rock resources 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 few pore spaces or fractures, and therefore contain little water and little or no interconnected permeability. The feasibility and economics of extraction of heat from hot dry rock has, for the past decade, been the subject of a $150 million research program at the Department of Energy's Los Alamos National Laboratory in New Mexico (18). Batchelor (19) describes similar research in England. Both projects indicate that it is technologically feasible to induce an artificial fracture system in hot, tight rocks at depths of about 3 km through hydraulic fracturing from a deep well. During formation of the fracture system, its dimensions, location and orientation are mapped using geophysical techniques. A second borehole is located and drilled such that it intersects the hydraulic fracture system. Water can then be circulated down one hole, through the fracture system where it removes heat from the rocks, and up the second hole (Figure 10).

The principal aim of the research at Los Alamos is to develop the engineering data needed for industry to evaluate the economic viability of candidate resources. The current plans are for a one-year flow test of the existing two-well system in order to determine production characteristics of the artificially created fracture system and its thermal drawdown and rate of water loss. Hot dry rock energy may contribute to our energy mix in the 1990s or in the next century.

Molten Rock (Magma) Resources

Experiments are underway at the Department of Energy's Sandia National Laboratories in Albuquerque, New Mexico to learn how to extract heat energy directly from molten rock. Techniques for locating a shallow, crustal magma body, drilling into it and implanting heat exchangers or possibly direct electrical converters are being developed (20). In Iceland, where geothermal energy was first used for space heating in 1928, technology has been demonstrated for economic extraction of thermal energy from young lava flows (21). A heat exchanger constructed on the surface of the 1973 lava flow on Heimaey recovers steam which results from downward percolation of water applied at the surface above hot portions of the flow. A space heating system which uses this energy has been operating successfully for over ten years.

GEOTHERMAL RESOURCES IN THE UNITED STATES

Figure 12 displays the distribution of known geothermal resources in the United States. Information for this figure was taken mainly from Muffler et al. (22) and Reed (23). Not shown are locations of hot dry rock or magma resources because very little is known. In addition, it should be emphasized that the present state of knowledge of geothermal resources of all types is limited.

Most of the hydrothermal resources and all of the presently known resources capable of electric power generation occur in the West. Large areas underlain by warm waters in sedimentary rocks exist in Montana, North and South Dakota, and Wyoming (the Madison Group of aquifers). Another important large area of low-temperature water is the north east-trending Balcones zone in Texas. The geopressured resource areas of the Gulf Coast and surrounding states are also shown. Resource areas indicated in the eastern states are highly speculative. Low temperature resources are much more plentiful than are high-temperature resources. Muffler et al. (22) and Reed (23) conclude that the cumulative frequency of occurrence increases exponentially as reservoir temperature decreases (Figure 13).

Let us consider the known geothermal occurrences in a bit more detail, beginning in the Western U. S. Figure 14 shows a physiographic map of the U.S. to help in locating the areas discussed, and Table 3 lists the geologic time scale.

Salton Trough/Imperial Valley, CA

The Salton Trough lies along the landward extension of the Gulf of California. It is composed of the Imperial Valley in the U.S. and the Mexicali Valley in Mexico. The area is one of complex, currently active plate tectonic processes. The crest of the East Pacific Rise spreading center is offset repeatedly northward up the Gulf of California by transform faulting (Figure 2). Both the rise crest and the transform faults come onto the continent under the delta of the Colorado River (Figure 15) and the structure of the Salton Trough suggests that they underlie the trough.

The Salton Trough has been an area of subsidence since Miocene times (7-23 million years before present, mybp). Sedimentation in the tough has paced subsidence, with debris from the Colorado River predominating. At present, 3 to 5 km of poorly-consolidated sedimentary material overlie a basement of Mesozoic crystalline rocks that intruded Paleozoic and Precambrian sedimentary rocks. Detailed analysis of drilling data and of surface and downhole geophysics indicates that at least some of the known geothermal occurrences (Cerro Prieto, Brawley and the Salton Sea) are underlain by pull-apart basins apparently caused by crustal spreading above a local section of the East Pacific Rise crest (24). Very young volcanic activity has occurred at Cerro Prieto where a rhyolitic volcanic cone is known, and along the southern margin of the Salton Sea where rhyolite domes occur. The Salton Sea domes are approximately 60,000 years old (25).

The Cerro Prieto hydrothermal field provides an example of a Salton Trough resource type. This field is water-dominated producing from depths of 1.5 to over 3 km. Fluid temperatures range from about 200 deg C to over 350 deg C (26). The rocks are composed of an upper layer of unconsolidated silts, sands and clays, and a layer of consolidated sandstones and shales overlying the crystalline basement (27). Two principal reservoir horizons occur in sandstones within the consolidated sequence. Enhanced production has been noted in the vicinity of faults, indicating that fracture permeability is important, although intergranular permeability due to dissolution of minerals by the geothermal fluids is believed to be important also (28). Reservoir recharge is apparently from the northeast and east and consists partly of Colorado River water (29). A conceptual model of fluid flow at Cerro Prieto (Figure 16) has been developed by Halfman et al. (30). They conclude that water flows upward from depth within permeable sandstone units that have a shallow dip. The permeable units are overlain by impermeable shales, and the water gains access to permeable units higher in the section through breaks in the shales.

The geothermal fluid from Cerro Prieto, after steam separation, contains about 25,000 ppm total dissolved solids. This figure is much lower than some of the other resources in the Salton Trough. For example, the Salton Sea hydrothermal field contains 20 to 30 percent by weight by solids.

The Geysers, CA

The Geysers geothermal area is the world's largest producer of electricity from geothermal fluids with about 1,800 MWe from 22 plants on line and an additional 800 MWe scheduled. This area lies about 150 km north of San Francisco. The portion of the resource being exploited is a vapor-dominated field having a temperature of 240 deg C. The ultimate potential of the vapor-dominated system is not known. Associated with the vapor-dominated field are believed to be several unexploited hot water-dominated reservoirs whose volumes and temperatures are unassessed (Figure 17).

The geology of The Geysers area is complex, especially structurally. Reservoir rocks consist mainly of fractured greywackes, which are sandstone-like rocks consisting of poorly sorted fragments of quartzite, shale, granite, volcanic rocks and other rocks. Fracturing has created the reservoir permeability. Overlying the reservoir rocks is a series of impermeable metamorphosed rocks (serpentinite, geenstone, melange and metagranite) that forms a cap on the system.

The presently known steam field is confined between the Mercuryville fault zone on the southwest and the Collayomi fault zone on the northeast (Figure 18). The northwest and southeast margins of the steam field are not definitely known. Surface manifestations of the steam field include two small areas, the largest one being known as The Big Geysers, an area of hot springs, fumaroles and hydrothermal alteration. The extent of surface manifestations is curiously small compared to the large size of the underlying steam resource.

To the east and northeast lies the extensive Clear Lake volcanic field composed of dacite, rhyolite, andesite and basalt. The interval of eruption for these volcanics extends from 2 million to 10,000 years ago, with ages progressively younger northward (31). The Clear Lake volcanics are very porous and soak up large quantities of surface water. It is believed that recharge of a deep, briny hot-water reservoir comes from water percolating through the Clear Lake volcanics, and that this deep reservoir supplies steam to the vapor-dominated system through boiling (Figure 17), although the deep water table has never been intersected by drilling. Geophysical surveys indicate the presence of a large magma chamber underlying the Clear Lake volcanic rocks and centered on Mt. Hanna, immediately northeast of the Collayomi fault zone (32).

Basin and Range

The Basin and Range province extends northward from Mexico into southern Arizona, southwestern New Mexico and Texas on the south, through parts of California, Nevada and Utah, and becomes ill-defined beneath the covering volcanic flows of the Columbia Plateau and the Snake River plain on the north (Figure 14). The northern portion of this area contains abundant geothermal resources of all temperatures. Resources along the eastern and western margins of the province appear to be both more abundant and of higher temperature.

Electrical power is presently being generated from Roosevelt Hot Springs (20 MWe) and Cove Fort/Sulphurdale (3.2 MWe) in Utah; from Beowawe (17 MWe), Desert Peak (9 MWe), Wabuska (0.6 MWe), and Steamboat Springs (5.4 MWe) in Nevada; and from Coso Hot Springs (30 MWe) in California. Exploration is being or has been conducted at probably 20 or more sites. Direct application of geothermal energy for industrial process heating and space heating are currently operating in this area at several sites including Brady Hot Springs (vegetable drying), Reno (space heating) and Salt Lake City (greenhouse heating).

The reasons for the abundance of resources in the Basin and Range seem clear. This area, especially at its margins, is an active area geologically. Volcanism only a few hundred years old is known from tens of areas. The area is also active seismically and faulting that causes the uplift of mountain ranges also serves to keep pathways open for deep fluid circulation.

As an example of a Basin and Range hydrothermal system, we will discuss Roosevelt Hot Springs, although it should not be supposed to be typical of all high-temperature occurrences in this province. The oldest rocks exposed (Figures 19 and 20) are Precambrian sedimentary rocks that have been extensively metamorphosed. These rocks were intruded during Miocene time (7-23 mybp) by granitic rocks (33) and (34). Rhyolite volcanic flows and domes were emplaced during the interval 800,000 to 500,000 years ago. The area has been complexly faulted by north- to northwest-trending high-angle faults and by east- west high-angle faults. The Negro Mag fault is such an east-west fault that is an important controlling structure in the north portion of the field. The north- trending Opal Mound fault apparently forms the western limit of the system. The oldest fault system is a series of low-angle denudation faults (Figure 20) along which the upper plate has moved west by about 600 m and has broken into a series of discrete blocks. Producing areas in the southern portion of the field are located in zones of intersection of the upper-plate faults with the Opal Mound and other parallel faults. Producing zones in the northern part of the region are located at the intersection of north-south and east-west faults. The permeability is obviously fracture controlled.

Cascade Range and Vicinity

The Cascade Range of northern California, Oregon, Washington and British Columbia is comprised of a series of volcanos, 12 of which have been active in historic times. The May 18, 1980 eruption of Mount St. Helens attests to be the youth of volcanic activity here. The Cascade Range lies above the zone of subduction of the Juan de Fuca plate beneath the North American plate, (Figure 2) and magma moving into the upper crust has transported large amounts of heat upward. In spite of the widespread, young volcanism, however, geothermal manifestations are not as plentiful as expected. High rainfall and snowfall in the Cascades are believed to suppress surface geothermal manifestations through downward percolation of the cold surface waters in the highly permeable volcanic rocks. In the absence of surface manifestation, discovery becomes much more difficult.

No producible high-temperature hydrothermal systems have yet been located in the Cascades. A vapor-dominated system is present at Lassen Peak in California, but it lies within a national park, and will not be developed. A hydrothermal system having temperatures greater than 200 deg C has been located at Newberry Caldera in Oregon through research drilling sponsored by the U. S. Geological Survey (35), but the known portion of the system lies within the caldera and will not be exploited for environmental reasons.

Industry's exploration efforts have increased somewhat in the last several years. The Department of Energy is currently sponsoring a cost-shared drilling program with industry to encourage more subsurface exploration and to help develop research data for devising new exploration techniques. To date, two holes have been drilled at Newberry volcano by GEO Operator Corporation, and one hole has been completed by Thermal Power Company north of Mt. Jefferson. A third research hole has been started on the southeast slope of Mt. Mazama, the volcano whose summit consists of the Crater Lake caldera. This hole has found interesting temperatures at shallow depths (+100 deg C at 1300 feet), but the hole remains unfinished at this writing.

The use of geothermal energy for space heating at Klamath Falls, Oregon is well established (36), and numerous hot springs and wells occur throughout the Cascades. Potential for discovery of resources in all temperature categories is great (37).

Snake River Plain

The basalt flows and other volcanic deposits of the Snake River Plain are an extension of the Columbia Plateau eastward across southern Idaho to the border with Wyoming. The Plain is divided into a western part and an eastern part. Thermal waters occur in numerous wells and springs in the western portion, especially on or near the edges of the plain. Geochemically indicated resource temperatures exceed 150 deg C at Neal Hot Springs and Vale, Oregon and Crane Creek, Idaho, but indicated temperatures for most resources are lower. Younger volcanic rocks occur in the eastern part of the plain, but no high-temperature resources are yet identified. This part of the plain is underlain by a high-flow, cold-water aquifer that is believed to mask surface geothermal indications.

The ages of volcanic eruptions decrease from west to east along the Snake River Plain, apparently reflecting the arcuate track of a mantle plume as the North American plate moved westward. Recent volcanic activity has taken place at Yellowstone, under which the hot spot currently lies. Future violent eruptions in the area are possible. The vapor- and water-dominated hydrothermal systems at Yellowstone will not be developed because they lie within a national park, but surrounding areas are highly prospective.

Direct use of hydrothermal energy for space heating is famous at Boise, Idaho, where the Warm Springs district has been heating homes geothermally for almost 100 years (38). Also, near this area, but lying in the Basin and Range, is the Raft River site where the Idaho National Engineering Laboratory of DOE constructed and operated a 5 MWe binary demonstration plant on a hydrothermal resource whose temperature is 147 deg C. This project is currently inoperative and the plant has been sold.

Rio Grande Rift

The Rio Grande Rift is a north-trending tectonic feature that extends from Mexico through central New Mexico and ends in central Colorado. It is a down-dropped area that has been filled with volcanic rocks and erosional debris from the bordering plateaus and mountains. The rift began to form in late Oliogocene times (23-38 mybp), and volcanic and seismic activity have occurred subsequently to the present.

There are several low- and intermediate-temperature hydrothermal convection systems in this area, but the only high-temperature system that has been drill tested to any significant extent and where production is proven is a hot water-dominated system in the Valles caldera (39) and (40). Deep drilling has encountered a hydrothermal convection system in fractured Tertiary volcanic, Paleozoic sedimentary and Precambrian granitic rocks at an average depth of 2 to 3 km. Temperatures as high as 300 deg C have been recorded. An attempt by DOE, Union Geothermal and Public Service Company of New Mexico to build a demonstration plant at that location failed when the steam supply proved to be inadequate. Recent research drilling, sponsored by DOE under the Continental Scientific Drilling Program, has developed an improved understanding of the area. Geologists believe that the area contains an important, undiscovered hydrothermal resource capable of electrical power generation. Also located near the caldera is the site of Los Alamos National Laboratory's DOE sponsored hot dry rock experiment at Fenton Hill.

Madison and other Aquifers

Underlying a large area in western North and South Dakota, eastern Montana and northeastern Wyoming are a number of aquifers that contain thermal waters. These aquifers have developed in carbonates and sandstones of Paleozoic and Mesozoic age. The permeability is both intergranular and fracture controlled in the case of the sandstones (e.g. the Dakota Sandstone) and fracture and solution cavities in the carbonates (e.g. the Madison Limestone). Some of the aquifers produce under artesian pressure. Depths to production vary widely but average perhaps 2,000 ft. Temperatures are 30 to 80 deg C (41) in the Madison but are lower in other shallower aquifers such as the Dakota. Direct use of the thermal water is being made at a few locations today (42), and it is evident that the potential for further development is substantial.

Balcones Zone, TX

Thermal waters at temperatures generally below 60 deg C occur in a zone that trends northeasterly across central Texas. Many of the large population centers are in or near this zone, and there appears to be significant potential for geothermal development in spite of the rather low temperatures.

An initial assessment of the geothermal potential has been documented by Woodruff and McBride (43). The thermal waters occur in a band broadly delimited by the Balcones fault zone on the west and the Luling-Mexia-Talco fault zone on the east. In many locations the thermal waters are low enough in content of dissolved salts to be potable, and indeed many communities already tap the warm waters for their municipal water supplies.

The geothermal aquifers are mostly Cretaceous (65-140 mybp) Sandstone units, although locally thermal waters are provided from Cretaceous limestones and Tertiary sandstones. The thermally anomalous zone coincides with an ancient zone of structural weakness dating back more than 200 million years. The zone has been a hinge line with uplift of mountain ranges to the north and west and downwarping to the south and east. Sediments have deposited in the area of downwarping, and the rate of sedimentation has kept pace with sinking, keeping this area close to sea level. Structural deformation of the sediments, including faulting and folding, and interfingering of diverse sedimentary units have resulted in the complex aquifer system of today. The source of the anomalous heat is not known with certainty.

Hawaiian Islands

The chain of islands known as the Hawaiian archipelago stretches 2,500 km in a northwest-southeast line across the Pacific Ocean from Kure and Midway Islands to the Big Island of Hawaii. Built of basaltic volcanic rocks, this island chain boasts the greatest volcanic masses on earth. The volcano Kilauea rises 9,800 m above the floor of the ocean, the world's largest mountain in terms of elevation above its base. The Kilauea, Mauna Loa and other vents on the big island are in an almost continual state of eruptive activity, but by contrast volcanos on the other islands have shown little recent activity. Haleakala on the island of Maui is the only other volcano in the state that has erupted in the last few hundred years, and the last eruption there was in 1790 (44).

Several of the Hawaiian islands are believed to have geothermal potential. The only area where exploration has proceeded far enough to establish the existence of a hydrothermal reservoir is in the Puna district on the Big Island near Kapoho along the so-called "East Rift", a fault zone on the east flank of Kileaua. Here a well was completed to a depth of 1965 m (45) with a bottom- hole temperature of 358 deg C. A 3 MWe generator is currently being operated at the site. Exploration is underway by several companies in areas adjacent to the operating plant. Elsewhere on the islands, potential for occurrence of low- to moderate-temperature resources has been established at a number of locations on Hawaii, Maui and Oahu, although little drilling to prove resources has been completed (46).

Alaska

Little geothermal exploration work has been done in Alaska. A number of geothermal occurrences are located on the Alaska Peninsula and the Aleutian Islands and in central and southeast Alaska. The Aleutians and the Peninsula overlie a zone of active subduction (Figure 2), and volcanos are numerous. A hydrothermal system was located at Makushin volcano on the island of Unalaska (47) and the island of Adak is also believed to have good discovery potential.

Low- and moderate-temperature resources are indicated in a number of locations in Alaska by occurrence of hot springs (22). One area that has been studied in more detail and has had limited drilling is Pilgrim Hot Springs (48). This site is 75 km north of Nome, Alaska. Initial drilling has confirmed the presence of a hot water reservoir about 1 sq km in extent that has artesian flow rates of 200-400 gallons per minute of 90 deg C water.

POTENTIAL FOR GEOTHERMAL DEVELOPMENT IN THE U.S.

Muffler et al. (22) have dealt with the problem of how much accessible resource exists in the U. S. both at known sites and those that are undiscovered. They conclude that about 1650 EH8J of energy are present in reservoirs of 215 identified hydrothermal systems in the U.S. having temperatures greater than 90 deg C and excluding national parks. Recoverable thermal energy at the surface from these systems is estimated to be 400 EH8J, which is sufficient to produce 23,000 megawatts of electricity for 30 years and to produce 42 EH8J of direct heat. The undiscovered hydrothermal resource base is estimated to be about five times greater than the known resources. These figures do not include possible hot dry rock or other more speculative resources. Table 4 is a summary of the current estimate of the geothermal resource base as taken from Muffler et al. (22). This table demonstrates our lack of resource knowledge through the ranges and relative amounts of undiscovered resources and through the many missing numbers. We can conclude, however, that the geothermal resource base is large in the U.S.

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Table 1

Geothermal Resource Classification

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

Table 3

Geologic Time Scale

(millions of years)
(from van Eysinga, 1978)

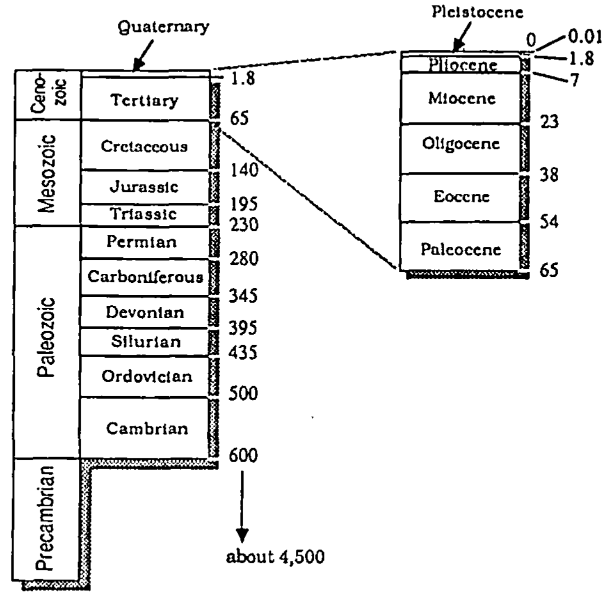


Table 2

Representative Analyses of Geothermal Fluids

Sample #		1	2	3	4	5	6
Temp. °C		42	89	255	<260	292	316
pH			7.9	8.4			
SiO_2	(ppm)	52	293	690	563	705	400
Ca	(ppm)	257	5	17	8	592	28,000
Mg	(ppm)	17	.8	.03	<2	.6	54
Na	(ppm)	578	653	1,320	2,320	6,382	50,400
K	(ppm)		71	255	461	1,551	17,500
Li	(ppm)	.5	.7	14.2	25.3	14.5	215
HCO	(ppm)		305		232	28	7,150
SO	(ppm)	932		36	72	<3.5	5
Cl	(ppm)	625	865	2,260	3,860	11,918	155,000
F	(ppm)	2.8	1.8	8.3	6.8		15
B	(ppm)	2.6	4.9			13.4	390
As	(ppm)		2.7	4.8	4.3		12

Sample Descriptions:

1. Hot spring, Monroe, UT.
2. Hot spring, Steamboat, NV.
3. Well 44, Wairakei, New Zealand.
4. Brine discharged from well 54-3, Roosevelt Hot Springs, UT.
5. Analyses calculated from flashed brine, well M-26, Cerro Prieto, Mex.
6. Brine discharged from well 11D, Salton Sea Geothermal Field, CA.

Table 4

Geothermal Energy of the United States

After Muffler et al. (1978) Table 20

RESOURCE TYPE	ELECTRICITY (MWe for 30 yr)	BENEFICIAL HEAT (10^{18} joules)	RESOURCE (10^{18} joules)
Hydrothermal			
Identified	23,000	42	400
Undiscovered	72,000-127,000	184-310	2,000
Sedimentary Basins	?	?	?
Geopressured (N. Gulf of Mexico)			
Thermal			270-2800
Methane			160-1600
Radiogenic	?	?	?
Hot Rock	?	?	?

TEMPERATURE VS DEPTH IN EARTH

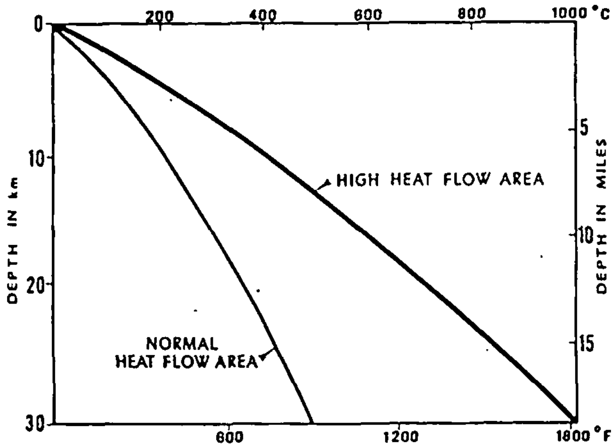


FIGURE 1

INTERIOR OF THE EARTH

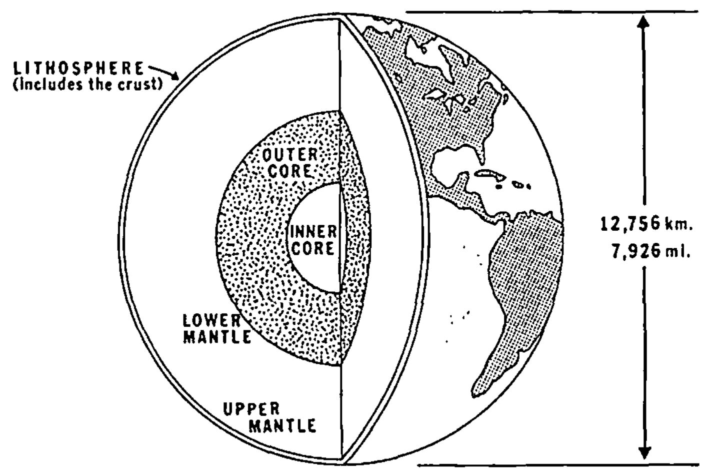
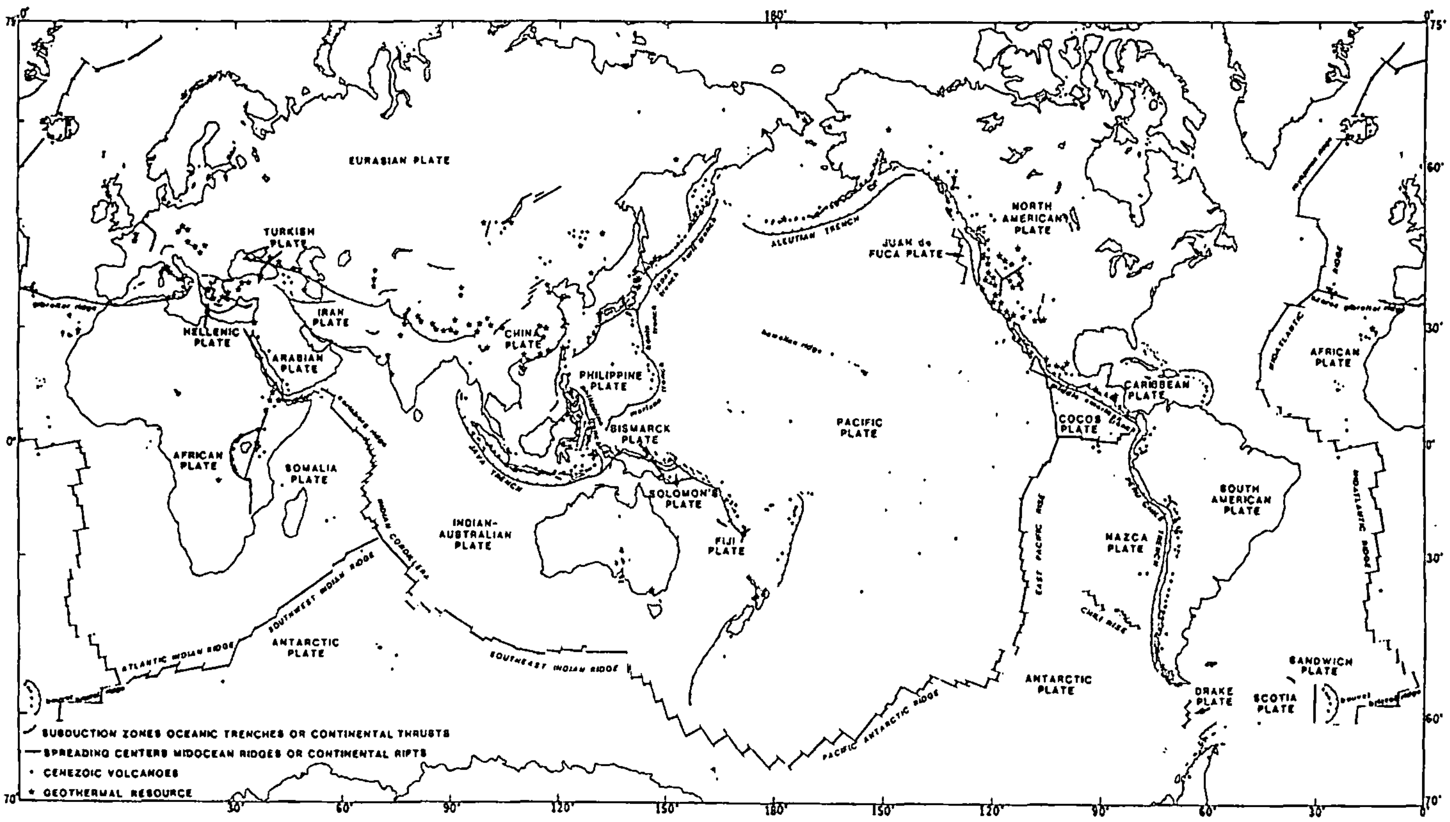


FIGURE 3



GEOTHERMAL RESOURCES AND PLATE TECTONIC FEATURES

FIGURE 2

CONCEPT OF PLATE TECTONICS

(NOT TO SCALE)

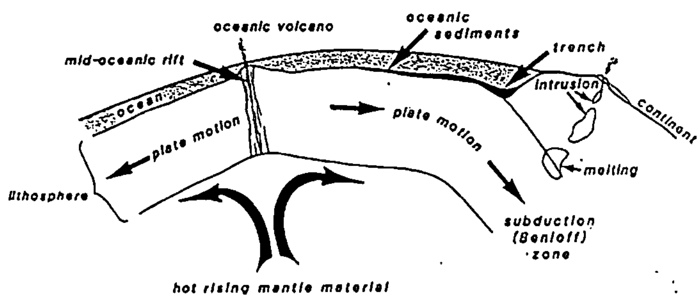


FIGURE 4

Geothermal Temperatures

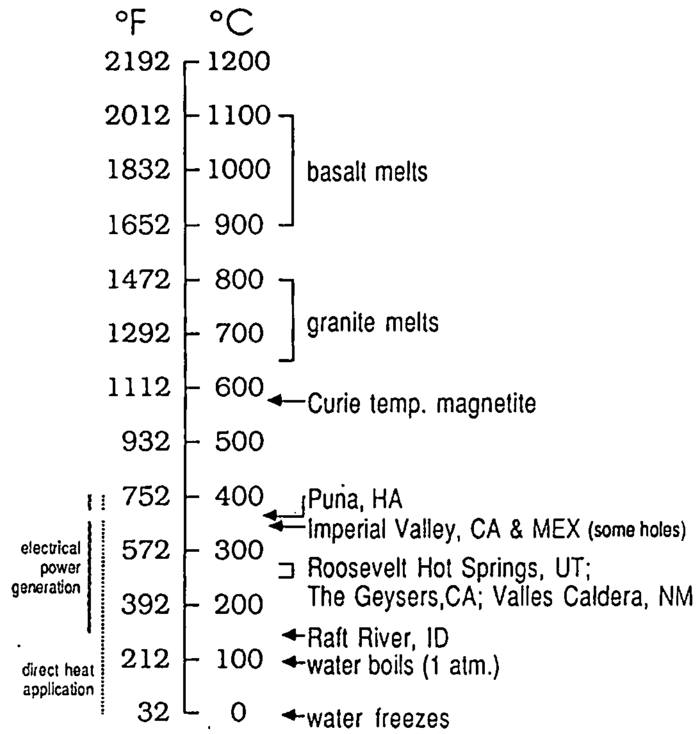


Figure 6

CRUSTAL INTRUSION

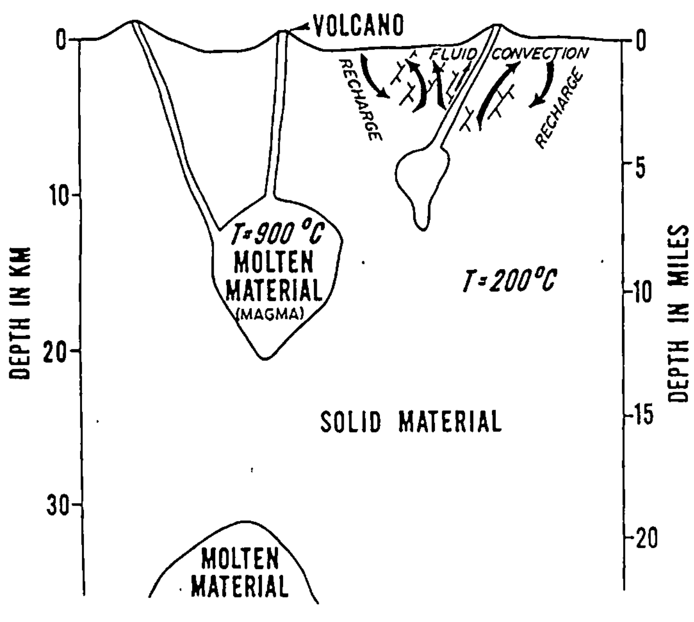


FIGURE 5

VAPOR DOMINATED GEOTHERMAL RESERVOIR

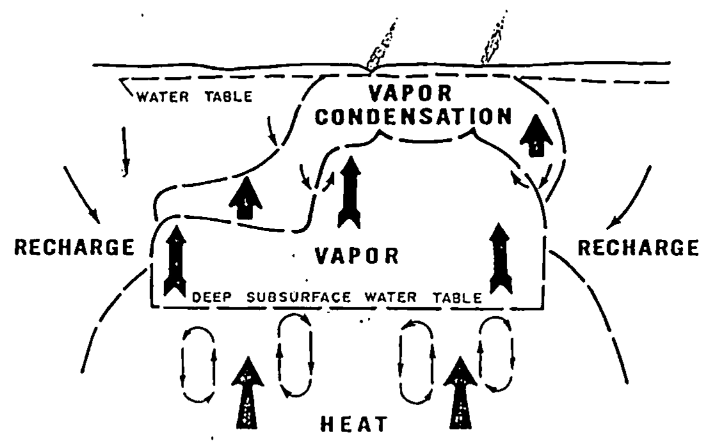


FIGURE 7

HYDROTHERMAL SYSTEM IN VOLCANIC TERRANE

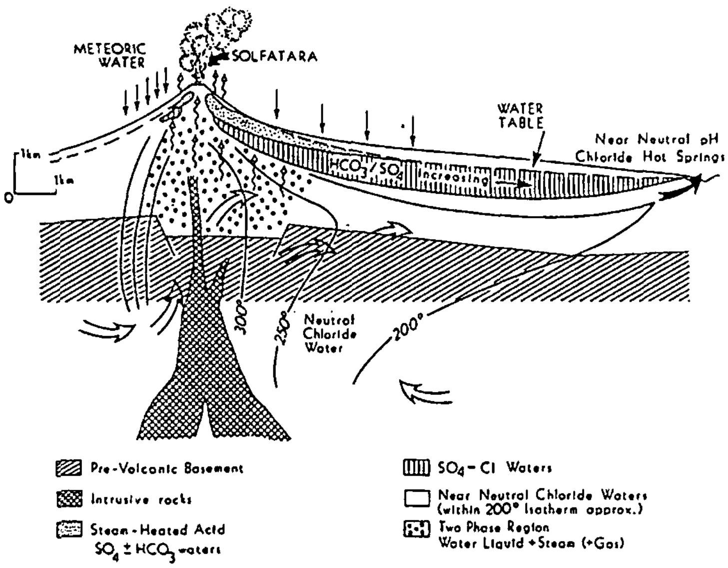
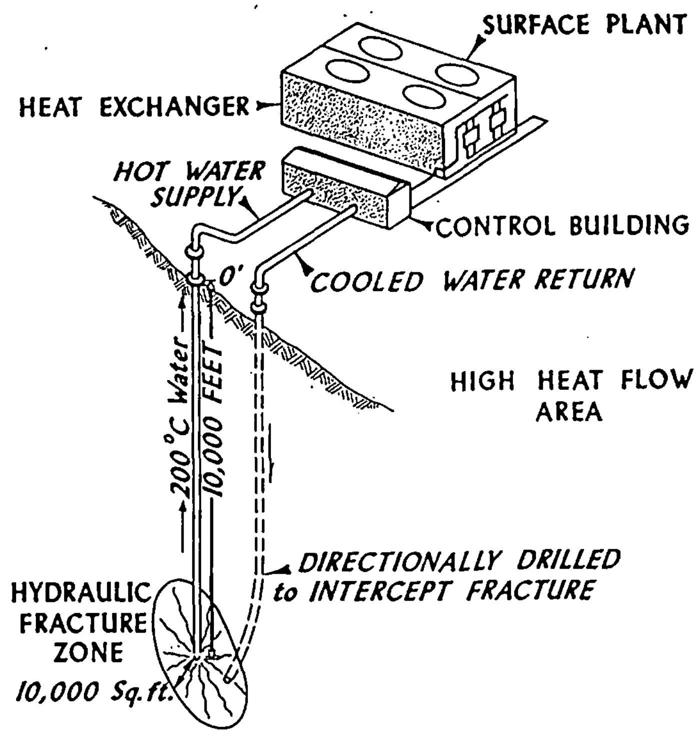


FIGURE 8



HOT DRY ROCK GEOTHERMAL RESOURCE

FIGURE 10

GEOPRESSURED GEOTHERMAL RESOURCES

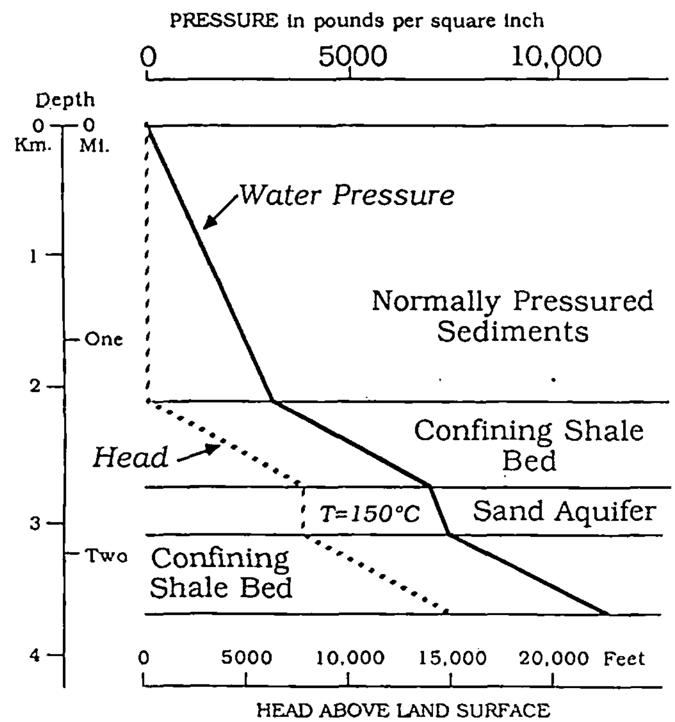


FIGURE 9

RADIOGENIC GEOTHERMAL RESOURCE

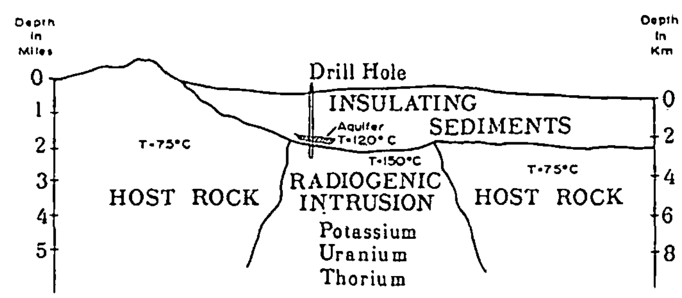


FIGURE 11

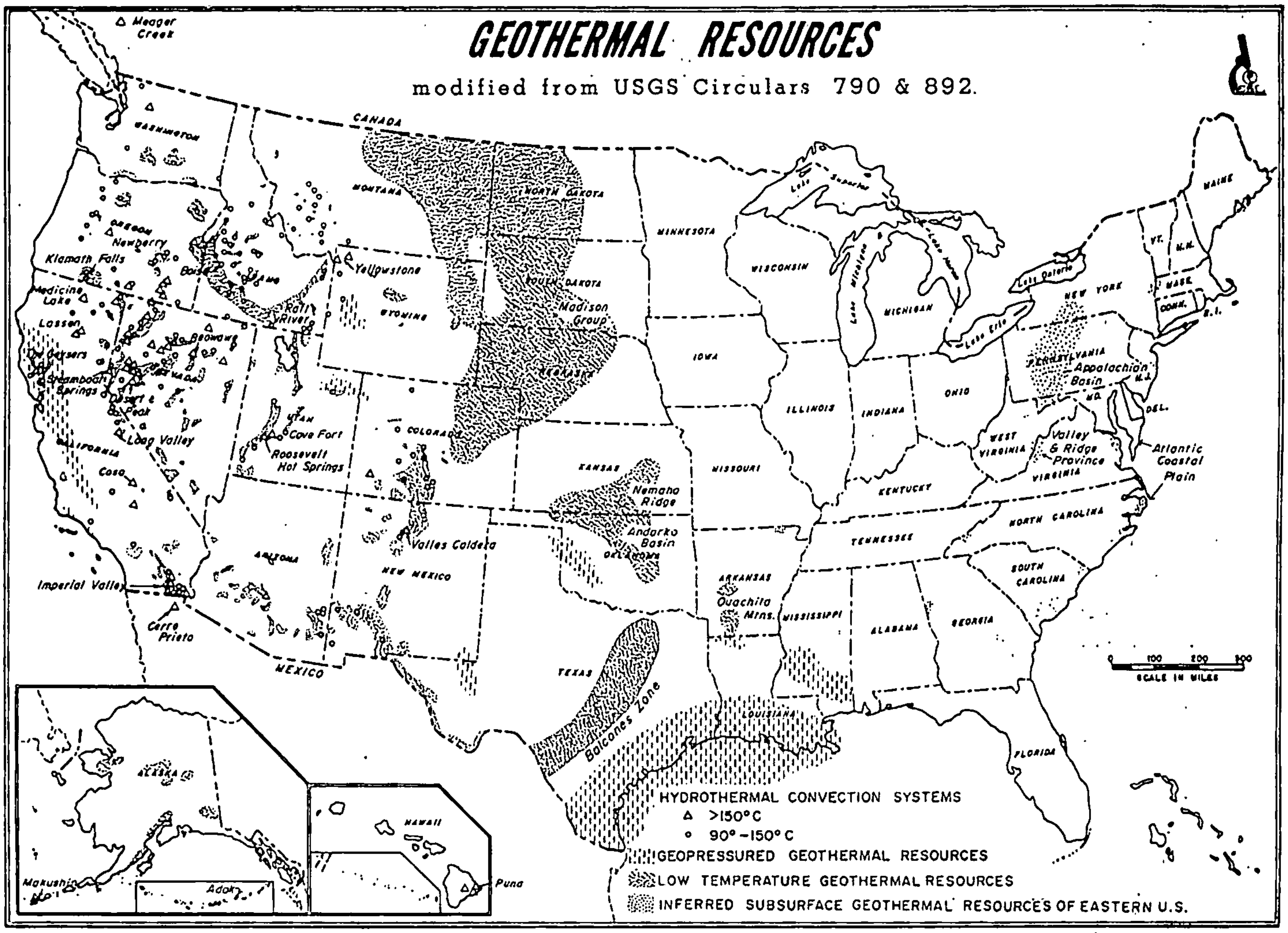
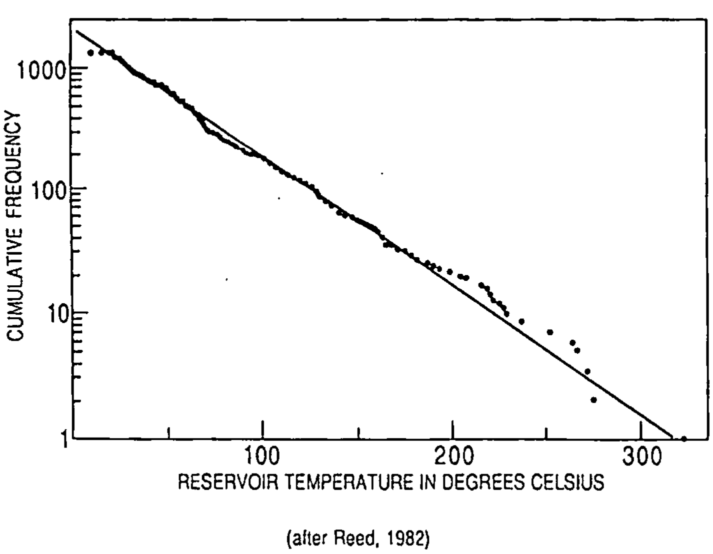
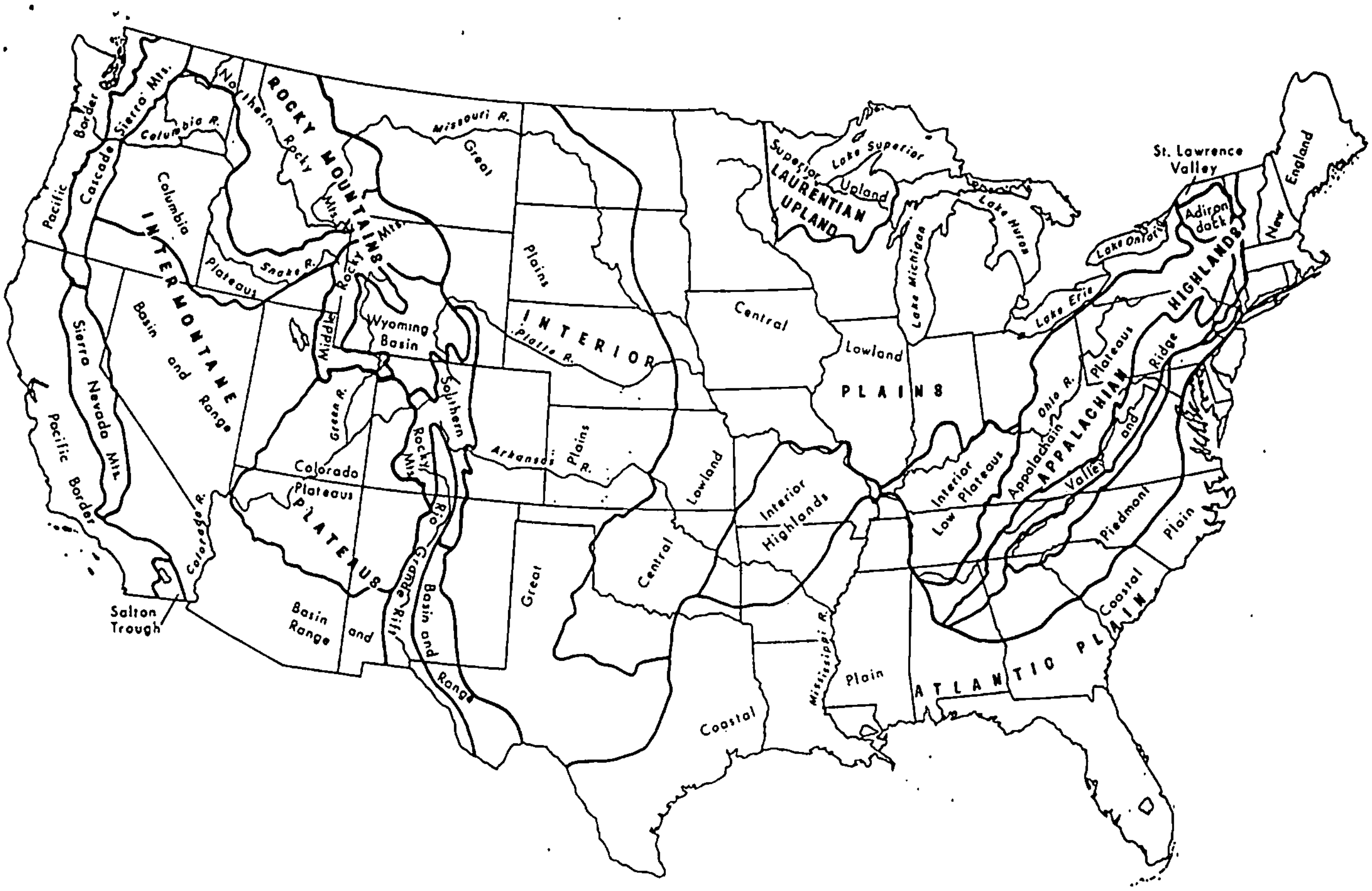


FIGURE 12



(after Reed, 1982)

FIGURE 13



PHYSIOGRAPHIC MAP of USA

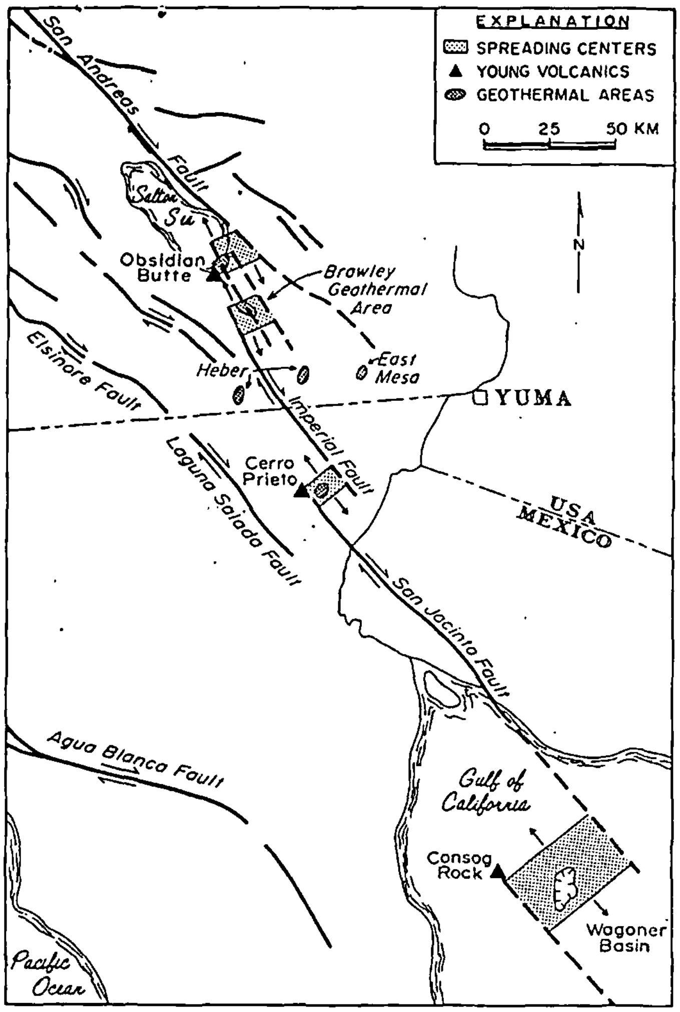
(after Fenneman, 1928)

SCALE

0 100 200 300 400 MILES

0 200 400 600 KILOMETERS

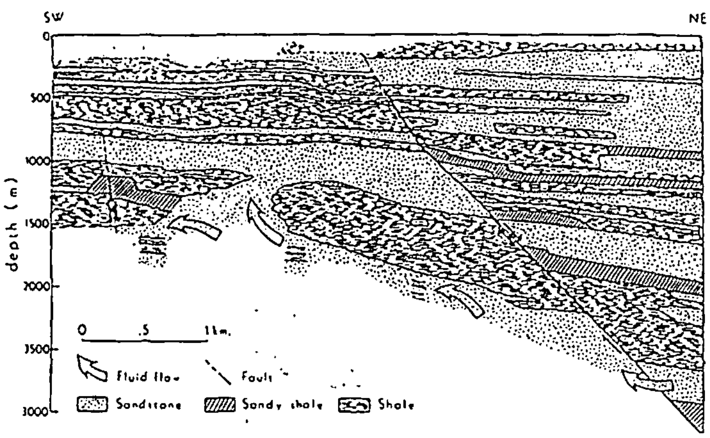
FIGURE 14



MAJOR STRUCTURES OF SALTON TROUGH

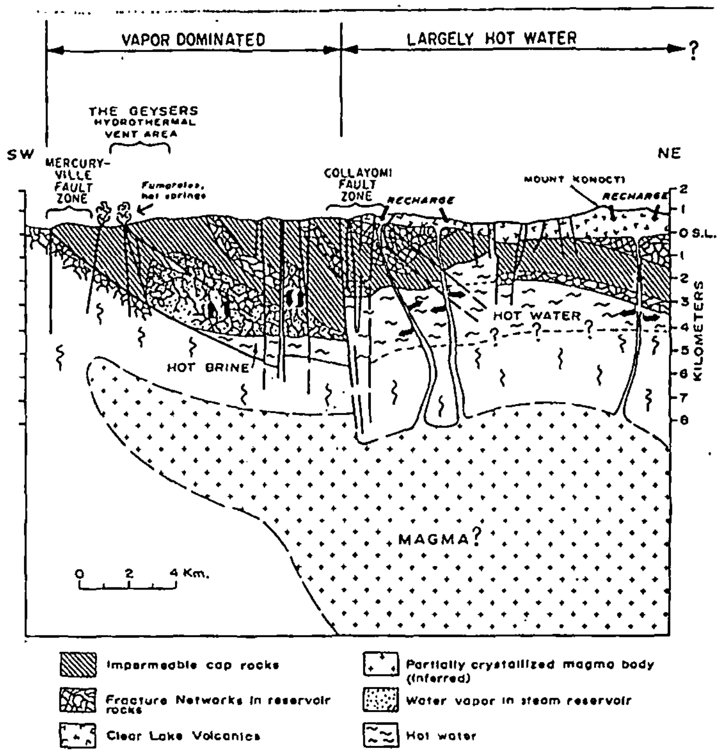
(after Palmer et al., 1975)

FIGURE 15



FLUID FLOW MODEL OF CERRO PRIETO, MEXICO

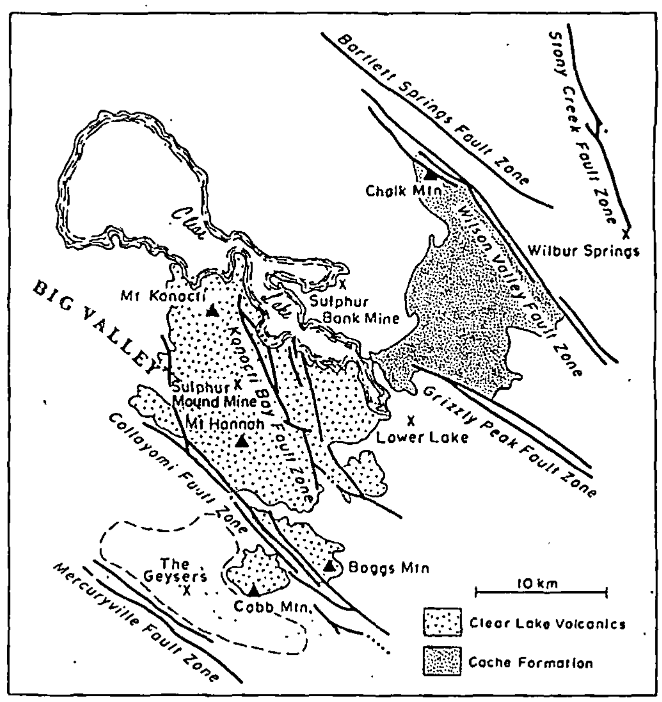
FIGURE 16



CRUSTAL MODEL FOR THE GEYSERS - CLEAR LAKE AREA, CA.

(after McLaughlin, 1977)

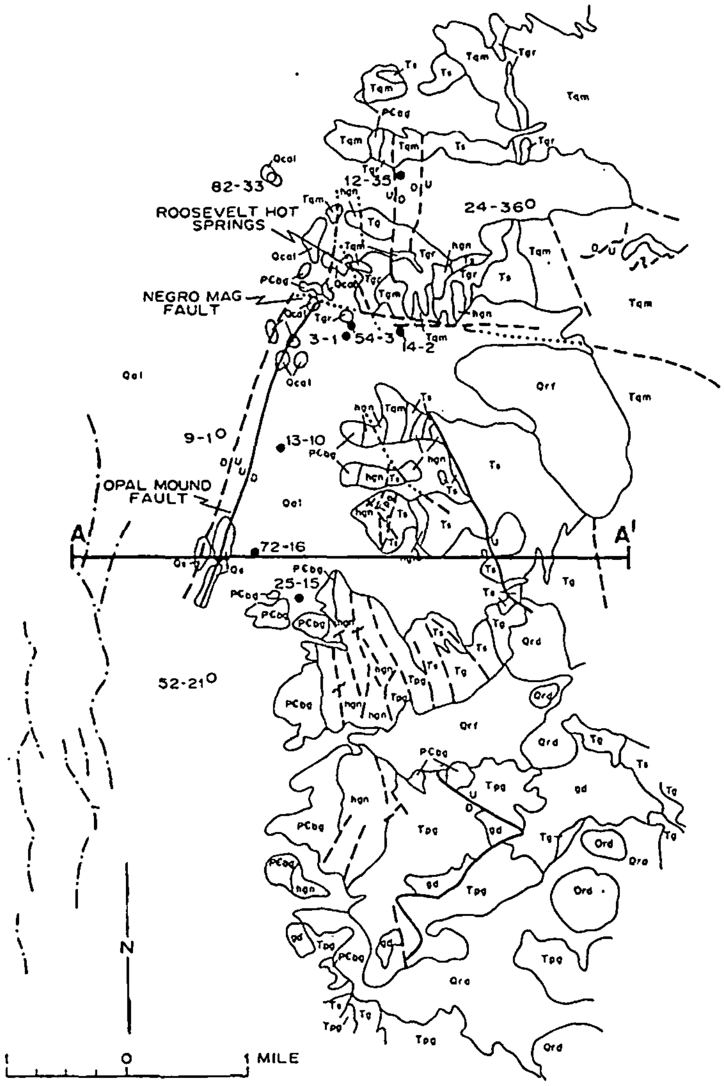
FIGURE 17



MAJOR STRUCTURES in
THE GEYSERS-CLEAR LAKE AREA

(After Goff, 1980)

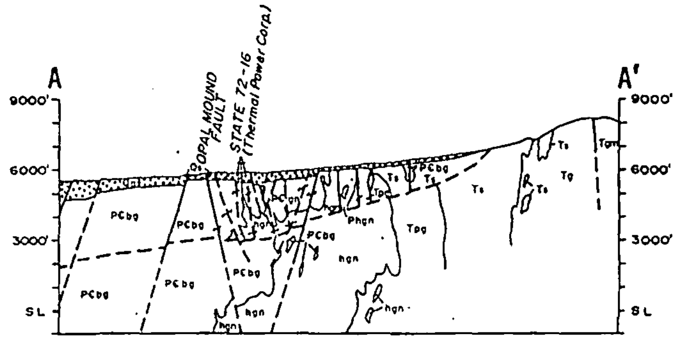
FIGURE 18



GEOLOGIC MAP
ROOSEVELT HOT SPRINGS, UTAH
(from Nielson et al., 1978)

FIGURE 19

GEOLOGIC CROSS SECTION
ROOSEVELT HOT SPRINGS, UTAH
(from Nielson et al., 1978)



EXPLANATION

Qal	- alluvium	Tg	- granite
Qcal	- silicified alluvium	Ts	- syenite
Qs	- siliceous sinter	Tpg	- porphyritic granite
Qrd	- rhyolite domes	Tqm	- quartz monzonite
Qra	- pyroclastic deposits	gd	- biotite diorite
Qrf	- rhyolite flows	hgn	- foliated hornblende granodiorite
Tgr	- fine-grained granite	PЄbg	- banded gneiss

FIGURE 20

NATURE AND OCCURRENCE OF GEOTHERMAL RESOURCES IN THE UNITED STATES

Phillip M. Wright

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

INTRODUCTION

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 paper is to present an overview of the geology of geothermal resources. It was written specifically with the non-geologist in mind. The use of highly technical geological language is avoided where possible, and the terms that are used are also defined. Emphasis is on resources in the United States, but the geological principles discussed have world-wide application. We will see that geothermal resources of high temperature are found mainly in areas where a number of specific geologic processes are active today and that resources of lower temperature are more widespread. We will present a classification for observed resource types and briefly describe the geology of each type. The geology of the United States will then be summarized to provide an appropriate background for consideration of the occurrence of geothermal resources. Finally we will be able to reach the conclusion that the accessible geothermal resource base in the United States is very large and that the extent of development over the next decades will be limited by economics rather than by availability.

THE EARTH'S INTERNAL HEAT

Although the distribution with depth in the earth of density, pressure and other related physical parameters is 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. Using present technology and under good conditions, holes can be drilled to depths of about 10 km, where temperatures range upward from about 150°C in areas underlain by cooler rocks to perhaps 600°C in exceptional areas.

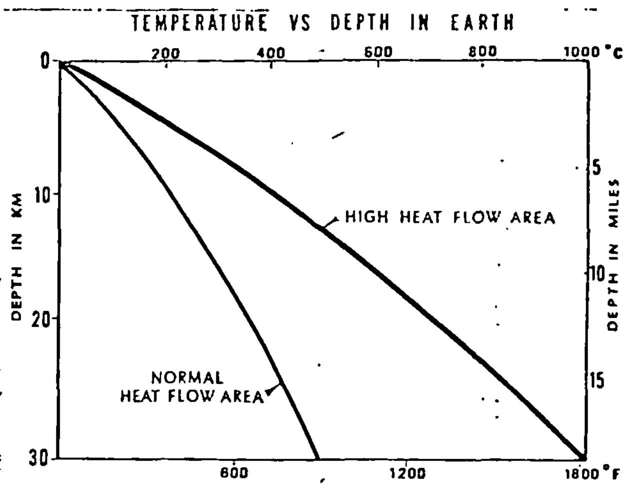


Figure 1

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 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.

INTERIOR OF THE EARTH

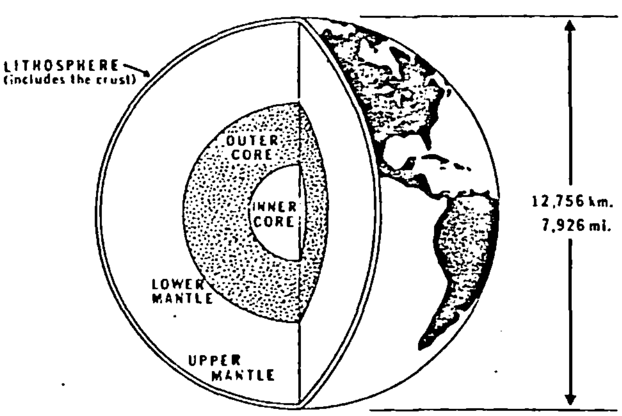


Figure 2

GEOLOGICAL PROCESSES

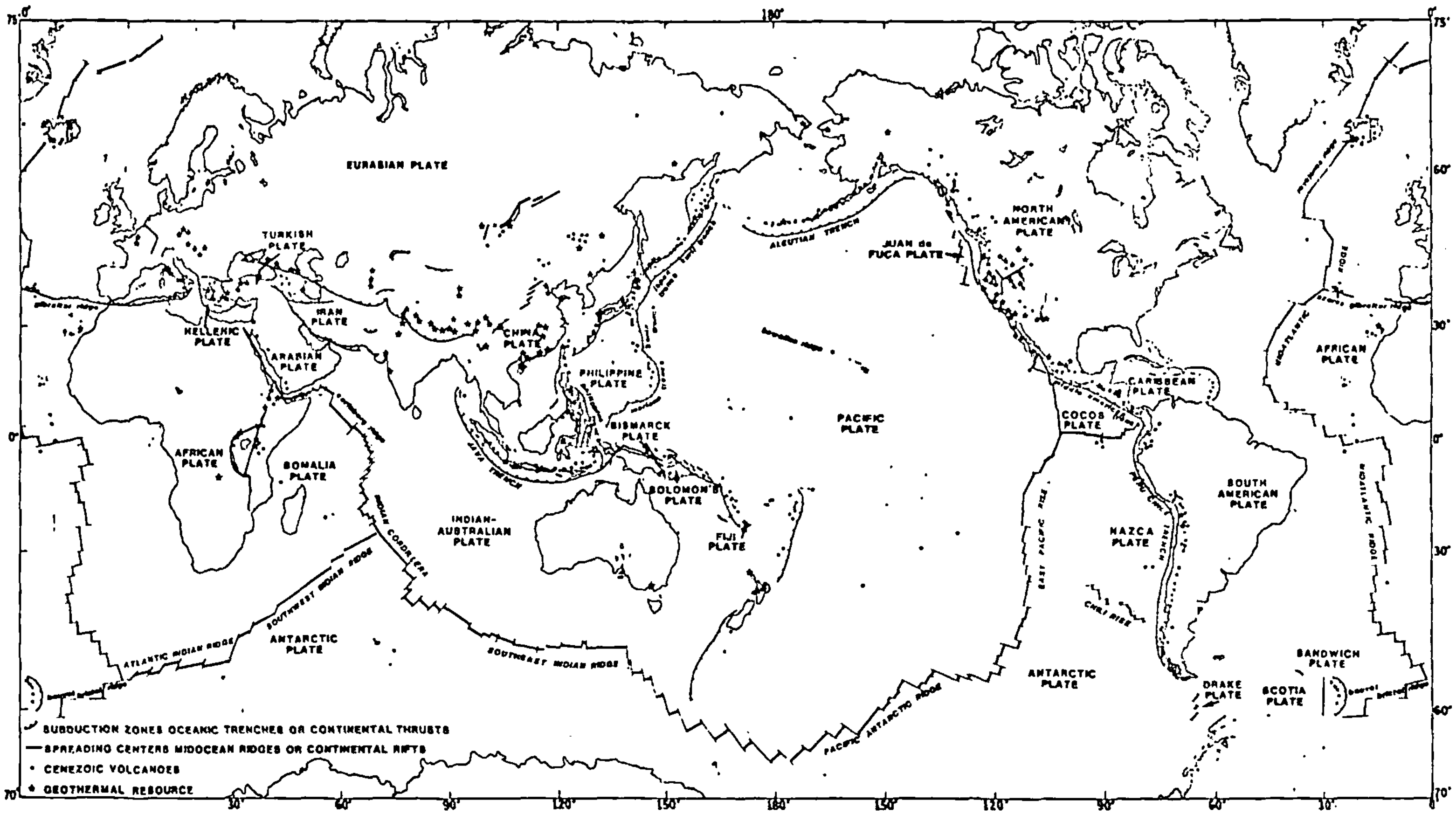
Geothermal resource areas, or geothermal areas for short, are generally those in which higher temperatures are found at shallower depths than is normal. This condition usually results from either 1) intrusion of molten rock to high levels in the earth's crust, 2) higher-than-average flow of heat to the surface with an attendant high rate of increase of temperature with depth (geothermal gradient) as illustrated in Figure 1, often in broad areas where the earth's crust is thin, 3) heating of ground water that circulates to depths of 2 to 5 km with subsequent ascent of the thermal water near to the surface, or 4) anomalous heating of a shallow rock body by decay of an unusually high content of radioactive elements. We will consider each of these phenomena in more detail below.

In many geothermal areas heat is brought right to the surface by circulation of ground water. If temperature is high enough, steam may be produced, and geysers, fumaroles, and hot springs are common surface manifestations of underlying geothermal reservoirs.

The distribution of geothermal areas on the earth's surface is not random but instead is governed by geological processes of global and local scale. 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 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



GEOTHERMAL RESOURCES AND PLATE TECTONIC FEATURES

3

Wright

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 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

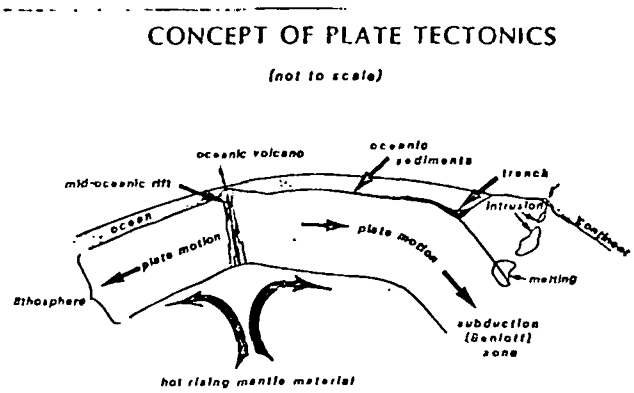


Figure 4

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 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.

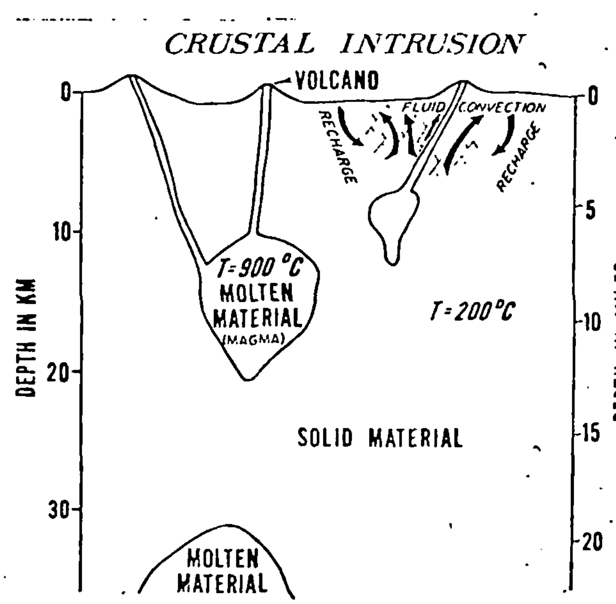


Figure 5

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 (Fig. 16) 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.

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

We have seen that the fundamental cause of many geothermal resources lies in the transport of heat near to the surface through one or more of a number of geological processes. We have also seen that the ultimate source of that heat is in the interior of the earth where temperatures are much higher than they are at the surface. We will now turn to an examination of various 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 with emphasis on those that are presently nearest to commercial use in the U.S 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.

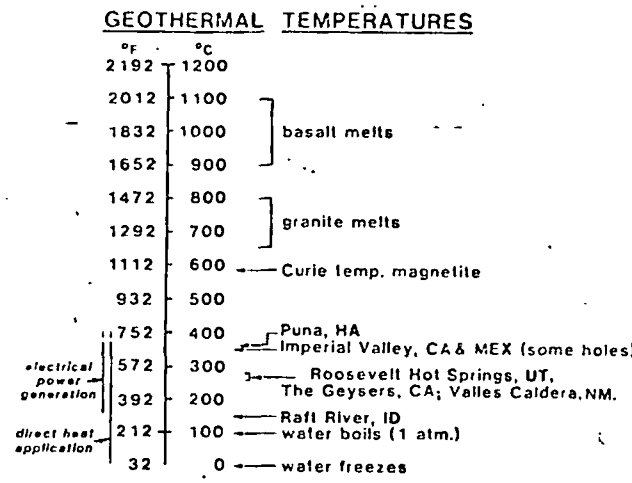


Figure 6

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

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 (Fig. 6) 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.

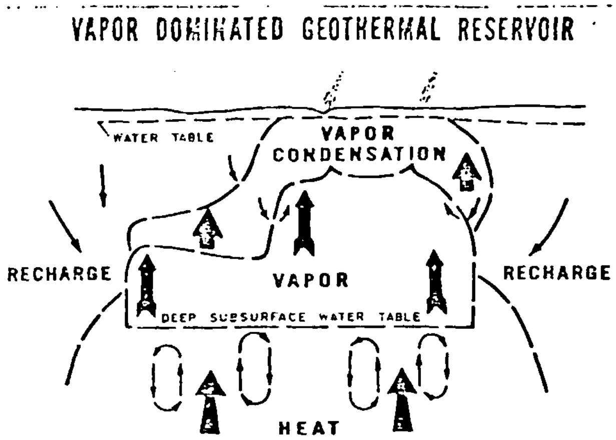


Figure 7

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 I). 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.

A well drilled into a vapor-dominated reservoir would produce superheated steam. The Geysers geothermal area in California (see Fig. 17 and the discussion below) 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 908 MWe (megawatts of electrical power, where 1 megawatt = 1 million watts), and 880 MWe of additional generating capacity is scheduled to come on line by 1986.

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 by industry 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 I). 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 geothermal 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.

WATER DOMINATED GEOTHERMAL SYSTEM

FLOW CONTROLLED BY FRACTURES

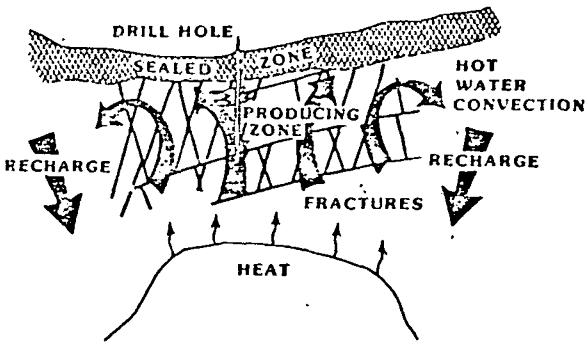


Figure 8

Examples of this type of geothermal resource are abundant in the western U.S. and include Roosevelt Hot Springs, Utah, and the Valles Caldera area, New Mexico. Approximately 50 areas having potential for containing such a resource have been identified (Muffler and others, 1978) so far in the West, with Nevada having a disproportionately large share.

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, in such areas as East Mesa, Heber, Brawley, the Salton Sea, and at Cerro Prieto, 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 specific resource areas appear to be controlled by faults that presumably allow deep fluid circulation to carry the heat upward to reservoir depths.

IMPERIAL VALLEY, CALIFORNIA GEOTHERMAL RESOURCE

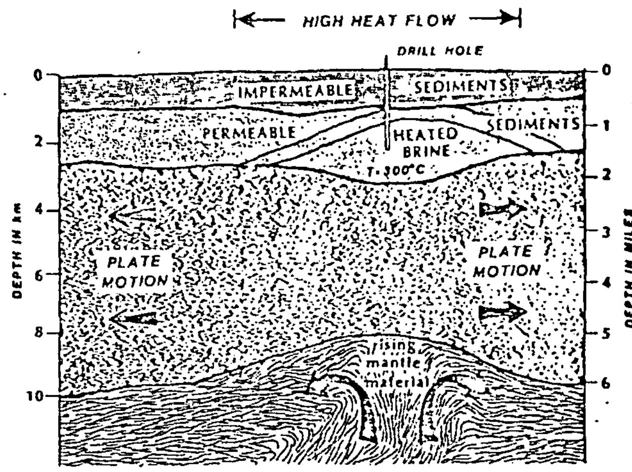
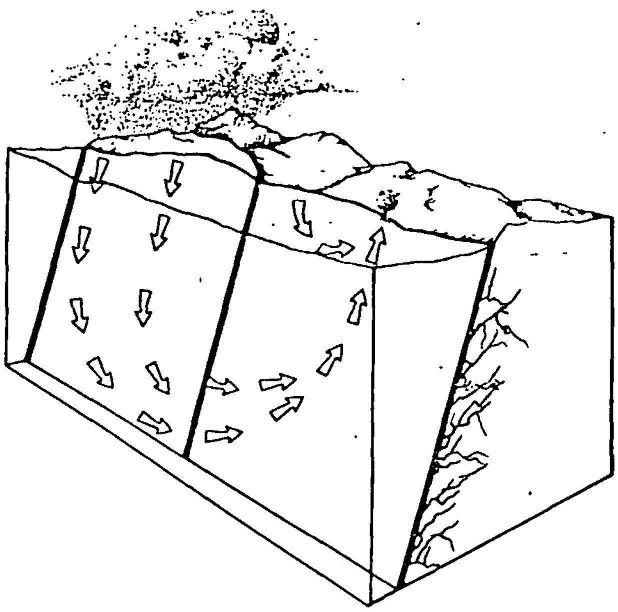


Figure 9

Virtually all of industry's geothermal exploration effort in the United States is presently directed at locating vapor- or water-dominated hydrothermal systems of the types described above having temperatures above 200°C. A few of these resources are capable of commercial electrical power generation today. Current surface exploration techniques are generally conceded to be inadequate for discovery and assessment of these resources at a fast enough pace to satisfy the reliance the U.S. may ultimately put upon them for alternative energy sources. Development of better and more cost-effective techniques is badly needed.

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. This type of warm water resource is especially prevalent in the western U.S. where active faulting keep conduits open to depth.



MODEL OF DEEP CIRCULATION HYDROTHERMAL RESOURCE

Figure 10

Sedimentary Basins

Some basins are filled to depths of 10 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). The Madison group carbonate rock sequence of widespread occurrence in North and South Dakota, Wyoming, Montana, and northward into Canada contains warm waters that are currently being tapped by drill holes in a few places for space heating and agricultural purposes. In a similar application, 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 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 11 (from Papadopulos, 1975) gives a few typical parameters for geopressured reservoirs and illustrates the origin of the above-normal fluid pressure. These geopressured fluids, found mainly in the Gulf Coast of the U.S. (Fig. 17), generally 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.

Industry has a great deal of interest in development of geopressured resources, although they are not yet economic. The U.S. Department of Energy (DOE), Division of Geothermal Energy, is currently sponsoring development of appropriate exploitation technology.

GEOPRESSURED GEOTHERMAL RESOURCE

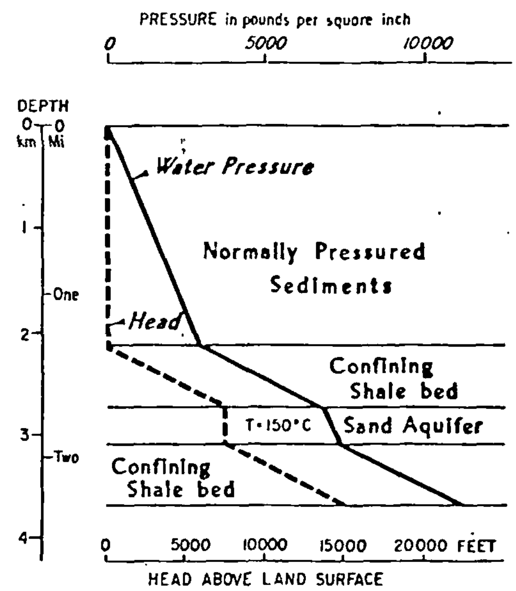


Figure 11

Radiogenic Resources

Research that could lead to development of radiogenic geothermal resources in the eastern U.S. (3c of Table 1) is currently underway following ideas developed at Virginia Polytechnic Institute and State University. The eastern states coastal plain is blanketed by a layer of thermally insulating sediments. In places beneath these sediments, rocks having enhanced heat-production due to higher content of radioactive elements are believed to occur. These rocks represent old intrusions of once molten material that have long since cooled and crystallized. 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. 12) is currently being completed under DOE funding. Success would most likely come in the form of low- to intermediate-temperature geothermal waters suitable for space heating and industrial processing. This could mean a great deal to the eastern U.S. where energy consumption is high and where no shallow, high-temperature hydrothermal convection systems are known. Geophysical and geological data indicate that radiogenically heated rock bodies may be reasonably widespread.

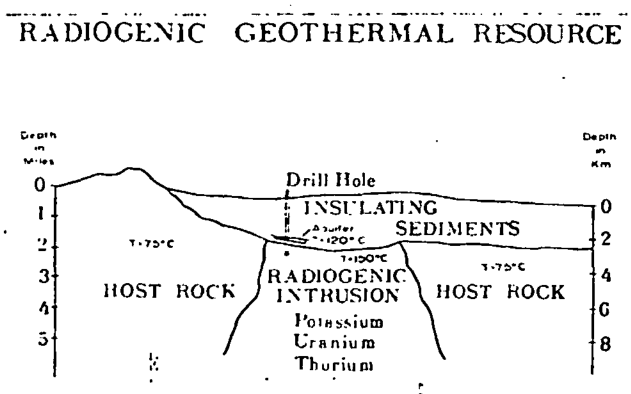
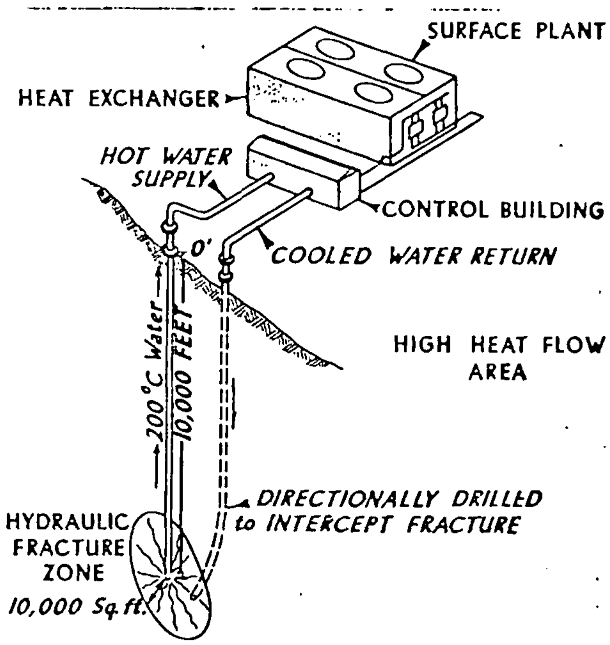


Figure 12

Hot Dry Rock Resources

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 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., 1975; 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. 13). 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.



HOT DRY ROCK GEOTHERMAL RESOURCE

Figure 13

Molten Rock

Experiments are underway at the 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.

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 (Hartley, 1980). Effective means have been and are still being developed to handle the scaling, corrosion and environmental problems caused by dissolved constituents in geothermal fluids.

GEOLOGY OF THE CONTINENTAL UNITED STATES

Before going on to a more detailed discussion of the occurrence of geothermal resources in the United States, let us turn to a summary of the geology of the U.S. This will form an appropriate context for consideration of the known and suspected geothermal occurrences.

Like all continental land masses, North America has had a long and eventful geologic history. The oldest rocks are dated at more than 2.5 billion years before present using radioactive dating methods. During this time the continent has grown through accretion of crustal material, mountain ranges have been uplifted and subsequently destroyed by erosion, blocks of rock have been displaced by faulting, both on a large scale as evidenced, for example, by the currently active San Andreas fault in California, and on the scale of an individual geothermal prospect, and volcanic activity has been widespread. In the discussion below some of these events will be described and will be keyed in time to the geological time scale, shown in Figure 14.

The U.S. can be divided into several distinct regions on the basis of geology. One way to do this is illustrated in Figure 15, which shows the major tectonic, or structural, divisions in the U.S. (Eardley, 1951). Areas of long-time stability are differentiated from areas of orogenic activity that has consisted of crustal downwarping accompanied by filling of basins with thick deposits of eroded sediments, mountain building with attendant faulting and folding of the rock strata, metamorphic changes of existing rocks by heat and pressure due to great depth of burial, intrusion of molten igneous rock bodies, some of great extent (batholiths), and eruption of volcanic rocks at the surface. A summary of these events, following Eardley (1951) closely will be given below for each of the tectonic divisions.

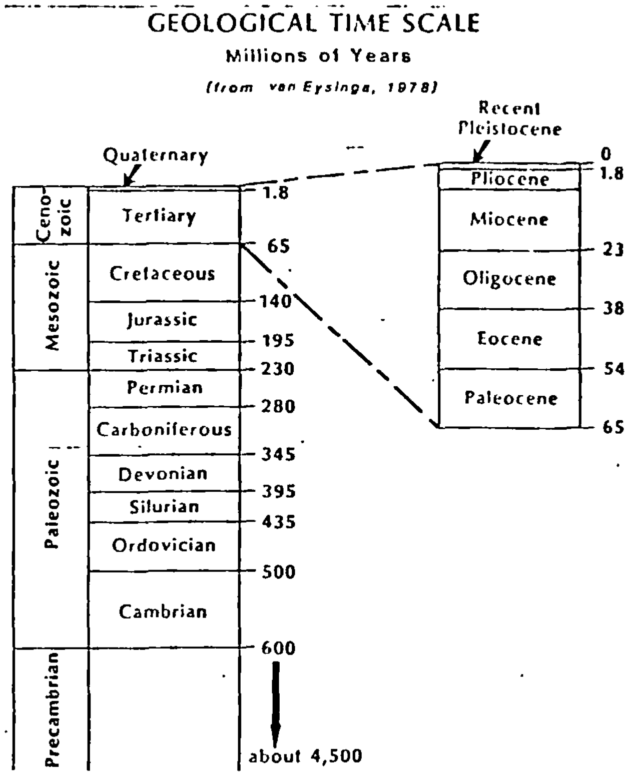


Figure 14

A second way to view the U.S. is in terms of present land forms or physiography as shown in Figure 16. This map will help the reader to correlate the discussion to follow with current names for various physiographic division. By reference to Figures 14, 15 and 16 this discussion will be more meaningful.

Canadian Shield

For the last billion years, the Canadian shield has been the great stable portion of the North American continent. It consists mainly of pre-Cambrian granitic intrusions and metamorphosed volcanic and sedimentary rock. A few occurrences of Paleozoic strata indicate that the Paleozoic formations were once much more widespread over the shield than now, and that they have been stripped off by a long interval of erosion during the Mesozoic and Cenozoic eras.

Central Stable Region

The central stable region consists of a foundation of pre-Cambrian crystalline rock, which is a continuation of the Canadian shield southward and westward, covered by a veneer of sedimentary sandstone, limestone and shale. The veneer varies greatly in thickness from place to place, and several broad basins, arches, and domes, developed chiefly in Paleozoic times, are present. Many of these basins have been the site of oil accumulation, and some contain aquifers having geothermal potential.

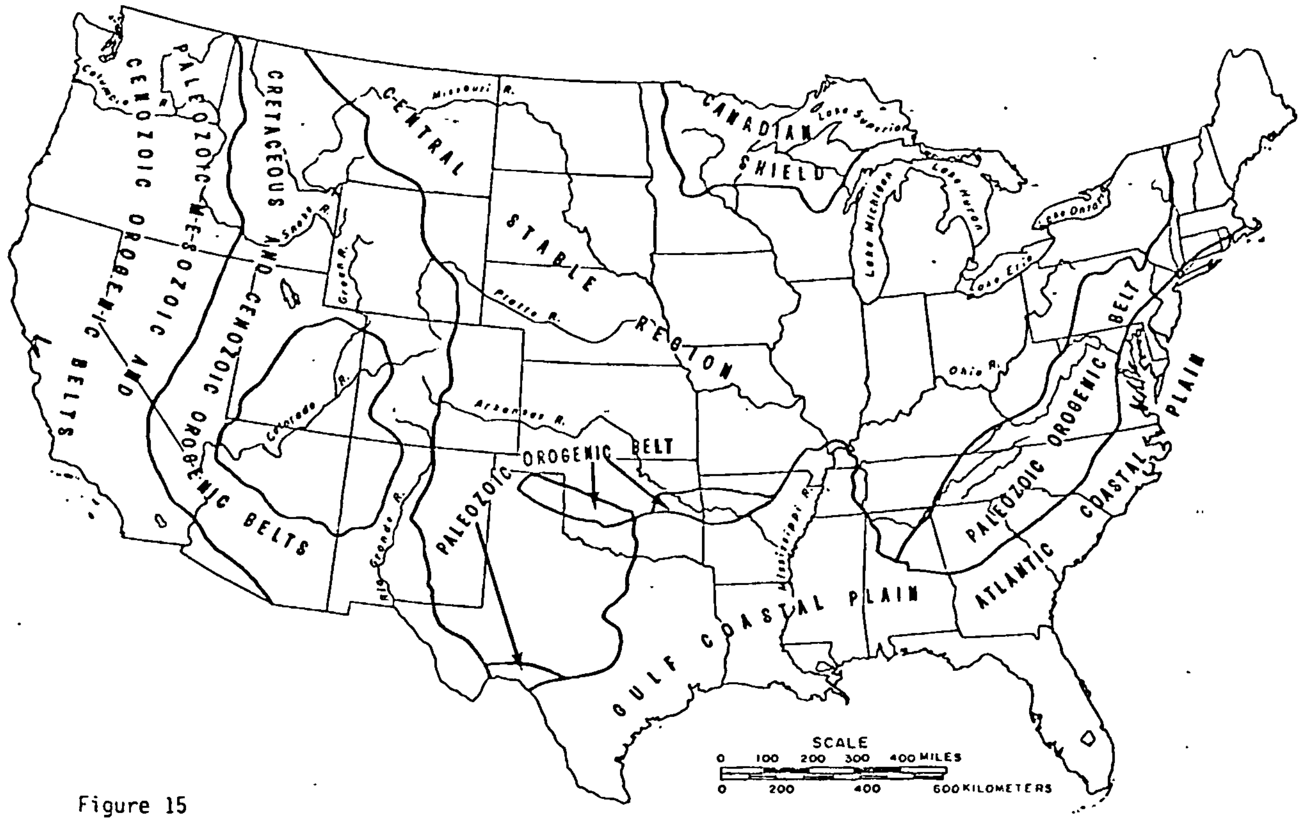


Figure 15

MAJOR TECTONIC DIVISIONS OF USA

(After Eardley, 1951)

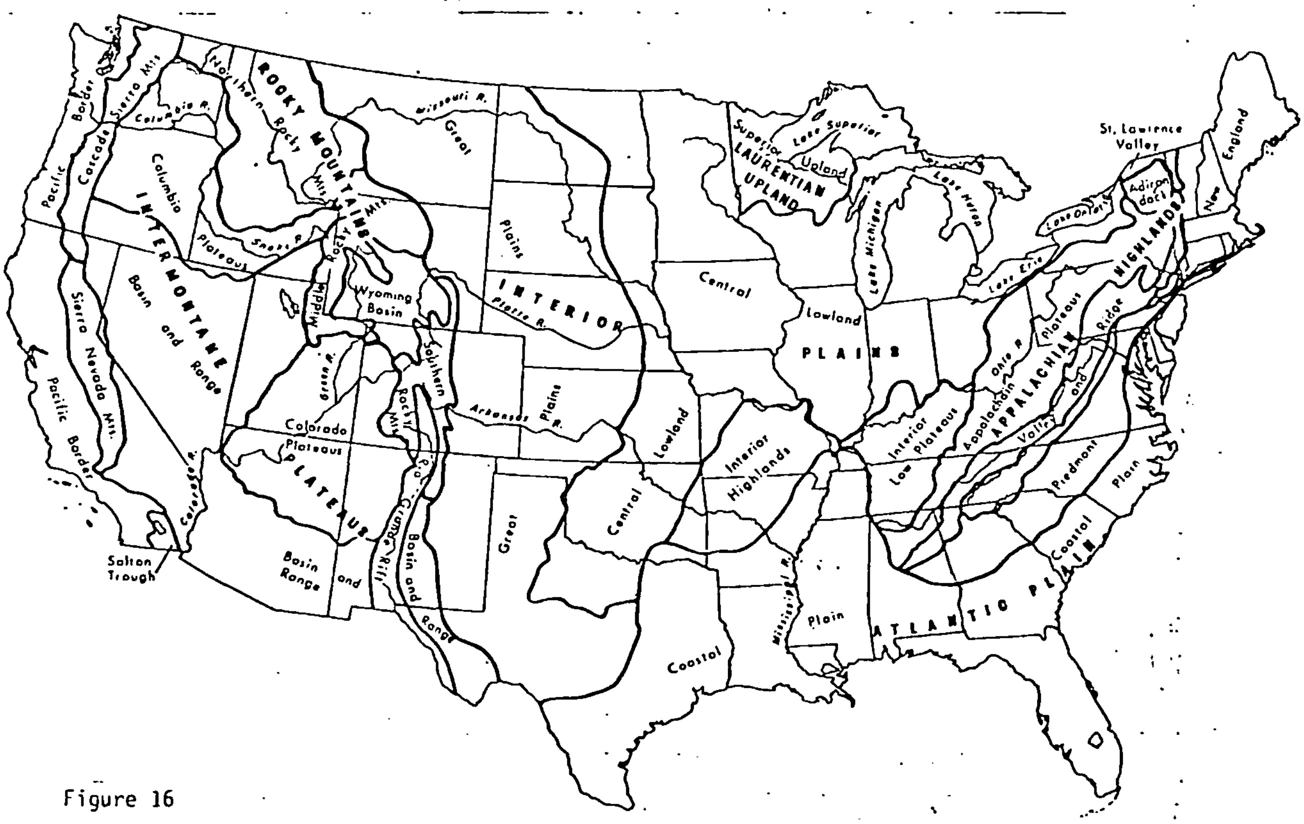


Figure 16

PHYSIOGRAPHIC MAP of USA

(after Fenneman, 1928)

In the southwestern corner of the central stable region, a system of ranges was elevated in Carboniferous time, and then during the Permian and Mesozoic it was largely buried. The ranges are known as the Ancestral Rockies in Colorado and New Mexico, and as the Wichita mountain system in Kansas, Oklahoma, and Texas. The late Cretaceous and early Tertiary Laramide orogenic belt was partly superposed on the Ancestral Rockies in Colorado and New Mexico, and a fragment of the central stable region was dismembered in the process to form the Colorado Plateau.

Orogenic Belts of the Atlantic Margin

The Paleozoic orogenic belts of the Atlantic margin bound effectively the southern, as well as the eastern, continental margin. The major belt is known as the Appalachian, and it consists of an inner folded and faulted division, the Valley and Ridge, and an outer compressed, metamorphosed, and intruded division, the Piedmont. Volcanic rocks and great intrusions of crystalline rock (batholiths) are important components of the outer division, but the inner folded and faulted belt is comparatively free of them. Both divisions are made up of very thick sequences of sedimentary rocks that have been metamorphosed.

The orogenic belt bordering the southern margin of the stable interior is mostly concealed by overlapping coastal plain deposits, but where exposed, it is a folded and faulted complex, somewhat similar to the inner Appalachian division.

The eastern extent or breadth of the Appalachian orogenic system and the nature and condition of the crust that lies east of it are not known, because of the cover of Atlantic Coastal plain sediments. The continental margin had begun to subside at least by early Cretaceous time, if not before. The gently sloping surface on the crystalline rocks has been traced eastward under this Cretaceous and Tertiary sedimentary cover to a depth of 10,000 feet, which is near the margin of the present continental shelf. Most units of the Coastal Plain sediments dip gently and thicken like a wedge oceanward as far as they have been traced by deep drilling and by seismic traverses. The Gulf coastal plain is continuous with the Atlantic coastal plain, and counting its shallowly submerged portions, it nearly encloses the Gulf of Mexico.

Orogenic Belts of the Pacific Margin

The great complex of orogenic belts along the Pacific margin of the continent evolved through a very long time. The oldest strata recognized are Ordovician. In Paleozoic time, the Pacific margin of the continent was a volcanic archipelago in appearance, and internally was a belt of profound compression and igneous intrusion. Inward from the archipelago, much volcanic material was deposited in a sagging trough and admixed with other sediments. The Permian, Triassic, and Jurassic were times of volcanism, and represent a continuation of essentially the same Paleozoic conditions well into the Mesozoic. In late Jurassic and early Cretaceous time, intense folding preceded batholithic intrusions (Nevadan orogeny) and the results of this great geologic activity now constitute large parts of the Coast Range of British Columbia, the ranges along the international border in British Columbia, Washington, and Idaho, the Klamath Mountains of southwestern Oregon and northern California, the Sierra Nevada Mountains of California, and the Sierra of Baja California. It is probable that this orogeny was caused by compression due to subduction of an oceanic plate beneath the western margin of the continent.

Following the Nevadan orogeny, a new trough of accumulation and a new volcanic archipelago formed west of the Nevadan belt, and a complex history of deformation and sedimentation carries down through the Cretaceous and Tertiary to the present, to result in the Coast Ranges of Washington, Oregon, and California. It is believed that subduction was active in this area until the last few million years (Dickinson and Snyder, 1979). Volcanism is active today in the Cascade Range.

The Columbia Plateau is a complex of flat-lying basaltic lava flows and airfall deposits that cover much of eastern Washington and Oregon. The main period of volcanism was Miocene, but the deposits merge smoothly eastward with the flows of the Snake River plain in Idaho where volcanism has been active in places in the past few hundred years. The volcanic rocks were deposited in a downwarped area and range in thickness up to perhaps 2 km. They were deposited on sedimentary rocks of Paleozoic and Mesozoic age. It is likely that the Basin and Range Province extends under the plateaus.

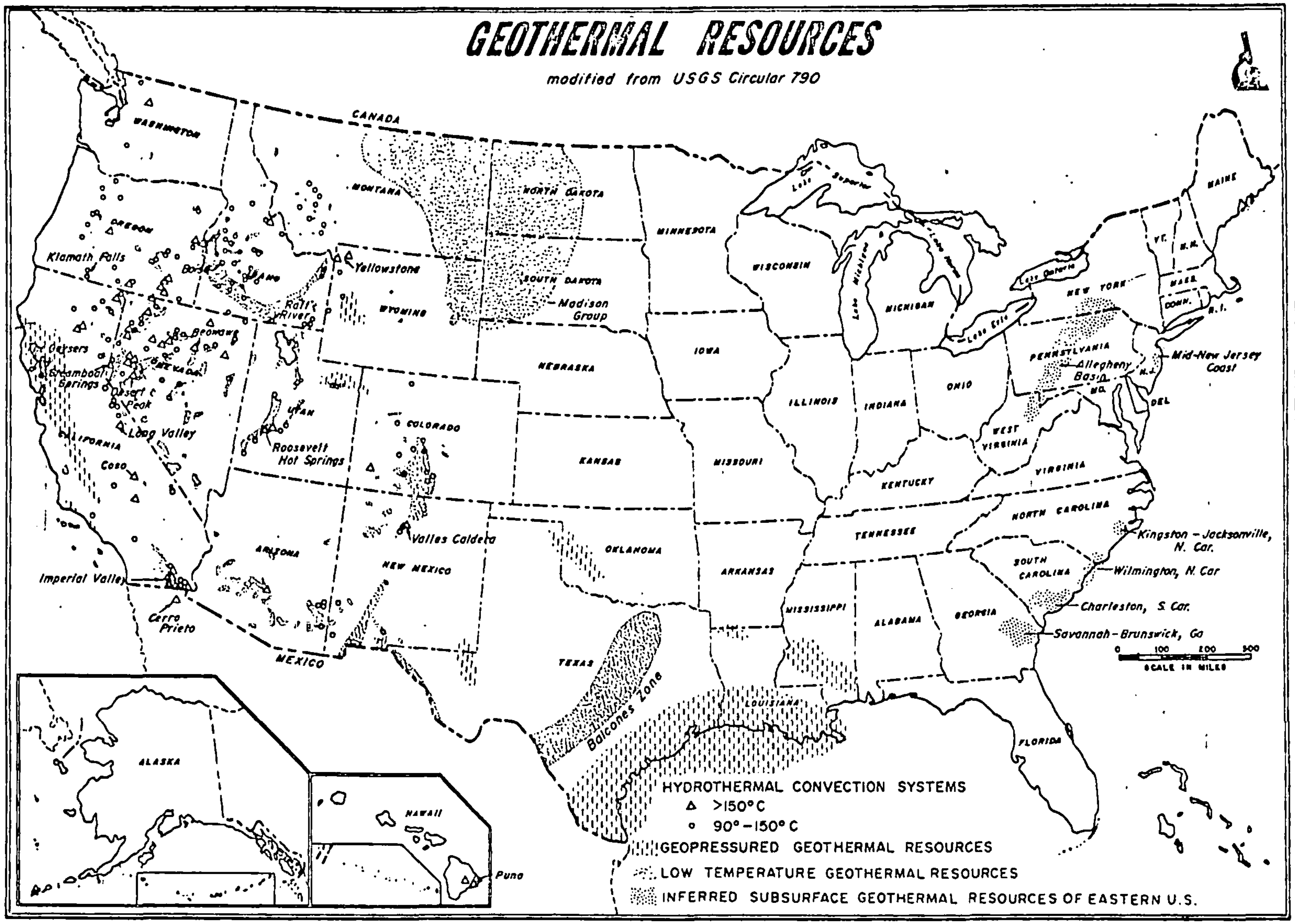
Orogenic Belts of the Rocky Mountains

During the complex and long orogenic history of the Pacific margin, the adjacent zone inward was one of gentle subsidence and sediment accumulation, comparatively free of volcanic materials during the Paleozoic.

The Paleozoic and all the Mesozoic sediments except the Upper Cretaeous of the Rocky Mountains may be divided into thick basin sequences on the west and fairly thin shelf sequences on the east. The line dividing the two lies approximately along the west side of the Colorado plateau and runs northward through western Wyoming and Montana to western Alberta. The shelf sequences were part of the central stable region until the late Cretaceous and early Tertiary (Laramide) orogeny. The eastern Laramide belt of folding and faulting extended through the shelf region of central and eastern Wyoming, central Colorado, and central New Mexico, forming the eastern Rocky Mountains and cutting off the Colorado plateau from the central stable region.

Following in the middle Tertiary, well after

Figure 17



Wright

the compressional Nevadan and Laramide orogenies of western North America, an episode of high-angle faulting occurred that created the Basin and Range physiographic province and gave sharp definition to many of its mountain ranges. The high-angle faults were superposed on both the Nevadan and Laramide belts; most of them are late Tertiary in age and some are still active. In many areas of the Basin and Range, volcanism occurred throughout the Tertiary and, especially along its eastern and western margins, it continues to the present time. Active volcanoes existed as recently as a few hundred years ago in parts of Idaho, Utah, Nevada, California, Arizona and New Mexico.

GEOTHERMAL RESOURCES IN THE CONTINENTAL UNITED STATES

Figure 17 displays the distribution of the various resource types in the 48 contiguous states. Information for this figure was taken mainly from Muffler et al. (1978), where a more detailed discussion and more detailed maps can be found. Not shown are locations of hot dry rock resources because very little is known. In addition, it should be emphasized that the present state of knowledge of geothermal resources of all types is poor. Because of the very recent emergence of the geothermal industry, insufficient exploration has been done to define properly the resource base. Each year brings more resource discovery, so that Figure 17 will rapidly become outdated.

Figure 17 shows that most of the known hydrothermal resources and all of the presently known sites that are capable or believed to be capable of electric power generation from hydrothermal convection systems are in the western half of the U.S. The preponderance of thermal springs and other surface manifestations of underlying resources is also in the west. Large areas underlain by warm waters in sedimentary rocks exist in Montana, North and South Dakota, and Wyoming (the Madison Group of aquifers), but the extent and potential of these resources is poorly understood. Another important large area much of which is underlain by low-temperature resources, is the northeast-trending Balcones fault zone in Texas. The geopressured resource areas of the Gulf Coast and surrounding states are also shown. Resource areas indicated in the eastern states are highly speculative because almost no drilling has been done to actually confirm their existence, which is only inferred at present.

Regarding the temperature distribution of geothermal resources, low- and intermediate-temperature resources are much more plentiful than are high-temperature resources. There are many, many thermal springs and wells that have water at a temperature only slightly above the mean annual air temperature, which is the temperature of most non-geothermal shallow ground water. Resources having temperatures above 150°C are infrequent, but represent important occurrences. Muffler et al. (1978) show a statistical analysis of the temperature distribution of hydrothermal resources and conclude that the cumulative frequency of occurrence increases exponentially as reservoir temperature decreases (Fig. 18). This relationship is based only on data for known occurrences having temperatures 90°C or higher. It is firmly enough established, however, that we can have confidence in the existence of a very large low-temperature resource base, most of which is undiscovered.

FREQUENCY OF OCCURRENCE VS TEMPERATURE

FOR GEOTHERMAL RESOURCES

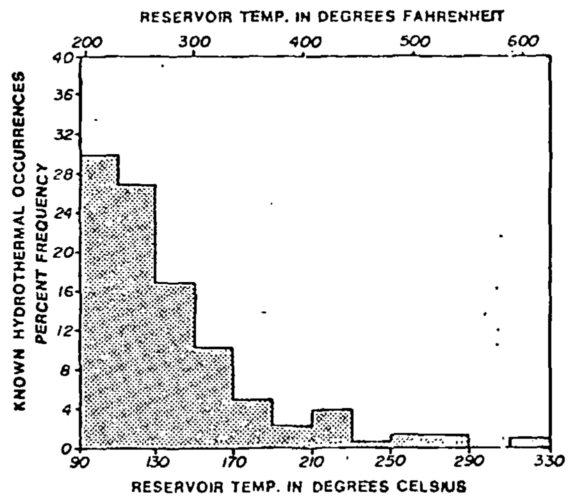


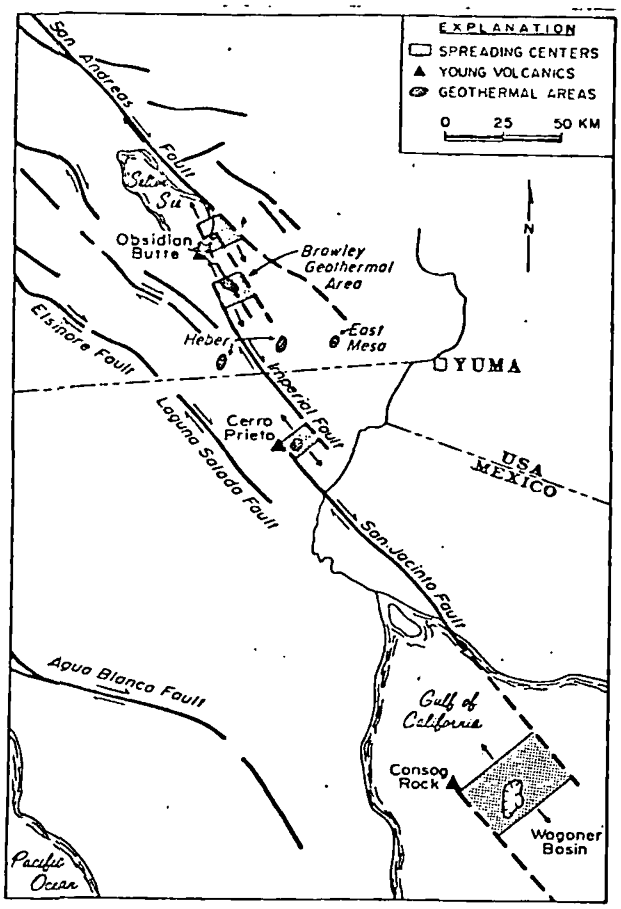
Figure 18

Let us consider the known geothermal occurrences in a bit more detail, beginning in the Western U.S.

Salton Trough/Imperial Valley, CA

The Salton Trough is the name given an area along the landward extension of the Gulf of California. It is an area of complex, currently active plate tectonic geologic processes. As shown on Figure 3, the crest of the East Pacific Rise spreading center is offset repeatedly northward, up the Gulf of California, by transform faulting. Both the rise crest and the transform faults come onto the continent under the delta of the Colorado River (Fig. 19) and the structure of the Salton Trough suggests that they underlie the trough. The offsetting faults trend northwest, parallel to the strike of the well-known San Andreas fault.

The Salton Trough has been an area of subsidence since Miocene times. During the ensuing years sedimentation in the trough has kept pace with subsidence, with shallow water sediments and debris from the Colorado River predominating. At present, 3 to 5 km of poorly-consolidated sediment overlies a basement of Mesozoic crystalline rocks that intruded Paleozoic and Precambrian sedimentary rocks. Detailed analysis of drilling data and of surface and downhole geophysics indicates that at least some of the known geothermal occurrences (Cerro Prieto, Brawley and the Salton Sea) are underlain by "pull-apart basins" apparently caused by crustal spreading above a local section of the East Pacific Rise crest (Elders, 1979). Very young volcanic activity has occurred at Cerro Prieto where a rhyodacite cone is known, and along the southern margin of the Salton Sea where rhyolite domes occur. The domes have an approximate age of 60,000 years (Muffler and White, 1969). The Cerro Prieto volcano has been difficult to date but may be about 10,000 years old (Wollenberg et al., 1980). Faulting is occurring at the present time as evidenced by the many earthquakes and earthquake swarms recorded there (Johnson, 1979).



MAJOR STRUCTURES OF SALTON TROUGH
(after Palmer et al., 1975)

Figure 19

The Cerro Prieto field is the best understood geothermal occurrence in the Salton Trough because of the drilling done there. We may take it as an example of a Salton Trough resource type. This field currently produces 150 MWe and there are plans by the Comision Federal de Electricidad in Mexico to enlarge its capacity to 370 MWe by 1985. The field is water-dominated and the more than 60 wells produce from depths of 1.5 to over 3 km. Fluid temperatures range from about 200°C to over 350°C (Alanso, et al., 1979). The rocks are composed of an upper layer of unconsolidated silts, sands and clays, and a layer of consolidated sandstones and shales overlying the crystalline basement (Puete Cruz and de la Pena, 1979). Two principal reservoir horizons occur in sandstones within the consolidated sequence, and enhanced production has been noted in the vicinity of faults, indicating that fracture permeability is important, although intergranular permeability due to dissolution of minerals by the geothermal fluids is believed to be important also (Lyons and Van de Kamp, 1980). Reservoir recharge is apparently from the northeast and east and consists, at least partly, of Colorado River water (Truesdell et al., 1980).

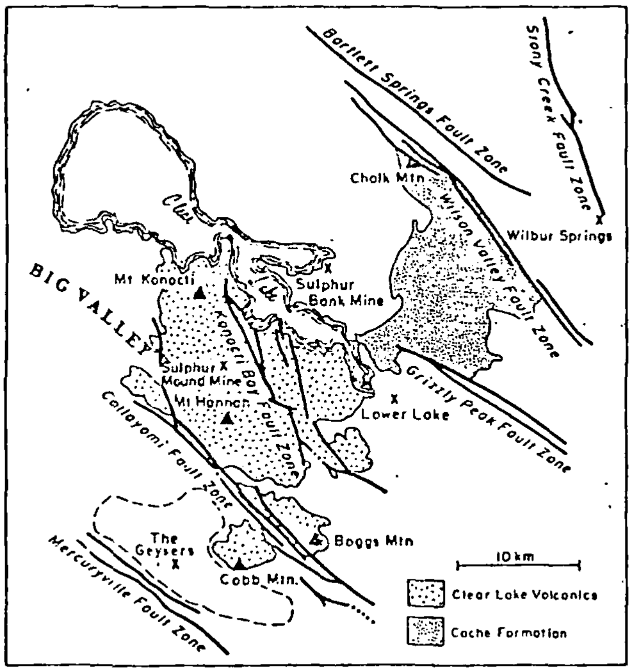
The geothermal fluid from Cerro Prieto, after steam separation, contains about 25,000 ppm total dissolved solids. This figure is much lower than some of the other resources in the Salton Trough. For example, the Salton Sea area contains 20 to 30 percent by weight by solids (Palmer, 1975). Primarily because of problems associated with this high salinity, no significant use has been made of Salton Sea fluids to date.

The heat source(s) for the several Salton Trough resources are unknown. Hot, partly molten rock at shallow depth (5-15 km) could underly at least some of the resource areas, or alternatively the active faulting could provide a mechanism where water could circulate to depths great enough to be heated by the enhanced geothermal gradient.

The Geysers, CA

The Geysers geothermal area is the world's largest producer of electricity from geothermal fluids with 908 MWe on line and an additional 880 MWe scheduled by 1986. This area lies about 150 km north of San Francisco. The portion of the resource being exploited is a vapor-dominated field having a temperature of 240°C, as previously discussed. The ultimate potential of the vapor-dominated system is presently believed to be around 2000 MWe. Associated with the vapor-dominated field are believed to be several unexploited hot water-dominated reservoirs whose volume and temperature are unknown.

The geology of The Geysers area is complex, especially structurally. Reservoir rocks consist mainly of fractured greywackies, sandstone-like rocks consisting of poorly sorted fragments of quartzite, shale, granite, volcanic rocks and other rocks). The fracturing has created the permeability necessary for steam production in quantities large enough to be economically exploitable. Overlying the reservoir rocks, as shown in Figure 21, is a series of impermeable metamorphosed rocks (serpentinite, greenstone, melange and metagranite) that form a cap on the system. These rocks are all complexly folded and faulted. They are believed to have been closely associated with and perhaps included in subduction of the eastward-moving plate (Fig. 3) under the continent. This subduction apparently ended 2 to 3 million years ago.



MAJOR STRUCTURES in
THE GEYSERS-CLEAR LAKE AREA
(After Goff, 1980)

Figure 20

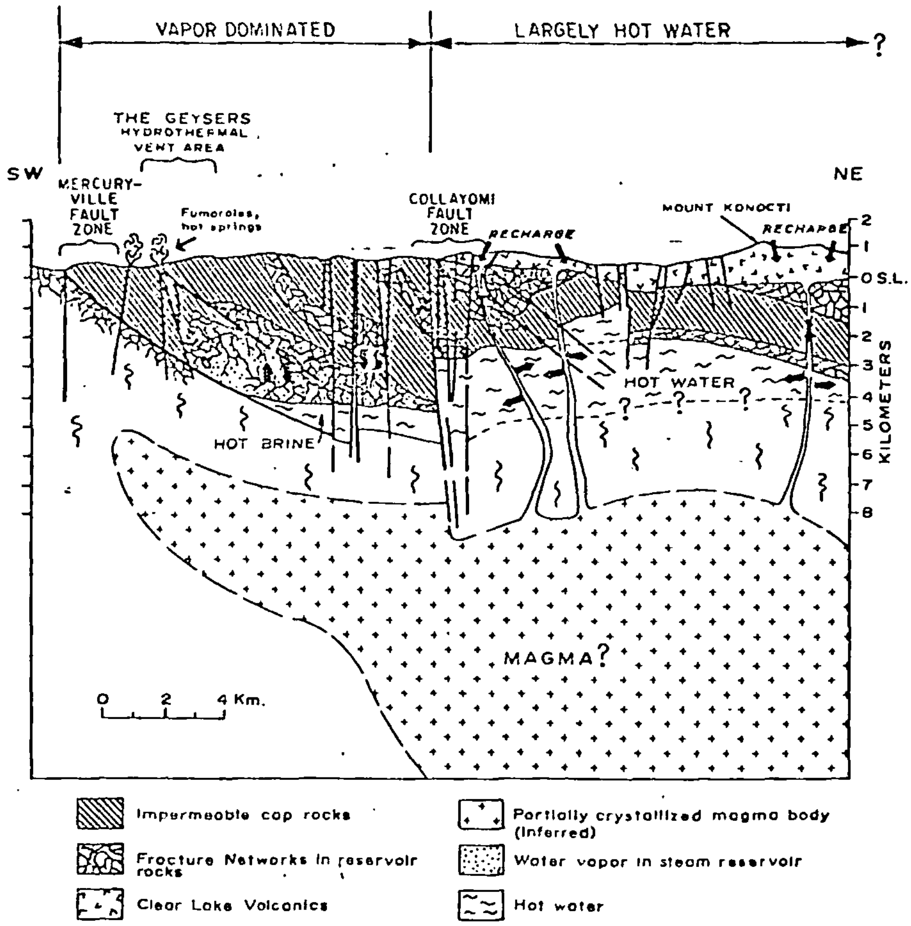
As shown in Figure 20, the presently known steam field is confined between the Mercuryville fault zone on the southwest and the Collayomi Fault zone on the northeast. The northwest and southeast margins are not defintely known. To the east and northeast lies the extensive Clear Lake volcanic field composed of dacite, rhyolite, andesite and basalt. The interval of eruption for these volcanics extends from 2 million years ago to 10,000 years ago, with ages progressively younger northward (Donnelly, 1977). The Clear Lake volcanics are very porous and soak up large quantities of surface water. It is believed that recharge of a deep, briny hot-water reservoir comes from water percolating through the Clear Lake volcanics, and that this deep reservoir may supply steam to the vapor-dominated system through boiling (Fig. 21) although these ideas are not universally supported by geologists and the deep water table has never been intersected by drilling.

The postulated water-dominated geothermal reservoirs do not occur everywhere in the Clear Lake volcanics. At several locations drill holes have found temperatures of 200°C at depths of only 2000 m, but the rocks are tight and impermeable (Goff, 1980). Fractured areas apparently host the water-dominated reservoirs at the Wilbur Springs district (Thompson, 1979), the Sulphur Bank Mine (White and Roberson, 1962) and other smaller occurrences. Potential in The Geysers area for discovery of additional exploitable resources is good.

The Basin and Range

The Basin and Range province extends from Mexico into southern Arizona, southwestern New Mexico and Texas on the south, through parts of California, Nevada and Utah, and becomes ill-defined beneath the covering volcanic flows of the Columbia Plateau on the north (Fig. 16). This area, especially the northern portion, contains abundant geothermal resources of all temperatures and is perhaps the most active area of exploration in the U.S. outside of the Imperial Valley and The Geysers areas. Resources along the eastern and western margins of the province are both more abundant and of higher temperature. Although no electrical power is presently being generated from geothermal resources in this area, plans have been announced to develop 20 MWe from Roosevelt Hot Springs in Utah and 10 MWe from an area yet to be selected in Nevada. Candidate sites in Nevada include Steamboat Springs, Dixie Valley, Desert Peak and Beowawe. Exploration is being conducted at probably 20 or more sites in the Basin and Range, including, in addition to those named above, Cove Fort, Utah; Tuscarora, McCoy, Baltazor, Leach Hot Springs, San Emidio, Soda Lake, Stillwater, and Humboldt House, Nevada; and Surprise Valley, Long Valley Caldera and Coso, California. Direct application of geothermal energy for industrial process heating and space heating are currently operating in this area at several sites including Brady Hot Springs (vegetable drying), Reno (space heating) and Salt Lake City (greenhouse heating).

The reasons for the abundance of resources in the Basin and Range seem clear. This area, especially at its margins, is an active area geologically. Volcanism only a few hundred years old is known from tens of areas, including parts of west central Utah on the east (Nash and Smith, 1977) and Long Valley caldera on the west (Rinehart and Huber, 1965). The area is also active seismically and faulting that causes the uplift of mountain ranges in this area also serves to keep pathways open for deep fluid circulation at numerous locations. Rocks in the Basin and Range consist of Paleozoic and Mesozoic sandstones, limestones and shales that lie on Precambrian metamorphic and intrusive rocks. These rocks were deformed, complexly in some places, during the Nevadan and Laramide orogenies, as discussed above, and some base and precious metal deposits were formed. Beginning in mid-Tertiary times volcanic activity increased many fold with both basaltic and rhyolitic rocks being erupted. Extentional stresses also began to operate and a sequence of north-south mountain ranges were formed which separate valleys that have been filled with erosional debris from the mountains (Eardley, 1951). In some places more



CRUSTAL MODEL FOR THE GEYSERS - CLEAR LAKE AREA, CA.

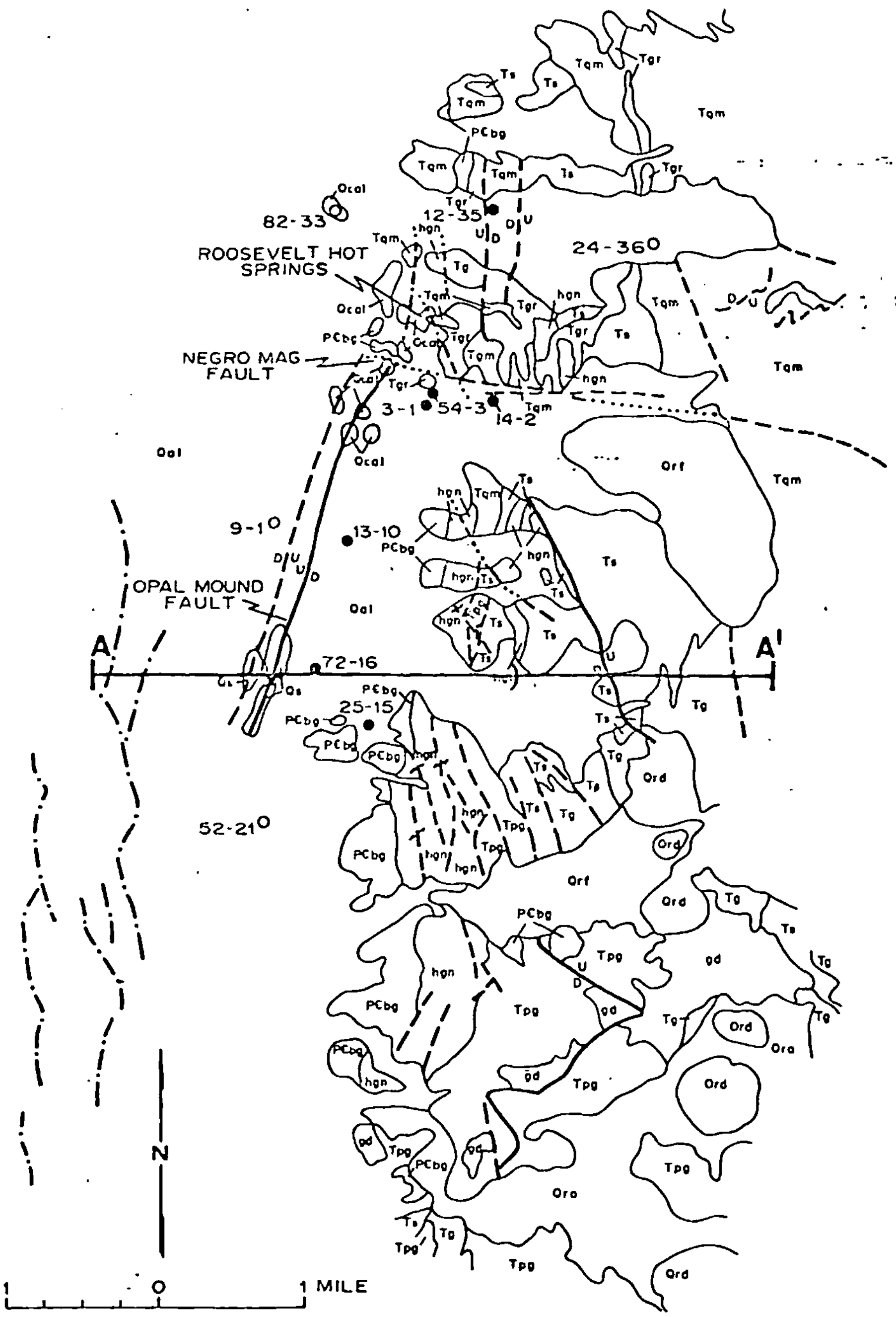
(after McLaughlin, 1977)

Figure 21

than 2 km offset has occurred along range-front faults, and the valleys may contain a hundred to as much as 3,000 m of unconsolidated erosional debris. This activity persists to the present time.

As an example of a Basin and Range hydrothermal system we will discuss Roosevelt Hot Springs, although it should not be supposed to be typical of all high temperature occurrences in this province. This geothermal area has been studied in detail for the past six years (Nielson et al., 1978; Ward et al., 1978). The oldest rocks exposed (Figs. 22 and 23) are Precambrian sedimentary rocks that have been extensively metamorphosed. These rocks were intruded during Miocene time by granitic rocks (diorite, quartz monzonite, syenite and granite). Rhyolite volcanic flows and domes were emplaced during the interval 800,000 to 500,000 years ago. The area has been complexly faulted by north to northwest-trending high angle faults and by east-west high-angle faults. The Negro Mag fault is such an east-west fault that is an important controlling structure in the north portion of the field. The north-trending Opal Mound fault apparently forms the western limit of the system. The oldest fault system is a series of low-angle denudation faults (Fig. 23) along which the upper plate has moved west by about 600 m and has broken into a series of discrete blocks. Producing areas in the southern portion of the field are located in zones of intersection of the upper plate fault zones with the Opal Mound and other parallel faults. Producing zones in the northern part of the region are located at the intersection of north-south and east-west faults. The permeability is obviously fracture controlled.

Seven producing wells have been drilled in the area (Fig. 22). Fluid temperature is about 260°C and the geothermal system is water-dominated. Average well production is perhaps 318,000 kg/hr (700,000 lbs./hr). Initial plans are for a 20 MWe power plant with two 50 MWe plants to be installed as knowledge of reservoir performance increases.



EXPLANATION

Qal - alluvium	Tg - granite
Qcal - silicified alluvium	Ts - syenite
Qs - siliceous sinter	Tpg - porphyritic granite
Qrd - rhyolite domes	Tqm - quartz monzonite
Qra - pyroclastic deposits	gd - biotite diorite
Qrf - rhyolite flows	hgn - foliated hornblende granodiorite
Tgr - fine-grained granite	PЄbg - banded gneiss

GEOLOGIC MAP
ROOSEVELT HOT SPRINGS, UTAH
(from Nielson et al., 1978)

Figure 22

GEOLOGIC CROSS SECTION ROOSEVELT HOT SPRINGS, UTAH

(from Nielson et al., 1978)

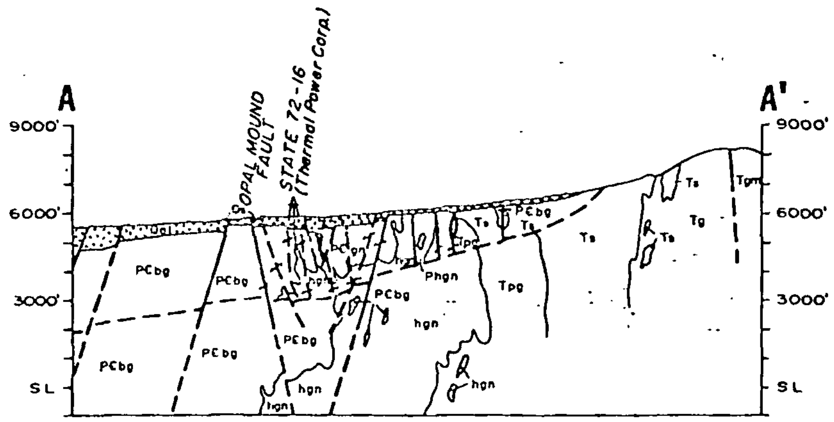


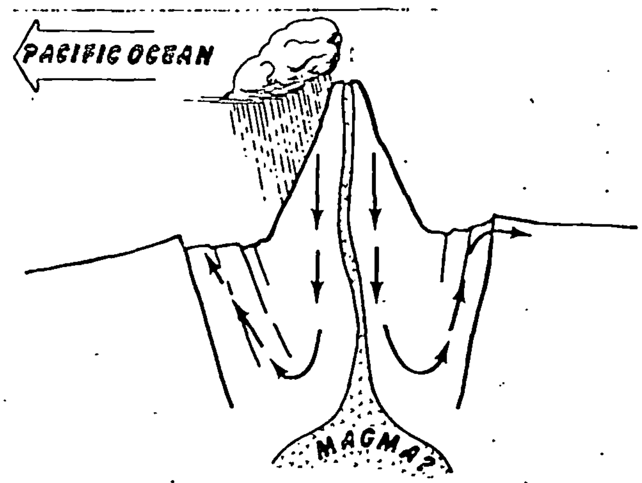
Figure 23

Cascade Range and Vicinity

The Cascade Range of northern California, Oregon, Washington and British Columbia is comprised of a series of volcanoes, 12 of which have been active in historic times. The May 18, 1980 eruption of Mount St. Helens attests to the youth of volcanic activity here. The Cascade Range probably lies over a subduction zone (Fig. 3) and magma moving into the upper crust has transported large amounts of heat upward. In spite of the widespread, young volcanism, however, geothermal manifestations are not as plentiful as one would suppose they should be (Fig. 17). Figure 24 illustrates in schematic form that the high rainfall and snowfall in the Cascades are believed to suppress surface geothermal manifestations through downward percolation of the cold surface waters in the highly permeable volcanic rocks. In the absence of surface manifestation, discovery of these resources becomes much more difficult.

No producible high-temperature hydrothermal systems have yet been located in the Cascades, although they are belived to exist. Geological and geochemical evidence indicates that a vapor-dominated system is present at Lassen Peak in California, but it lies within a national park, and will not be developed. Elsewhere hydrothermal systems having predicted temperatures greater than 150°C are postulated at Newberry Caldera in Oregon and Gamma Hot Springs in Washington, but drill evidence has not been obtained (Muffler et al., 1978). Industry's exploration effort so far in this area has been minimal.

The use of geothermal energy for space heating at Klamath Falls, Oregon is well known (Lund, 1975; Lund, 1980), and numerous hot springs and wells occur in both Oregon and Washington. Potential for discovery of resources in all temperature categories is great.



CASCADES GEOTHERMAL ENVIRONMENT

Figure 24

Columbia Plateaus

The Columbia Plateaus area is an area of young volcanic rocks, mostly basalt flows, that cover much of eastern Washington and Oregon and continue in a curved pattern into Idaho, following the course of the Snake River (see below).

There are no hydrothermal resources having temperatures >90°C known through drilling in this area. However, there are numerous warm springs and wells that indicate the presence of geothermal resources potentially suitable for direct heat uses.

Snake River Plain

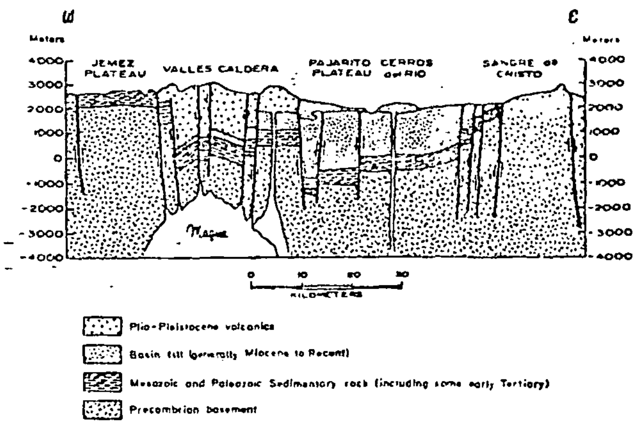
The basalt flows and other volcanic deposits of the Snake River Plain are an extension of the Columbia Plateau eastward across southern Idaho to the border with Wyoming. The plain is divided into a western part and an eastern part. Thermal waters occur in numerous wells and springs in the western portion, especially on or near the edges of the plain. Geochemically indicated resource temperatures exceed 150°C at Neal Hot Springs and Vale, Oregon and Crane Creek, Idaho, but indicated temperatures for most resources are lower. Younger volcanic rocks occur in the eastern part of the plain, but no high-temperature resources (T>150°C) are yet identified, although numerous areas have warm wells and springs. This part of the plain is underlain by a high-flow cold-water aquifer that is believed to mask surface geothermal indications.

Direct use of hydothermal energy for space heating is famous at Boise, where the Warm Springs district has been heating homes geothermally for almost 100 years (Mink et al., 1977). Also in this area is the Raft River site where DOE is currently constructing a 5 MWe binary demonstration plant on a hydrothermal resource whose temperature is 147°C.

Rio Grande Rift

The Rio Grande Rift is a north-trending tectonic feature that extends from Mexico through central New Mexico and ends in central Colorado (Figs. 16 and 17). It is a down-dropped area that has been filled with volcanic rocks and erosional debris from the bordering plateaus and mountains (Fig. 25). The rift began to form in late Oligocene times, and volcanic and seismic activity have occurred subsequently to the present. Young volcanism, faulting and high heat flow characterize the area today.

There are several low- and intermediate-temperature hydrothermal convection systems in this area, but the only high-temperature system that has been drill tested to any significant extent and where production is proven is a hot water-dominated system in the Valles Caldera (Dondanville, 1978). Surface manifestations at the Baca No. 1 location in the caldera include fumaroles, widely distributed hot springs and gas seeps. Hydrothermal alteration extends over 40 km^2. Deep drilling has encountered a hydrothermal convection system in fractured Tertiary volcanic, Paleozoic sedimentary and Precambrian granitic rocks at an average depth of 2 to 3 km. Temperatures as high as 300°C have been recorded and the average production temperature will likely be 260°C. There are current plans for a 50 MWe flash steam plant at this location. Also located near the caldera is the site of Los Alamos National Laboratory's hot dry rock experiment at Fenton Hill. Both the hot dry rock site and the hydrothermal convection system probably derive their heat from magma that has provided the material for the several episodes of volcanism that created the caldera structure.



SCHEMATIC CROSS SECTION
RIO GRANDE RIFT, NM

(after Eichelberger and Westrich, 1980)

Figure 25

Elsewhere in the Rio Grande Rift, there are numerous hot springs and wells. Discovery potential is high, although there are no known sites where discovery of fluids in excess of 150 to 170°C is indicated by present data (Harder et al., 1980).

The Madison and other Aquifers

Underlying a large area in western North and South Dakota, eastern Montana and northeastern Wyoming are a number of aquifers that contain thermal waters. These aquifers have been developed in carbonates and sandstones of Paleozoic and Mesozoic age. The permeability is both intergranular and fracture controlled in the case of the sandstones (e.g. the Dakota Sandstone) and fracture and open spaces in the carbonates (e.g. the Madison Limestone). At least some of the aquifers will produce under artesian pressure. Depths to production vary widely but average perhaps 2,000 ft. Temperatures are 30-80°C (Gries, 1977) in the Madison but are lower in other shallower aquifers such as the Dakota.

The U.S. Geological Survey is completing an intensive study of these aquifers, and the results will form a much firmer basis for hydrothermal development than presently exists. Direct use of the thermal water is being made at a few locations today, and it is evident that the potential for further development is substantial.

The Balcones Zone, Texas

Thermal waters at temperatures generally below 60°C occur in a zone that trends northeasterly across central Texas. Many of the large population centers are in or near this zone, and there appears to be significant potential for geothermal development in spite of the rather low temperatures.

An initial assessment of the geothermal potential has been documented by Woodruff and McBride (1979). The thermal waters occur in a band broadly delimited by the Balcones fault zone on the west and the Luling-Mexia-Talco fault zone on the east. In many locations the thermal waters are low enough in content of dissolved salts to be potable, and indeed many communities already tap the warm waters for their municipal water supplies.

The geothermal aquifers are mostly Cretaceous sandstone units, although locally thermal waters are provided from Cretaceous limestones and Tertiary sandstones. The thermally anomalous zone coincides with an ancient zone of structural weakness dating back more than 200 million years. The zone has been a hinge line with uplift of mountain ranges to the north and west and downwarping to the south and east. Sediments have been deposited in the area of downwarping, and the rate of sedimentation has kept pace with sinking, keeping this area close to sea level. Structural deformation of the sediments, including faulting and folding, and interfingering of diverse sedimentary units have resulted in the complex aquifer system of today.

The source of the anomalous heat is not known with certainty but several postulates are (Woodruff and McBride, 1979): 1) deep circulation of ground waters along faults; 2) upwelling of connate waters, originally trapped in sediments now deeply buried; 3) stagnation of deep ground waters owing to faults that retard circulation; 4) local hot spots such as radiogenic heat sources (intrusions) within the basement complex, or; 5) other loci of high heat flow.

A minor amount of direct use is being made of these waters at present, and potential for further development is good.

Other Areas--Eastern Half of U.S.

Hydrothermal resources in other areas of the continental U.S. besides those mentioned above are very poorly known. There is believed to be potential for thermal waters of about 100°C at a number of locations along the Atlantic Coastal plain associated with buried intrusions that are generating anomalous heat through radioactive decay of contained natural uranium, thorium and potassium. Examples of such areas are shown at Savannah-Brunswick, Charleston, Wilmington, Kingston-Jacksonville and the mid-New-Jersey Coast. One drill test of such an area (Delmarva Penninsula near Washington, D.C.) has been conducted with inconclusive results regarding amount of thermal water that could be produced. This is the only geothermal test well so far in the east. Less than a dozen warm springs and wells are known at present. The Allegheny Basin is outlined on Figure 17 because it has potential for thermal fluids in aquifers buried deeply enough to be heated in a normal earth's gradient. Parts of Ohio, Kansas, Nebraska, and Oklahoma as well as other states are belived to have potential for low-temperature fluids. No drill tests have been conducted, however.

Hawaiian Islands

The chain of islands known as the Hawaiian archipelago stretches 2500 km in a northwest-southeast line across the Pacific ocean from Kure and Midway Islands to the Big Island of Hawaii. Built of basaltic volcanic rocks, this island chain boasts the greatest volcanic masses on earth. The volcano Kilauea rises 9800 m above the floor of the ocean, the world's largest mountain in terms of elevation above its base. The Kilauea, Mauna Loa and other vents on the big island are in an almost continual state of activity, but by contrast volcanoes on the other islands have shown little recent activity. Haleakala on the island of Maui is the only other volcano in the state that has erupted in the last few hundred years, and the last eruption there was in 1790 (MacDonald and Hubbard, 1975).

Several of the Hawaiian islands are believed to have geothermal potential. The only area where exploration has proceeded for enough to establish the existence of a hydrothermal reservoir is in the Puna district near Kapoho along the so-called "East Rift", a fault zone on the east flank of Kileaua. Here a well was completed to a depth of 1965 m (Helsley, 1977) with a bottom-hole temperature of 358°C. Little is known in detail of the reservoir at present, but it is believed to be fracture-controlled and water-dominated. A 3 MWe generator is currently being installed and is scheduled for start-up in mid-1981. Success of this project would undoubtedly spur further development at this site.

Elsewhere on the islands potential for occurrence of low- to moderate-temperature resources has been established at a number of locations on Hawaii, Maui and Oahu, although no drilling to establish existence of a resource has been completed (Thomas et al., 1980).

Alaska

Very little geothermal exploration work has been done in Alaska. A number of geothermal occurrences are located on the Alaska Peninsula and the Aleutian Islands and in central and southeast Alaska. The Aleutians and the Peninsula overly a zone of active subduction (Fig. 3), and volcanoes are numerous. None of the identified hydrothermal convection systems here have been studied in detail.

Low- and moderate-temperature resources are indicated in a number of locations in Alaska by occurrence of hot springs (Muffler et al., 1978). One area that has been studied in more

TABLE 2

GEOTHERMAL ENERGY OF THE UNITED STATES
After Muffler et al. (1979) Table 20

RESOURCE TYPE	ELECTRICITY (MWe for 30 yr)	BENEFICIAL HEAT (10^{18} joules)	RESOURCE (10^{18} joules)
Hydrothermal			
Identified	23,000	42	400
Undiscovered	72,000-127,000	184-310	2,000
Sedimentary Basins	?	?	?
Geopressured (N. Gulf of Mexico)			
Thermal			270-2800
Methane			160-1600
Radiogenic	?	?	?
Hot Rock	?	?	?

detail and has had limited drilling is Pilgrim Hot Springs (Turner et al., 1980). This site is 75 km north of Nome, Alaska. Initial drilling has confirmed the presence of a hot water reservoir about 1 km^2 in extent that has artesian flow rates of 200-400 gallons/minute of 90°C water. Geophysical data suggest that the reservoir is near the intersection of two inferred fault zones. Further exploration work will be required to determine the potential of this reservoir.

POTENTIAL FOR GEOTHERMAL DEVELOPMENT

A small industry exists in the U.S. that is beginning the development of high-temperature hydrothermal resources for electrical power production. Developers involved are mainly large petroleum companies and potential users of the hydrothermal fluids are electric utilities. Exploration for high-temperature resources is being conducted at a rather low level, mainly because development of geothermal resources is not yet economic.

There is virtually no industry activity to develop geothermal resources for direct heat uses in the U.S. Good inventories of low- and moderate-temperature resources are only now becoming available in map form through efforts of the Federal geothermal program. And there has been very little drill testing that is necessary to prove resource viability so that money could be obtained for construction of utilization systems.

Muffler et al. (1978) have dealt with the problem of how much accessible resource exists in the U.S. both at known sites and those that are undiscovered. They conclude that the undiscovered resource base is on the order of 3 to 5 times greater than the resources known today. These figures do not include possible hot dry rock or other more speculative resources. Table 2 is a summary of the current estimate of the geothermal resource base as taken from Muffler et al. (1978). This table demonstrates our lack of resource knowledge through the ranges and relative amounts of undiscovered resources and through the many missing numbers We can conclude, however, that the geothermal resource base is large in the U.S

The amount of geothermal energy that will be in use at various times in the future is a topic of much discussion. It is no trivial exercise to estimate this number. Table 3 shows the best current estimates (Anon., 1980; Anon., 1981a; Anon., 1981b).

TABLE 3

GEOTHERMAL DEVELOPMENT POTENTIAL

	Estimated Use by Year 2000	
	ELECTRICAL (MW)	DIRECT HEAT (10^{15} BTU)
Hydrothermal	12,800	0.57
Geopressured	2,000	3.0 (methane)
Hot Dry Rock	700	0.007

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"twist off" is said to have occurred. There are techniques for reconnecting to the broken off section of pipe. It may be necessary to disconnect and retrieve the drill stem length by length until the stuck section is located. This section may be explosively freed.

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 reponse 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.

A second, important problem in geothermal well logging is general lack of probes that will work in an environment where temperature exceed 100-150°C. Electric components have generally not been available for temperatures this high and indeed have not generally been required by the petroleum industry, which makes the most use by far of well logging. Sandia National Laboratories in the USA has had an active and successful research program sponsored by the

U.S. Department of Energy, to develop electronic components and logging tools for use in the geothermal environment, and so appropriate logging equipment is now becoming available.

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 performedfor cased intervals. It is common practice not to repeat logs in sections of the well that have been previously logged, but simly to provide adequate overlap with the previous logging run to facilitate _____ of logs that may be made with different instruments and different calibrations on successive logging runs.

Few developers or drilling contrctors offer logging services themselves. Geophysycal 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 1.6-1 is given a brief summary of logs that have been applied to geothermal well logging, and a brief explanation of these logs follows herewith.

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 automatically during the logging process. Three- or four-arm cliper 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 $\pm$ 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, 1979). As an example of the utility of these plots, bulk density is plotted against neutron porosity in Figure 1.6-G 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, 1979). 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.

WRIGHT

III. EXPLORATION RESULTS AND RECOMMENDATIONS

A. Drilling Success Ratios

The determination of the success of a geothermal well is somewhat subjective. In general, we will consider a well successful if it encounters fluids at a temperature and flow rate which will achieve the desired purpose of the well, such as the generation of electrical power. In a number of instances, wells drilled for electrical applications have encountered fluids which are suitable for direct heat applications but not for the generation of electricity. These wells are considered failures in our evaluation. This is justified by the fact that few if any of these wells are now being used in direct heat projects, although some are undergoing testing. This phenomena is due to the lack of interest of most of the major geothermal producers in direct heat applications.

An annual update of geothermal drilling in the western United States is provided in issues of Geothermal Energy Magazine (Smith et al., 1976, 1977, 1978, 1979, 1980; Ehni, 1981). Some of the results of the past five years of geothermal drilling are presented in Table III-1. Clearly the available data are too scanty to provide statistically significant conclusions. However, some trends are perhaps developing. The total wells drilled are increasing regularly. Most of these are either producers or step-out wells within known districts, principally The Geysers and the Imperial Valley. Table III-2 illustrates general data for The Geysers geothermal field, including development, step-out, workover, and exploration wells. This area is characterized by impressive success ratios (successful wells divided by total wells

Table III-1. Drilling data for the geothermal industry in the western United States (from Smith et al., 1976, 1977, 1978, 1979, 1980; Enhi, 1981).

YEAR	TOTAL HOLES	TOTAL FOOTAGE	TOTAL WELLS	TOTAL PRODUCERS	SUCCESS RATIO	WILDCATS TOTAL	WILDCATS PRODUCERS	WILDCATS SUCCESS RATIO
1975	51	355,143	46	37	.80	6	1	.17
1976	65	437,752	52	39	.75	21	2-3	.09-.14
1977	52	374,129	47	25	.53	15	0	0
1978	58	433,703	49	30	.61	13	2	.15
1979	77	552,329	61	42	.69	17	2	.12
1980	82	595,002	66	51	.77	15	2	.13

Total holes include production wells, wildcats, injection wells, and deep observation holes.
Total wells include total holes minus wells drilled for injection, observation, and workover.

Table III-2. Drilling data for The Geysers geothermal field, including development, stepout, and exploration holes (references of Table III-1. Footage drilled includes workovers. NA indicates data not available.)

YEAR	1975	1976	1977	1978	1979	1980
drilled	24	30	32	25	31	38
producers	20	20	19	14	25	35
workover	NA	NA	NA	2	2	4
success ratio	.83	.66	.59	.56	.81	.92
footage	197,373	243,936	260,495	237,481	289,311	323,329

drilled) unequaled elsewhere in the geothermal industry. Table III-3 gives drilling data for the western U.S. excluding that done in The Geysers geothermal field.

In contrast with the success rates at The Geysers, Table III-4 presents information from Union Oil Company's Baca Project of the Valles Caldera in New Mexico. The project has been underway since 1973. Since 1978 the U.S. Department of Energy and Public Service Company of New Mexico have participated with Union in the 50 MW Baca Geothermal Demonstration Power Plant. All data generated by that project are now in the public domain. Prior to the initiation of the GDPP, Union had drilled 10 wells, four of which were able to produce 320,000 lbs/hr of steam. Since then twelve bottom hole locations have been drilled with only one producer, which contributes 30,000 lbs/hr steam to the total. The entire project is now in jeopardy. Part of the reason for this difficulty is that the project area and the size of the power plant have been defined by agreement. Therefore it is not possible for Union to explore their adjacent lands or change the size of the power plant. This does not change the fact that, rather than improving drilling success as the project matured, Union suffered the opposite. The principal problem with the lack of productivity has been the low permeability of the geothermal reservoir. Temperatures in most of the holes are at least 250°C with geothermometers suggesting reservoir temperatures over 300°C.

Success ratios of drilling in frontier environments can be formulated by looking at data from the Industry Coupled Program sponsored by the Department of Energy between 1978 and 1981. The

Table III-3. Drilling data for the western U. S. excluding The Geysers. Footages are for total drilled including injection, observation, workover, development, stepout, and exploration (references of Table III-1.)

	TOTAL WELLS	TOTAL FOOTAGE	TOTAL PRODUCERS	SUCCESS RATIO
1975	22	157,771	17	.77
1976	22	193,816	19	.86
1977	17	113,634	6	.35
1978	24	196,222	16	.67
1979	30	263,018	17	.57
1980	28	271,673	16	.57

Table III-4. Drilling results from Union Oil Company's Baca Project

TOTAL WELLS: 19 original
6 redrill
25

FOOTAGE: 117,788 original
8,947 redrill (incomplete)

TOTAL COMMERCIAL WELLS: 5 (2 additional, mechanical failures)

TOTAL STEAM: 350,000 lbs/hr (15 MWe)

SUCCESS RATIO: 0.20

SUBTOTAL WELL COSTS: $15,739,000 (4 wells not included) program was designed to eliminate some of the risk by the financial participation of the federal government. The government had no influence over the exploration projects of the companies but was entitled to data generated by those projects. These data were then placed in the public domain to be used by the exploration community. The wells drilled under this program are listed in Table III-5 with the general statistics provided in Table III-6. The success ratio is a low 0.13. In defense of this low ratio it should be stated that these were comparatively high-risk wells, and all but about three could be considered rank wildcats. The data presented are probably a good indication of the probable success ratios of wildcat wells drilled within the Basin and Range province.

Since 1975 there have been four new geothermal fields discovered within the United States. In 1975 Phillips Petroleum Co. discovered Roosevelt Hot Springs system near the town of Milford, Utah. Sunedco discovered a geothermal system in Dixie Valley, Nevada in 1978. That same year McCullough Geothermal (now MCR Geothermal Corp.) and Geothermal Kinetics Inc. (GKI) discovered the South Brawley field in the Imperial Valley. -In 1979 Phillips Petroleum Co. discovered the Steamboat Hot Springs system south of Reno, Nevada. At the present time several fields are undergoing evaluation; these include Chevron's Beowawe project in Nevada and Phillip's Desert Peak project, also in Nevada.

B. Drilling Costs

Costs of drilling within the U.S. have been escalating at rates above the average annual rate of inflation. Chappell et al. (1979) have

Table III-5. Summary of holes drilled under DOE's Industry Coupled Program

Area	Company	Well	Depth	T Max	Status
Roosevelt Hot Springs, UT	Getty	52-21	7500'	398°F	Non-commercial
	GPC	GPC-15	1900'	162°F	T-gradient
Cove Fort/Sulphurdale, UT			42-7	7735'	344°FNon-commercial
		31-33	5221'	294°F	Non-commercial
		14-29	2620'	198°F	Abandoned
Beowawe, NV	Chevron	85-18	5927'	354°F	Producer
	Getty	76-17	9005'	336°F	Non-commercial
Dixie Valley, NV	Southland	45-14	9022'	385°F	Non-commercial
	Royalty	66-21	9780'	336°F	Non-commercial
Stillwater, NV	Union	Debraga #2	6946'	336°F	Non-commercial
		Richard			
		Weishaupt #1	10014'	353°F	Non-commercial
Humboldt House, NV	Phillips	Campbell	8061'	312°F	Non-commercial
		"E" No. 2			
Desert Peak, NV	Phillips	B-23-1	9641'	414°F	Future Producer
Soda Lake, NV	Chevron	63-33	2000'	297°F	T-gradient
		11-33	2000'	367°F	T-gradient
Colado, NV	Getty	44x-10	7965'	282°F	Non-commercial
Leach H.S., NV	Aminoil	11-36	8565'	260°F	Non-commercial
McCoy, NV	AMAX	14-7	2010'	140°F	T-gradient
		66-8	2510'	216°F	T-gradient
Tuscarora, NV	AMAX	66-5	5454'	225°F	Non-commercial

Table III-6. Drilling results from DOE's Industry Coupled Program

WELLS DRILLED: 20 (5 deep thermal gradient; 15 production)

FOOTAGE DRILLED: 124,723

PRODUCTIVE WELLS: 2

SUCCESS RATIO: 0.13

Table III-7. Correction factors for drilling costs.

YEAR	$/ft INDEX *	% INCREASE	1979 $ INDEX
1973	84.5		2.29
1974	100.0	+18.3	1.93
1975	118.2	+18.2	1.63
1976	128.2	+ 8.5	1.51
1977	144.9	+13.0	1.33
1978	165.9	+14.5	1.16
1979	193.1	+16.4	1.00
1980	204.1	+ 5.7	.95
1981	(231.7)	(+13.5)	(.83)

* Oil & Gas Journal (May 19, 1980; May 11, 1981)

( ) estimate equivalent to average of cost from 1973 to 1980 reviewed the costs of geothermal wells and inflation factors for those well costs. The inflation factors for geothermal drilling were found to be similar to those published for oil and gas drilling (Oil and Gas Journal, May, 1979), and since the oil and gas inflation factors are based on such a large data base, they were used to correct the drilling costs to constant dollars. We have continued this practice, and subsequent drilling costs will be stated in equivalent 1979 prices. The correction factors for drilling costs are shown in Table III-7. Surprisingly, data for 1980 shows a lower than normal increase of only 5.7%. Inflation factors for 1981 driling are estimated as being equivalent to the average increase of 13.5% of costs/foot over the past 7 years. The correction factors of Table III-7 have been applied to the costs of 50 geothermal wells drilled between 1973 and 1981. Figure III-1 shows well costs as a function of depth for those wells. These data are principally for production wells although two deep thermal gradient holes are also included. The data base is somewhat selective in that it does not contain data from wells drilled in The Geysers geothermal field or from the Imperial Valley. In addition, it does contain information from wells that were drilled for low- and intermediate-temperature direct heat applications (8 wells) as well as those that were drilled to develop electrical potential. We have attempted to fit regression lines to the data shown on Figure III-1 but have not achieved very satisfactory results. It is clear that there is a large variation in geothermal well costs. Many times this can be attributed to the resolve of a group to push on to a target objective in spite of extreme drilling conditions. In general, however, it is felt that Figure III-1 indicates the general range of expenditures that can be expected in the drilling

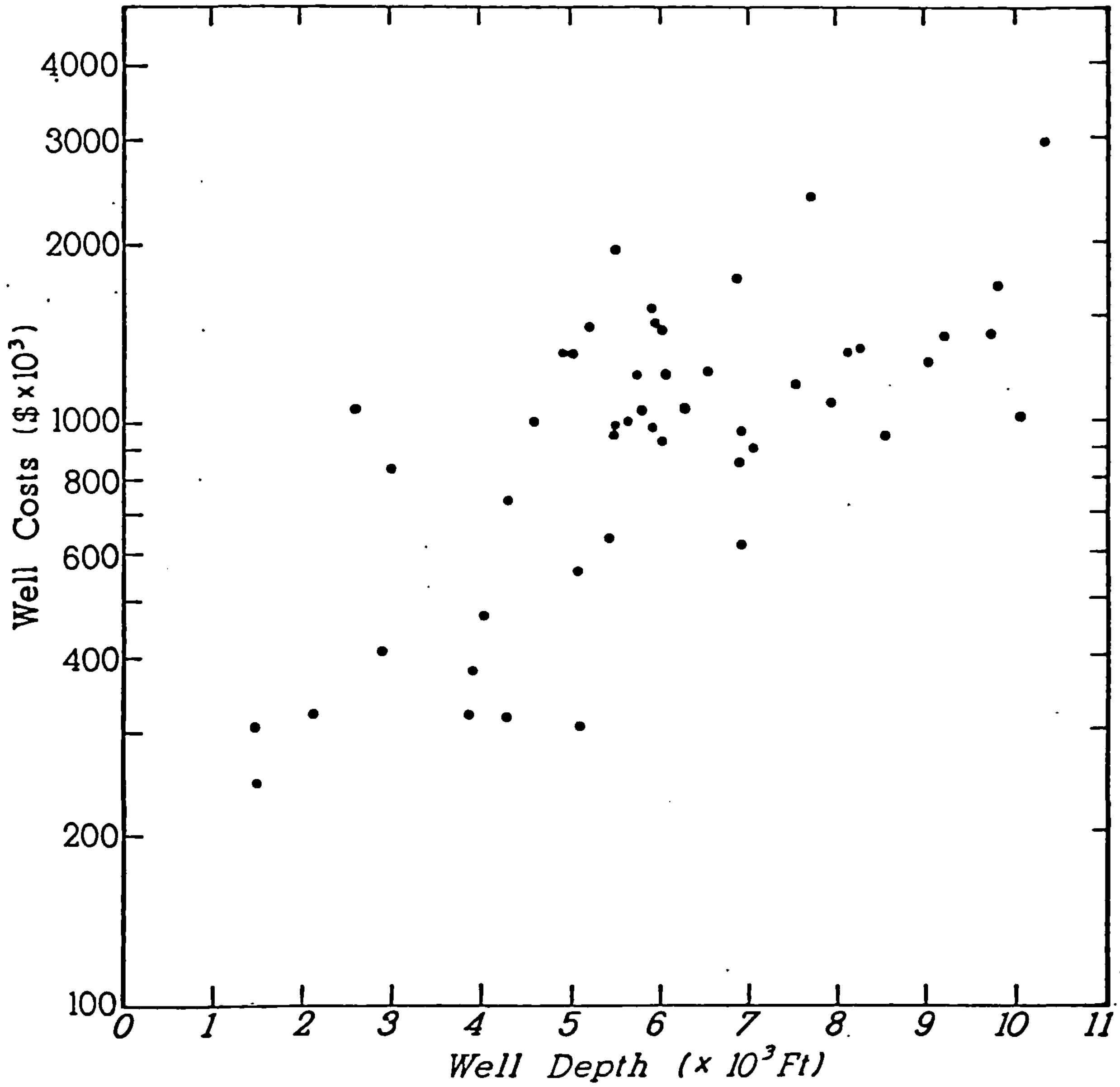

Figure III-1. Geothermal well costs in 1979 dollars as a function of depth.

of deep tests in geothermal prospect areas including The Geysers.

The same type of variation is found in drilling costs from individual project areas. Figure III-2 shows the cost per foot of wells drilled between 1973 and 1979 in the Baca project area of New Mexico. It can be seen that there is a large variation in costs but that this variation stays relatively constant although the costs are undergoing a steady increase due to inflation.

As was discussed in the section on exploration methods, it is standard procedure to drill thermal gradient holes during an exploration program. These holes are normally drilled by rotary methods. The depths are, of course, determined by local conditions, but 500 feet is an average during the initial stages of an exploration program. Companies presently budget about $10/foot for the drilling and preservation of these holes. Many companies also put a time limit on each hole to avoid high costs in the event that difficult drilling is encountered. For initial thermal gradient programs prior to the definition of a prospect area, it is common to set a limit of 500 feet or two days of drilling, which evercomes first.

Deep thermal gradient tests up to and exceeding 2000 feet are also commonly drilled during an exploration program. In the reconnaissance stage these are often drilled as a cooperative venture by several companies interested in the same district. In addition, they are often drilled in prospect areas prior to drilling a much more expensive production well. In general, the deep thermal gradient holes are drilled by rotary methods and preserved for future thermal measurements using either PVC or iron pipe. Oil and Gas Journal (June 1, 1981)

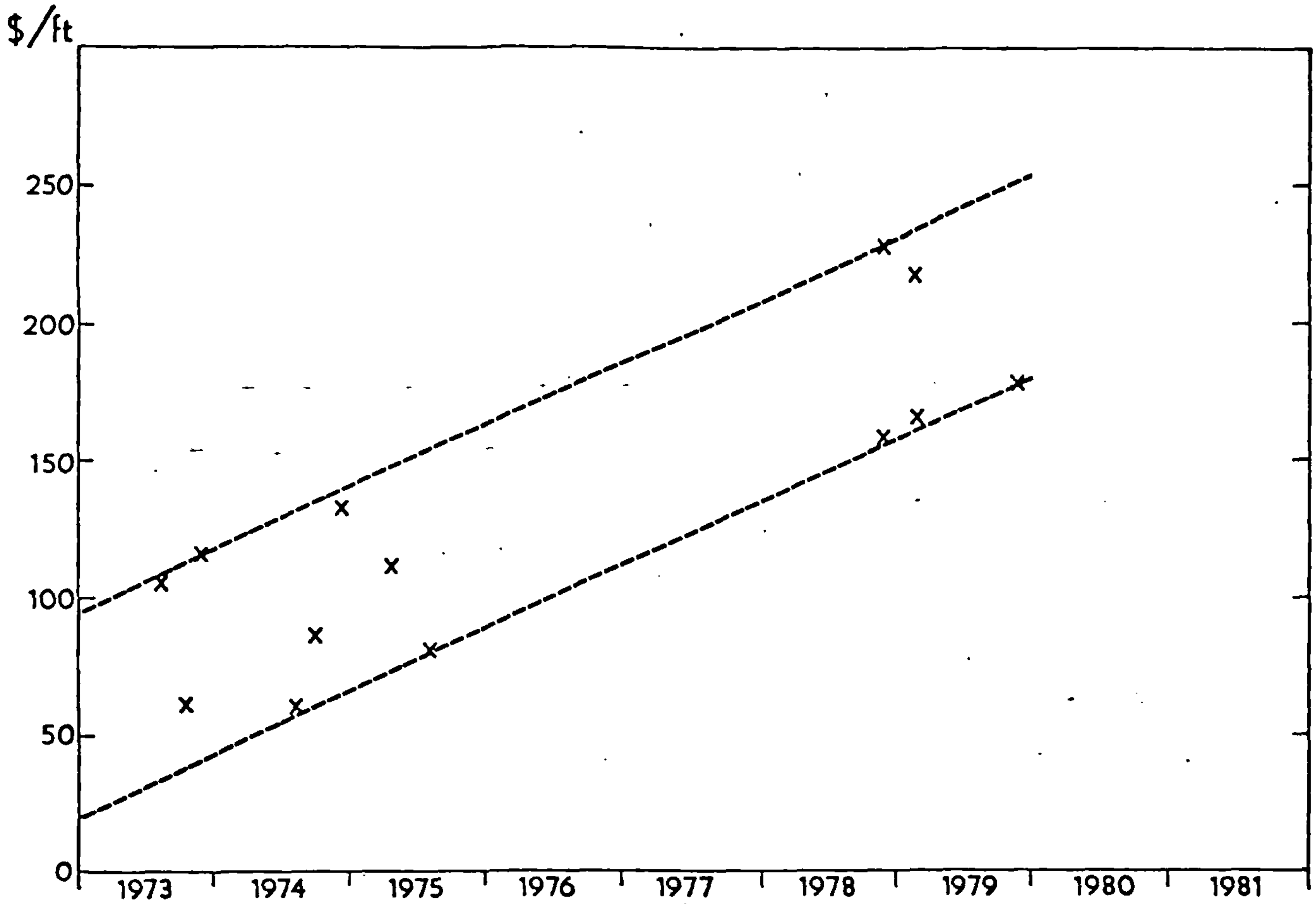

Figure III-2. Drilling costs of the Baca project, New Mexico.

RECOMMENDATIONS FOR RESEARCH

Research needs to be performed in several areas:

1) Establish the effects of pressure, temperature and chemical reactions upon permeability.
2) Characterize the fluid flow and diffusional pathways in crystalline rock such as tortuosity in microcracks and the degree of chemical equilibria between flow regimes and the host rock. Oxygen isotopes would be particularly useful in the latter and could be applied both to natural systems and laboratory experimental studies.
3) Develop improved indirect methods to determine permeability. It would be especially valuable to be able to utilize well logging methods to characterize fractures in terms of frequency, orientation and aperture.

BACKGROUND

Porosity and permeability are of fundamental importance to solution mining because they govern the fluid flow characteristics of the rock mass to a significant extent. Porosity is related to the fluid storage capacity of the system, and steady state fluid flow is simply a function of permeability and the pressure gradient. Accordingly, it is important to establish what the porosity and permeability are in potential targets for solution mining. Because these targets are often of difficult access, it is also important to know if porosity and permeability can be predicted, or if not, what kind of measurements are appropriate. Must porosity and permeability be measured in situ? Can they be measured or inferred remotely or by instrumental technique? Are laboratory measurements suitable?

EVALUATION OF STATE OF THE ART

Porosity

Formulations and models. Porosity is simply the ratio of pore volume to bulk volume. For crystalline rocks a convenient model is presented by Norton and Knapp (1977) whereby the total porosity, ϕ_T, is given by $\phi_T = \phi_F + \phi_D + \phi_R$ where ϕ_F is the flow porosity in which mass transport is dominated by fluid flow, ϕ_D is the diffusional porosity in which transport is by diffusion in the aqueous phase, and ϕ_R is the residual porosity composed of pores unconnected to ϕ_F or ϕ_D (Fig. 1.5-1). ϕ_F contributes significantly to the permeability of the rock along planar features such as joints, fractures or bedding planes, and ϕ_F = nd for any given parallel fracture array, where n is the fracture abundance (L^{-1}), and d is the fracture aperture. For the Mayflower pluton in the Park City district, Utah, $\phi_F = 2 \times 10^{-3}$ to 3×10^{-5} compared to ϕ_T of 0.01 to 0.03 (Villas, 1975; Villas and Norton, 1977). In the Sherman granite $\phi_F = 5 \times 10^{-6}$ and $\phi_T = 10^{-2}$ (Pratt et al., 1974). In general, ϕ_F constitutes a small proportion of the bulk total porosity; nonetheless, flow porosity is primarily responsible for the permeability of the rock mass.

Experimental evidence indicates that diffusion porosity in fractured crystalline rocks is on the order of 10^{-3} to 10^{-4} in rocks with total porosities of 10^{-2} to 10^{-1}. Because ϕ_F and ϕ_D are small, the bulk of the total porosity of crystalline rocks is in the residual porosity; $\phi_R > 0.9\phi_T$ (Norton and Knapp, 1977). This is significant for it means that more than 90% of crystalline bulk rock porosity is not accessible to fluids either through advective or diffusional transport.

Porosity Data. Davis (1969) has summarized porosity data for a wide range of rock types. The hydrologic characteristics of sedimentary rocks are so well known from research by the petroleum industry that they are not dealt with here; comprehensive studies are summarized by Davis (1969). Porosities of crystalline rocks are less well studied. Norton and Knapp (1977) have determined porosities on over 75 samples of fresh, altered and mineralized rocks. Unaltered igneous rocks should have finite porosity due to cooling effects. For particular geologic settings, total porosities exhibit moderate ranges: Chino, New Mexico skarns, 0.73×10^{-2} - 9.43×10^{-2}; San Manuel, Arizona altered quartz monzonite, 1.46×10^{-2} - 6.84×10^{-2}; Butte, Montana quartz monzonite, fresh = 1.01×10^{-2}, sericitized = 6.35×10^{-2}, argillized = 0.0753×10^{-2}; Bingham, Utah 2.64×10^{-2} - 12.5×10^{-2}; Ronda, Spain metamorphics, 0.66×10^{-2} - 8.42×10^{-2}. Norton and Knapp (1977) observed that the majority of the hydrothermally altered plutonic rocks and calc-silicates they studied had higher total porosities than their unaltered equivalents; the major component of the increased porosity is in larger residual porosity in the altered rocks. Fabric analyses indicate a correlation between mineral grain size and pore distribution in which pores occur predominately between grains and the more continuous pores are concentrated around larger mineral grains (Norton and Knapp, 1977).

Permeability.

Formulation and units. Permeability is a second rank tensor associated with a flux and a gradient:

$$q_i = -K_{ij} \frac{\partial P}{\partial X_j} \qquad \text{(Eq 1.5-1)}$$

where when $\partial P/\partial X_j$ is the pressure gradient, q_i is the volume rate of flow per unit time and K_{ij} are constants. This is an empirical relationship known as Darcy's Law. It has been shown experimentally that K_{ij} is a function of the characteristics of both the fluid and the medium. In order to separate the effects of the two and concentrate on characteristics of the medium, Darcy's

Law is often modified as follows (shown in scalar form):

$$q = \frac{k}{\mu} \left(\frac{\partial P}{\partial X}\right) \qquad \text{(Eq 1.5-2)}$$

where μ is the viscosity of the fluid, and k is the permeability with dimensions of L^2. For the purposes of this section we shall use the unit of the darcy (10^{-8} cm^2 ≈ 10^{-11} ft^2). Units of hydraulic conductivity also known as coefficient of permeability, common in ground-water studies, superimpose fluid properties on permeability and have dimensions of LT^{-1}, and 1 darcy ≈ 10^{-3} cm/sec = 10^{-5}m/s. In the U. S. the meinzer is a unit of hydraulic conductivity. For pure water at 15.6°C (60°F), 1 darcy = 18.2 meinzer = 18.2 gallon/day/ft^2.

Observations. There are three fundamental types of permeability determinations 1) laboratory measurements, 2) in situ measurements, and 3) inferred permeabilities from geological phenomena. Brace (1980) has summarized the various measurement techniques and compiled available permeability data on crystalline and argillaceous rocks. These data are illustrated in Fig. 1.5-2. The range in permeabilities spans more than 12 orders of magnitude, and wide ranges occur within similar rock types.

Laboratory measurements of granites and metamorphic rocks range from 10^{-4} to 10^{-12}d. In situ measurements vary from greater than 1 d to 10^{-10}d for granites and gneisses; in general k falls between 1μd and 100 md, and in situ values are often somewhat greater than laboratory values. Of particular significance is that in situ values in crystalline rock vary by over 4 orders of magnitude at a given site, and at most sites some volume of the rock has a permeability of >100 md.

Laboratory permeability determinations of well characterized crystalline rocks include the following: Westerly granite, 4×10^{-8}d (Heard et al., 1979), $3\text{-}5 \times 10^{-9}$d (Brace et al., 1968); Sherman granite, $40\text{-}100 \times 10^{-6}$d (Pratt et al., 1977); Climax granite, $< 10^{-9}$d (Ramspott, 1979); Barre granite, 10^{-6} - 10^{-7}d (Kranz et al., 1979); White Lake gneiss, 10^{-10} - 10^{-12}d (Heard et al., 1979). In situ measurements at the Stripa Mine, Sweden yield $k = 10^{-5}$ - 10^{-6}d for low fracture-density rock, and $k = 10^{-2}$d for well fractured rock (Witherspoon et al., 1979a,b). Variable fracture permeability undoubtedly contributes to the wide range in values of k for other in situ measurements.

Estimates of permeabilities in the deeper crust can be made from inferred permeabilities (Fig. 1.5-1C.). Permeabilities may be inferred either from fracture spacing and aperture analysis (discussed below) or from geological phenomena involving time and distance controlled predominantly by hydraulic diffusion (induced seismicity, for example). As expected, jointed and fractured rock have high permeabilities, $> 10^{2}$ - 10^{-2}d. Upper crustal seismic phenomena suggests k values of 10^{-1} - 10^{-4}d. Volcanic terranes will be heterogeneous in permeability; lithologies can vary from dense crystalline rock, highly fractured rock, porous but relatively impermeable pyroclastic material, and interbedded clastic sediments. The gross permeability will be anisotropic with greatest permeabilities parallel to volcanic flow directions. Oceanic crust, which may have significant through-going fractures, has inferred k of 10^{-1} - 10^{-3}d (Fenn and Cathles, 1979) to 4.5×10^{-4}d (Ribando et al., 1976). Inferred permeabilities for the Skaergaard intrusion (Norton and Taylor, 1979), which sample deeply into the crust, are 10^{-4} - 10^{-5}d in gabbro at 4-8 km, and 10^{-8}d in host gneiss at 7-10 km. Very low permeabilities (10^{-10}d) are necessary in order to attain pore pressures equivalent to lithostatic pressure at depth and to provide effective sealing in argillaceous rocks above deep oil and gas reservoirs.

Fracture Permeability. The common observation that laboratory measurements yield lower permeabilities than in situ measurements is generally attributed to fractures, faults, and joints in the rock mass not present in small laboratory samples. For example, the in situ permeability of the jointed Sherman granite was measured as 1.2×10^{-3}d compared to 10^{-4}d for an intact sample in the laboratory (Pratt et al., 1977).

In crystalline rocks in nature, fracture permeability is the dominant control on fluid flow. In the Skaergaard intrusion, for example, isotopic evidence demonstrates that fluid flow occurred primarily along fractures. Interstitial fluids maintained isotopic equilibrium by diffusion through microcracks to the fracture-flow system. Plagioclase $\delta^{18}O$ values are lowest near major fractures, indicating higher water/rock ratios (Taylor and Forester, 1979); $\delta^{18}O$ values increase away from major fracture zones.

Determination of the three-dimensional distribution of permeability in the target volume is important. The basic parameters of the natural fracture system that must be known are spacing, continuity, aperture and orientation. Mapping studies give good information on continuity and fracture system geometry, but do not define aperture well. In situ hydrologic experiments yield aperture information (Witherspoon et al., 1979a,b). When these are combined with fracture geometry directional permeabilities can be estimated. Fracture continuity is best determined by pressure measurements. Witherspoon et al., (1979a) have demonstrated that experimental laboratory measurements of fracture permeability on small diameter cores may be significantly less than in field tests. At the very least, they recommend that large sample sizes on the order of 1 m diameter be employed.

Data on the permeability of naturally fractured rock is sparse; in most cases estimates of permeability have been based on field observations combined with a simple planar fracture model. The model assumes that the fractures constitute a set of equidistant, parallel planes of infinite extent (Snow, 1968, 1970). For a single parallel plate opening, the volume rate of flow is a function of the aperture of the fracture, d, and the viscosity of the fluid, μ (Snow 1968, 1970; Norton and Knapp, 1977):

$$q = \frac{d^3}{12\mu} \left(\frac{\partial P}{\partial x}\right) \qquad \text{(Eq. 1.5-3)}$$

From (Eq. 1.5-2) and (Eq. 1.5-3) it follows that

$$k = \frac{nd^3}{12} \qquad \text{(Eq. 1.5-4)}$$

where n is the fracture abundance (in fractures-cm^{-1}). In ore bodies, fractures may be closely spaced. At Bingham, Utah, n = 0.5 fractures/cm, and in the Mayflower pluton, Park City, Utah, n = 6-21 fractures/m, with other stocks in the area ranging from 0.6 - 3.0 fractures/m (Villas, 1975; Villas and Norton, 1977; Norton and Knapp, 1977). Unaltered igneous bodies have n ≈ 0.1 fracture/m. Fracture apertures in the Mayflower pluton range from 40 to 270 μ m and average 145 μ m (Villas, 1975). These values yield wide ranges in permeabilities in the Mayflower Pluton from < 10^{-6} to 7d (Villas and Norton, 1977).

In addition to fracture geometry, fracture asperity or surface roughness contributes to permeability variation. A perfectly planar fracture may close completely upon application of normal stress, whereas surface roughness will affect permeability in two ways. Rough surfaces will inhibit closure of the fracture and will also change the path length or tortuosity of the flow path. Pratt et al. (1977) demonstrated that when a joint "closed" upon application of normal stress in situ, permeability remained 3 orders of arger than for the unjointed rock. In the laboratory, Kranz et al. erved a rapid decrease in permeability (from an initial 8×10^{-5}d) rre granite as fractures closed, the rate of which was a function of roughness. However, they achieved low permeabilities equivalent to actured rock (10^{-6} to 10^{-7}d) at confining pressures of 2 kb.

essure effects. The data compilation of Brace (1980, Fig. 1.5-3) ts a subtle decrease in permeability with depth; however it is not matic, and formulations in which permeability decreases exponentially depth (c.f. Fenn and Cathles, 1979, p. 249) are not generally icable. Experiments demonstrate a decrease in permeability with reasing pressure. The permeability of Westerly granite declined from 350×10^{-9}d at 100 b pressure to 4×10^{-9}d at 4000 b. The permeability change was losely correlated with electrical resistivity, and led to an extrapolation of $k = 0.5 \times 10^{-9}$d at 10 kb (Brace et al., 1968). Heard et al. (1979) produced a factor of two reduction in permeability in Westerly granite with applied pressure. However the permeability of the White Lake gneissic granite appeared to be unaffected by either pressure or differential stress, even though differential stress produced a factor of 2 variation in permeability in Westerly granite. Similarly, Witherspoon et al. (1979a) produced a decrease in hydraulic conductivity through a fracture with increased applied normal stress. In the Eleana argillite at the Nevada Test Site, permeability varies with a negative power of the effective pressure in experiments up to 24 MPa (approximately equivalent to 2 km depth), and permeability ranged from 10^{-11} to 10^{-8}d, depending upon the effective pressure. A single through-going fracture increased k by 3 to 4 orders of magnitude at low pressure and 1 to 2 orders of magnitude at 24 MPa (Lin, 1978).

Temperature effects. There is little data on the effects of temperature ı permeability, nor is there consistent agreement as to its effects. In situ ɘasurements on the Climax granite at the Nevada Test Site, at a depth of 425 , recorded a systematic decrease in permeability as the rock mass was heated nd an increase in k as temperature declined. Permeabilities were less than 1 ɿanodarcy, with a minimum of 0.02 nd at the highest temperature of approximately 325°C (Ramspott, 1979). The decrease in permeability is presumably due to closing of cracks accompanying thermal expansion of the rock mass. Heard (1980) heated Climax quartz monzonite to 300°C. In this instance the thermal expansion was accompanied by new microcrack formation which was inferred to enhance the permeability by a factor of 2 to 5. Laboratory measurements of Westerly granite (on cylinders 1.59 cm in diameter, 3.81 cm in length) demonstrate that the initial permeability at elevated temperatures is higher by 1 to 2 orders of magnitude than at room temperature. In this case, enhanced permeability is attributed to thermal stress cracking. However, permeabilities ultimately declined significantly due to dissolution and reprecipitation of plagioclase and quartz (Summers et al., 1978). Similar results were achieved by Morrow et al. (1981) in Westerly granite where permeability decreased by 1 to 2 orders of magnitude at elevated temperature, again as a result of dissolution and redeposition of quartz and feldspar along fractures.

INTRODUCTION

The flow of thermal energy from the earth's interior drives many of the geologic processes that have shaped and reshaped the earth throughout geologic time. Our understanding of the interrelationships among thermally driven geologic processes and the formation and subsequent evolution of natural resources has advanced greatly in the past twenty years and continues to do so today. Just a few short years ago, active hydrothermal systems located on the sea floor in close proximity to spreading centers were discovered and were found to be precipitating sulfide and other minerals in large quantity. These so-called "ocean smokers" are now being related to certain of the massive sulfide orebodies whose exploitation has been underway for centuries and toward whose discovery billions of dollars have spent worldwide. It is also becoming evident that these oceanic hydrothermal systems exert a major control on the chemistry of the oceans, contrary to the ideas held until recently that the chemical input from rivers was the main factor of significance. On land, many disseminated and other mineral deposits are recognized as the end products of past hydrothermal processes, and by analogy we assume that in today's active hydrothermal systems ore-forming processes are underway. In the formation of petroleum resources, too, the earth's heat is recognized as a key ingredient, and studies of maturation, encompassing the thermal history of a prospecting area, and how it has influenced extraction and migration of hydrocarbons from source rocks into reservoirs, give important clues in oil exploration. There is little doubt that research on the earth's thssing the thermal history of a prospecting area, and how it has influenced extraction and migration of hydrocarbons from source rocks into reservoirs, give important clues in oil exploration. There is little doubt that research on the earth's thermal regime and on geothermal processes will continue to contribute substantially both to our basic understanding of the earth and to our ability to discover and develop a variety of natural resources.

A significant amount of recoverable energy is to be found in present-day hydrothermal systems, but, except at a few of the highest-grade, best located resourcs, generation of electricity from hydrothermal energy and direct uses are not yet economically viable. The fledgling geothermal industry competes today with cheaper fossil energy costs, but we believe that this situation will not always be so. In spite of currently falling oil prices, in the long term fossil energy costs must rise, while at the same time significant advances will be made in more efficient and less expensive ways to discover and develop geothermal resources. We believe that in years to come, geothermal energy will make an increasingly significant contribution to the world's energy needs.

It is worth noting that in the United States a majority of the geothermal resource and development work in the earth sciences has been federally sponsored by the National Science Foundation (until 1975), the Energy Research and Development Agency (until 1979) and the Department of Energy (to the present). Published reports from the Idaho National Engineering Laboratory, Lawrence Berkeley Laboratory, Lawrence Livermore National Laboratory, Sandia National Laboratories, Los Alamos National Laboratory, and Battelle Pacific Northwest Laboratory as well as from other contractors who operate primarily through the Idaho and San Francisco Operations Offices of the U.S. Department of Energy contain a wealth of public domain data and results, a significant amount of which has not appeared in journals. In addition, the U.S. Geological Survey has maintained an active program of research and regional assessment of the geothermal resource base in the U.S. (White and Williams, 1975; Muffler, 1979; Reed, 1983).

In substantially all of the rest of the world, geothermal development is sponsored by federal governments, and there is reasonably good access to data, although not all of it appears in journals. Active programs in geothermal exploration and development are being carried out in China, El Salvador, France, Iceland, Indonesia, Italy, Japan, Kenya, Mexico, New Zealand, Philippines, and to a lesser extent in other countries. Expertise arising from first-hand experience in Iceland, Italy, New Zealand, Mexico, Japan, and the U.S., primarily, is being used by more underdeveloped countries to assist their geothermal efforts. The United Nations sponsors both scientific work and education in these countries. La Organizacion Latinoamericana de Energía (OLADE), headquartered in Quito, Ecuador, provides coordination and support for geothermal development in Central and South America. In short, an infrastructure for geothermal development is being built throughout the world, and, while it is small compared to the corresponding petroleum or minerals infrastructures, it is making important contributions.

NATURE AND OCCURRENCE OF GEOTHERMAL ENERGY

Because the earth is hot inside, heat flows steadily outward from the surface, where it is lost permanently by radiation into space. The mean value of this heat loss is estimated to be about 10.2×10^{12} cal/sec (Williams and Von Herzen, 1974) or about 42.6 million megawatts. At present, only a small portion of this heat, that concentrated in geothermal resources, can be captured for man's benefit.

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 decay of radioactive elements in rocks; 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. Some theoretical models of the earth indicate that heat produced by radioactive decay can account for essentially all of the present heat flux (MacDonald, 1965). Other work (Davis, 1980) indicates 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. However, the distribution of radioactive elements within the earth is poorly known, as is the earth's early formational history, and the relative contribution to the observed surface heat flow of these two mechanisms is not yet fully resolved.

Geothermal Resource Types

Geothermal resources have three common components:

1) a heat source
2) permeable rock, and

the near-surface rocks are permeable and the water table low, such as in the Basin and Range province, much of this water may dischrge into these near-surface rocks and never reach the surface. White (1968) indicates that about 95% of the total discharge of the Steamboat Springs, Nevada thermal system may escape into the alluvial aquifers.

2. Heat Sources

The heat sources for hydrothermal convective systems are either magmatic sources or the thermal gradient of the earth. Smith and Shaw (1975) have reviewed the data from systems which are thought to derive their heat from igneous systems. Although these authors have proposed some theories that they admit are based on speculative data, the theories have held up quite well over the years since their paper was written. They have concluded that geothermal systems associated with felsic igneous systems which are younger than one million years have a high potential for being high temperature. This is because the granitic plutons that provide heat for these systems are able to reside within the upper 10 km of the crust. Evidence indicates that the more mafic magmas undergo rapid transport from their area of generation and are more likely to be erupted at the surface as flows rather than forming high-level plutonic bodies. Thus one exploration criteria for high-temperature geothermal systems is the association with volcanics of less than one million years. Examples of such systems are the calderas at Yellowstone, Valles, and Long Valley as well as Coso, California, Roosevelt Hot Springs, Utah, and Steamboat, Nevada. The andesitic systems of the Cascades province deserve special consideration because, even though exploration is just beginning, it is evident that high level plutons can exist, and with the possible exception of Meager Mountain system in British Columbia, there have been no discoveries in the province. However, a number of companies are involved in initial exploration programs. In addition, young mafic provinces are of high potential if they show evidences of differentiation to felsic end members. Examples of this would be the Imperial Valley which is thought to be located above a segment of the East Pacific Rise which has been overridden by the continent (Elders, 1979; Robinson et al., 1976).

Hydrothermal fluids are also warmed by the natural geothermal gradient of the earth. Gradients in much of the U.S. are in the range of 10°-30°C/km. However, average gradients in the Basin and Range and Cascade provinces are more like 45°-60°C/km. One of the more commonly used geothermal exploration tools is the measurement of thermal gradients in drill holes in order to calculate heat flow and identify areas of anomalous temperatures for further exploration. These methods will be discussed in more detail in a subsequent section.

Most explored hydrothermal systems have been found to reach a 'base temperature' which is characteristic of the system. This is a maximum temperature reached in drilling of the system, and temperatures do not increase above this level as greater depths are explored. The fluids have thus achieved this maximum temperature at depth and have then flowed toward the surface with little subsequent heat loss. The base temperatures in both Wairakei, New Zealand and Roosevelt Hot Springs system are approximately 260°C.

DRILLING, LOGGING, COMPLETION AND WELL SIMULATION

Introduction

Geothermal drilling is generally done with somewhat modified, conventional drilling equipment. For wells that are expected to produce large quantities of hot water and/or steam for electric power production, large rotary drilling rigs of the type used for oil and gas wells are used, but if smaller quantities of lower temperature fluids for use in space heating are sought, a conventional water well rig might suffice. Because drilling is one of the most expensive steps in geothermal development and at the same time is typically frought with problems and setbacks it is important to choose the correct drilling contractor, equipment and techniques for the job at hand and not to compromise the drilling program too much for the sake of attempting to save money. One will usually come out ahead cost-wise by going at this job right in the first place rather than skimping and then running into expensive and time consuming difficulty. Even drilling contractors well experienced in other types of drilling will have trouble at first with geothermal drilling because there is not a direct transfer of oil field or water well or mining technology to the geothermal drilling operation. The geothermal environment is characterized by temperatures above those in which most drilling equipment and muds were designed to operate. In addition, geothermal brines are especially corrosive due not only to their chemical composition but also to their elevated temperature. Special equipment and procedures are necessary to ensure the safety of the drilling crew and to minimize the possibility of blow-out. Besides all of these extra considerations, a good geothermal environment will have had a history of complex geological and geochemical processes that vastly increase the chances for difficulty during drilling attributable to zones of lost circulation, high angle faults or fractures that deflect the hole, or caving or sloughing of wall rock into the borehole. A wise developer will examine the prospective contractor's geothermal drilling credentials carefully before choosing his drilling team.

Overview of the Drilling Operation

The drilling machine proper, including the derrick or mast and the power unit which rotates the drill stem is termed the rig. It is supported by pumps and other equipment for circulating mud or water into the well. At the bottom of the hole a rotating bit breaks and abraids the rock, producing rock chips or cuttings. The bit is attached to the drill stem which is made from lengths of pipe that are provided with rotary power at the surface, on the floor of the drill rig. At the top of the drill string is a swivel connection to a high pressure hose through which drill mud, water or air is forced down the drill string and out through nozzels in the bit. Circulation of drilling fluid is essential to carry the cuttings away up the annulus between the drill string and the casing and to cool and lubricate the bit. The entire drill string is raised or lowered in the hole by draworks on the drill rig floor which operate cables that go over the crown block at the top of the derrick. This capability is needed in order to add length of drill pipe as the hole is deepened and to remove the string from the hole when the bit is to be replaced.

Collapse of the well is prevented by steel casing which lines the hole and is tightly cemented into place. The open, or uncased bottom portion of the hole may or may not need support to prevent sloughing or collapse and such support as is needed may be given during by the circulating, mud which has a density and therefore a pressure at the bottom of the hole considerably greater than that of plain water. At the upper end of the hole, a largedimater surface casing is firmly set in place with cement to bear the weight of any uncemented casing below and to provide a firm anchor. Below the surface casing and extending to the surface inside this casing is generally an intermediate casing, also carefully cemented in place. Blow out preventers are attached to the portion of the intermediate casing that projects above the surface. Their purpose is to close tightly around the drill string to prevent blow out of hot water or steam if a high pressure zone is unexpectedly encountered. The surface wellhead equipment, including blowout preventers is sometimes set underground in a cellar below the drilling floor.

For simpler drilling operations where fluid temperatures below the surface boiling temperature are expected, certain equipment such as the blow out preventers are not necessary, and the entire rig is generally smaller and less complex.

Let us now consider this process in more detail.

Selection of Drilling Equipment

Although other factors may be important, there are three primary, interrelated considerations in selecting the drilling equipment. These are: (1) anticipated depth of drilling, (2) planned well diameter, and (3) nature and severity of problems anticipated.

The primary choice has to do with the size of the rig. Deeper and/or larger diameter wells require larger rigs, as shown approximately by Table 1, and these larger rigs are more expensive to operate on a daily basis. It is important at the outset not to undersize the rig for a given operation to save costs because when trouble occurs (factor 3), extra rig capacity will inevitably be needed.

The well diameter is a function of the productivity to be expected from the well inorder to meet the how water or steam requirements of the utilization system. Unlike petroleum wells which need produce only a few gallons per hour of crude oil, geothermal wells in an electrical power application may be required to transmit 100 tons per hour of fluid, which is equivalent to about 3300 pounds or 450 gallons per minute. Obviously large diameter is important to reduce fractional losses in the pipe and to keep steam velocities subsonic. In addition, the planned depth of drilling affects well diameter. It is wise to start deep wells at large diameter so that if difficulty is encountered the well may be cased and continued at smaller diameter. Such casing with attendant reduction in diameter may be required several times during the drilling of a well.

Specification of the depth, diameter and rig size determines, within limits, the capacity of the mud pumps, size of the mud pit, compressors (for air drilling), etc.

Drill Site Preparation

Preparation of te drill site prior to bringing in the rig can be a significant expense. In the first place, the rig will require a leveled site, which could be a major undertaking in mountainous terrain. in addition to the psace the rig itself will occupy, there must be adequate cleared and leveled area around the rig for (1) stcking of drill pipe and casing, (2) constructing the mud pit, a pond of sufficient capacity to hold the volume of mud plus reserve required for the planned mud well depth and diameter and of sufficient surface area to provide the possibility of some cooling of the mud, (3) packing of the mud pumps and compressors if our drilling is to be ___, and (4) providing adequate room for vehicles, a doghouse and for working around the rig and its auxillary equipment.

In a simple water well rig application, the drill site may require only an area of, say, 15 ft by 30 ft, but for a large diameter, deep well for electric power application, the drill site may encompass 5 acres. Depending n environmental restrictions the site may have to be restored after the well is completed and the mud pit may have to be lined during drilling to prevent leakage of fluid into the ground water system.

We have discribed here an expensive site preparation procedure. Little wonder that geothermal drillers are turning more and more to directional drilling of two or more holes from the same drill pad to save site preparation cots.

Drill String and Bits

The drill string or pipe is made of steel in 10-foot sections for a small rig and in larger sections, up to 40 feet for a large rig. Joints are probably of the interval flush-butt type (Fig. 2). The rotary drill bit is usually of the tricone type (Fig. 3), and is somewhat different from the typical petroleum bit because harder rock, higher temperatures and more corrosive formation fluids are usually encountered in geothermal drilling. The driller attempts to maximize bit life because changing bits costs time and money. The cost of a new bit is small compared to the cost of lost drilling time to pull the entire drill string from the hole to replace the bit. Bit pressure and rotation speed are the principal factors that are controlled by the driller to optimize bit life and rate of progress.

Drilling Fluids and Loss of Their Circulation

Some type of fluid must be circulated to the bit and back out of the hole in order to remove cuttings. Otherwise the bit and drill string would soon foul and bind. The drill fluid also performs the important function of cooling and lubricating the bit. It has become common practice in oil field drilling to use especially compounded drilling muds that have a density greater than water to create a bottomhole mud pressure that exceeds hydrostatic pressure in the adjacent rock. This helps to prevent blowout of high pressure fluids as well as sloughing of rock from the sides of the borehole. Sloughing is also inhibited by the mud cake, a coating of the solid components of the mud that adheres to the walls of the hole. Indeed, development of mud technology over the past 30 years has revolutionized petroleum drilling. But use of mud in geothermal drilling can be a mixed blessing. On the one hand it undoubtedly helps prevent sloughing and blowout some problems, but on the other hand severe or irreparable damage can be done to the production capability of a well if mud penetrates into permeable zones and solidifies. It is to prevent such mud damage that techniques for drilling with air and foam have been developed in some goethermal fields. Nevertheless the extra measure of control that mud drilling affords is attractive, and so high temperature muds have been and continue to be developed. Conventional clay based drilling muds lose stability above about 150°C. High temperature stability of muds can be increased by adding a blend of chrome-lignite/chrom-lignosulphronate compounds.

Minimal well damage can be expected if the formation is competent enough to allow drilling with air as the circulating fluid. However, very high pressure and volume of air are required to remove the rock cuttings adequately. A velocity of 2000-5000 ft/min is usually maintained up the annulus of the hole. As might be imagined, this leads to abraision of the

drill string and casing this is much more serious than it is with mud drilling. To maintain air flow compressor capacity must be increased significantly as the hole deepens.

Drilling with air will not work at all in situations where a significant amount of water enters the hole. If the amount of water entering is the order of 2-3 gallons per minute, it may be possible to add a foaming agent and continue drilling, but if water flow exceeds 5 gallons per minute the only remedy is to seal off the zones of water production by casing or to switch to mud.

A common problem in drilling is to lose circulation of the drilling fluid (mud, foam or air), e.g. a significant portion of the fluid enters the formation and does not return to the surface. When this happens, the drill cuttings settle around the zone of lost circulation due to the abrupt decrease in fluid velocity at that point and the drill string and possibly the bit become bound in. At first sign of significant lost circulation, the driller will first attempt to plug the formation with any organic material available that can be shredded and pumped down the hole. Hay, cotton, nut shells, sawdust and many other materials have been used. Organic materials have the advantage in geothermal drilling that they first swell inthe formation, but later degrade at hgh temperature to allow production from the permeable zone. If circulation can thus be restored, drilling can continue. If not, the zone must be cased off.

On occasion the drill stem becomes stuck due to caving or to the results of lost circulation. Techniques for freeing the pipes include circulating a lubricant into the hole and application of a great deal of torque and/vertical pull to the pipe. Under such circumstances the pipe sometimes breaks, and a "twist off" is said to have occurred. There are techniques for reconnecting to the broken off section of pipe. It may be necessary to disconnect and retrieve the drill stem length by length until the stuck section is located. This section may be explosively freed.

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 reponse 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.

A second, important problem in geothermal well logging is general lack of probes that will work in an environment where temperature exceed 100-150°C. Electric components have generally not been available for temperatures this high and indeed have not generally been required by the petroleum industry, which makes the most use by far of well logging. Sandia National Laboratories in the USA has had an active and successful research program sponsored by the

U.S. Department of Energy, to develop electronic components and logging tools for use in the geothermal environment, and so appropriate logging equipment is now becoming available.

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 performedfor cased intervals. It is common practice not to repeat logs in sections of the well that have been previously logged, but simly to provide adequate overlap with the previous logging run to facilitate _____ of logs that may be made with different instruments and different calibrations on successive logging runs.

Few developers or drilling contrctors offer logging services themselves. Geophysycal 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 1.6-1 is given a brief summary of logs that have been applied to geothermal well logging, and a brief explanation of these logs follows herewith.

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 automatically during the logging process. Three- or four-arm cliper 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 $\pm$ 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, 1979). As an example of the utility of these plots, bulk density is plotted against neutron porosity in Figure 1.6-G 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, 1979). 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.

RECOMMENDED SAMPLING PROCEDURES

ROTARY DRILLING (see Hills, 1949; Low, 1951; Morris, 1968)

It is the job of the geoscientist to specify 1) sampling interval, 2) sample size, 3) collection method, 4) washing procedures, and 5) packaging and labeling specifications. In addition, there are a number of subsidiary considerations:

Sampling Interval

Grab samples at 3-meter intervals are recommended, with tighter sampling as dictated by frequency of geologic changes for shallow mud-drilled holes or for air-drilled holes, where transit time up the hole is short. Shorter sampling intervals are not justified if transit time is more than 20 minutes unless bit progress is less than 3 meters per hour and the hole is relatively free from caved or fractured zones where cuttings become temporarily hung-up and mixed.

Frequent grab samples are generally more useful than continuous samples for recognition and definition of complex lithologic or alteration variations and for the detection of sporadic caving and other types of sample contamination. Individual grab samples allow the option of subsequent compositing.

Sample Size

A minimum of 500 gm of cuttings should becollected. When lost circulation material is mixed with the mud, the size of thesamples caught must be increased to maintain minimum cuttings sample size. Sufficient sample must be available to allow a number of splits for various research purposes.

Collection Method

Rigs drilling with mud will move the returning mud over a shale shaker and through a sand filter to remove cuttings before recirculaing the mud to the hole. For hard rock the shale shaker will remove almost all of the cuttings, and an adequate sample can be caught as a grab from the shale shaker or from a bucket held in the return mud stream at each sampling interval. In softer materials the sand filter may be sampled also. Clays are usually lost from the sample because they mix with the mud and pass through shaker and filters. If high clay content is suspected, mud samples should be taken from the return mud stream before the shale shaker. These mud samples can be compared to samples of virgin mud and additives to determine extent of clay meterials introduced from the rock at depth.

For rigs drilling on air, cuttings are blown from the hole as a fine sand. Although large chips may be cut by the drill bit, abrasion during turbulent passage up the hole reduces their size. Caved material can often be recognized by its larger size. Grab samples from the cuttings stream are caught in the usual manner.

Washing Procedures

If the drill mud is oil-based, the cuttings must be washed immediately. Otherwise later removal of the oil is very difficult. For water-based drill muds, the geoscientist has an option. Several factors must be considered. Once an unwashed sample has dried, it is difficult to disarticulate and clean. In addition, lost circulation material can be rejected immediately if drill site washing is done. However, washing under laboratory conditions rather than at the drill site has control advantages. The drill site geoscientist must decide what is best for each situation.

Samples are washed by placing in a bucket and decanting the drill mud and very fine particles with water. Lost circulation material such as walnut shells, cottonseed hulls, a shredded cellophane often float off with this washing. Washing is usually continued until the water is clean. The samples are then dried, either naturally or by forced drying. For forced drying, use low heat (150°F) to prevent alterations to the sample. Screening of the samples after drying is not recommended.

It is strongly recommended that an unwashed reference sample be taken at each interval even though themain sample is washed at the drill site.

Packaging and Labeling

Unwashed samples should be placed in heavy plastic bags, tagged properly, and sealed as well as possible. Unwashed samples which arrive at the laboratory still moist are easier to trat subsequently because the drill mud has not dried to a hard cake. An exterior tag and an interior tag are both recommended for each sample. Washed, dried samples can be shipped in clean fabric bags.

The label should show hole number, hole location, date, footage represented and the sampler's intitials.

Subsidiary Considerations

1. Air-rotary samples return to the surface in a matter of a few minutes even from 3,000 meter holes. The samples can thus be considered torepresent the depth reached by the drill string. Mud-rotary samples

take much longer to return to the surface. A rough rule of thumb is 1 minute of return time for each 30 m of hole depth. The lat time should be measured by the site geoscientist at each opportunity and recorded (see Hills, 1949). Correction for lag can be made later.

2. The site geoscientist should obtain copies of all driller's logs, mud logs, mud additive substances, times, amounts, and data on penetration rate and bit weight.

3. Samples of all drill fluids should be periodically obtained for correlation with cuttings samples.

CORE DRILLING

Cores may be obtained either as a core run in a rotary hole or in an all-cored hole. Procedures for drill-site handling are the same.

The core barrel is emptied by the drill crew by shaking, tapping and flushing. The core is then transferred to boxes. The boxes are labeled and the core is turned over to the site geoscientist.

The geoscientist should estimate percentage core recovery by measuring the actual core and comparing to the footage drilled. The core should be washed carefully (exceptions noted below). A quick geologic log should be made by recording rock type, planar features alteration, mineralization and other important features on a standard log form (see Appendix). Note should be made of ground ends, overdrilled pieces, and of washouts and rubbelized zones, and the driller should be appraised of these. The core may be photographed box by box. The hole number, core top, bottom, and footage should be indicated clearly in the photographs by appropriately placed and marked 3" x 5" cards. The core should neither be split nor sampled at the drill site.

Some studies, such as core fluid analysis, require special handling and preservation of the core. To preserve the original fluids requires that the core be sealed. The core should be wiped clean, not washed. All pieces should be marked with footage and an arrow pointing to the top of the hole. Short pieces may be wrapped in plastic and saled with fiber tape. Longer pieces should be placed in polyethylene tubes. The ends of the tube should be heat sealed or folded several times and taped tightly closed. This method of preservation is good for a few days of storage. Longer term storage requires that the core and plastic tubing be sealed with hot parraffin. The plastic wrapped sample must first be wrpped in aluminum foil to protect the plastic from the heat of the wax and to facilitate unwrapping the sample inthe laboratory. A 3" x 5" card with thehole number and saple depth should be taped to the outside of the foil before applying the was. The sample is then dipped in hot wax until a coating approximately 1/4 inch tick covers the entire sample. These samples may be shipped in regular core boxes if the coreis carefully packed in foam rubber sothat the seal is not broken during shipment.

If the core will be used for fracture intensity studies, split-tube core barrels should be used. These core barrels are held together with clips which allowhalf of the tube to be lifted off, exposing the core. The core may be turned into a trough the length of the core barrel. The core is then washed or wiped and the fractures counted and measured. It is advisable to photograph the core in the trough.

<u>CABLE TOOL DRILLING</u> (see Hills, 1949, Low, 1951)

A grab sample of 500-1,000 gm should be collected from the first bailer after each run of the bit (Hills, 1949). Decisions regarding drill-site washing of samples and packaging and labeling of the same as for rotary drilling. The site geoscientist has little control over sampling interval -- samples are available only when the hole is bailed.

Well Completion

Well completion is the process of placing final casing in the well, cleaning and conditioning the well to achieve maximum productivity, and installing well head equipment to control the flow of produced fluids.

Once a producing horizon has been reached, and drilling has been terminated, there are decisions to be taken on how to complete the well. In some areas of competent country rock, open hole completion may be elected, that is no casing will be installed in the production zone. If it is anticipated that casing might occur and that the production zone will consequently require support, an appropriate, perforated casing will be installed. One common option is to use _____ casing down to the top of the production zone and then to install stalled liner, a casing that has narrow, long stal parallel to its axis through which fluids may flow into the well, below that, in the production zone. The stalled liner is normally hung free from the bottom of the solid casing so that it can be easily removed if the well requires cleaning or stimulation.

Prior to installation of slotted liner, mud is circulated in the well until all drill chips are removed, as there are usually of sufficient size to plug the openings in the liner. After liner installation, clear water is circulated into the well to replace the mud.

Meanwhile the well-head is equipped with an appropriate control valve and an initial flow test may be performed on the well, at reduced volume with the rig in place. If this flow test meets minimum expectations and if the well system appears to meet engineering psecifications, the rif will be removed the blow-out preventers and other well-head equipment will be replaced with a "Christmas Tree" consisting of a drain control valve, a secondary control valve and severalsampling valves and ports for taking fluid samples and measuring well-head pressure and temperature. At this point the well is ready for its first full-scale flow test.

Some wells will flow under artesian pressure when the valves are opened at the well-head. Other wells may not have artesian pressure and may indeed be inhibited by the column of relatively cool water that sites in the casing above the production zones. For these latter wells, a gas lift may be attempted. A small-diameter metal tube is placed down the well to a level at least several hundred feet below the level of standing water in the well. Through this tube is bubbled a relatively small quantity of gas. Bottled nitrogen is often used because it is easily bought and transported in cylinders, is available at high pressure and is inert, thereby not contributing to the chemical reactivity of the fluids in the well. The high pressure air compressors associated with the drilling rig could be used, but have the significant disadvantage of introducing oxygen into the well.

At any rate, as the introduced gas bubbles upward it expands and forces water upward and out of the well. this decreases the hydrostatic pressure on the production zones at depth, which are thereby induced to flow. If the temperature of the geothermal water is high enough, the reduction in pressure may cause some of the water to flash to steam in the well bore, and production of a steam-water mixture at a high rate might ensue.

The initial high-volume production from a well is usually accompanied by a great deal of rock debris that varies from dust or sand in size up to large climbs. This debris can cause very rapid erosion of well casing and surface pipe, especially where there are abrupt changes in diameter or bends in surface pipe. There is also danger that the flow control volumes at the surface might be damaged to the point of not being able to close off flow. A source of cold water and a means of introducing it into the well should be kept at the rady in case it is needed to hill the flow from the well

Stimulation

A number of methods have been very successfully developed for increasing the production rate of crude oil or gas from a well by the petroleum industry. There has been a great deal of speculation that geothermal wells could also benefit from application of these techniques or from modifications suited to the geothermal reservoir. Several research field experiments have actually been conducted that indicate that indeed benefits may be derived in some cases but much remains to be learned.

In a geothermal well, low production rates may be due to one or more of three principal causes (assuming that a reservoir of high potential production rate exists and that the borehole has simply failed on some account to yield the latest production):

1. Well bore damage (skin damage). This is not commonly the result of plugging of permeable zones, by lost circulation, plugging material or by drill muds.
2. Low formation permeability.
3. Failure of the borehole to intersect a fracture that is part of the interconnected reservoir system.

Well stimulation techniques are designed to reduce the restrictions to flow from the reservoir into the well bore. These methods include both physical stimulation techniques such as hydraulic fracturing and the use of explosives, and chemical stimulation techniques such as acid treatment.

To date the most promising technique for stimulating a geothermal well appears to be the hydraulic fracturing technique. In a hydraulic fracturing treatment, fluid is injected down the well casing or tubing at a rate higher than the reservoir will accept. This rapid injection produces a buildup in well bore pressure until a pressure large enough to overcome compressive earth stresses and the tensile strength of the rock is reached. In the majority of cases the stress regime in the earth is such that a vertical or near-vertical fracture having a shape illustrated schematically in Figure 5 well form. Continuous fluid injection increases the fracture length and width.

In order to achieve significant stimulation, the fracture conductivity (permeability times width) after the well is returned to production must be larger than that allowed by natural reservoir permeability. To obtain high conductivity a granular, solid propping agent, such as sand, is injected along with the fracturing fluid and is deposited within the fracture. The proppant must be strong enough to maintain a high permeability when subjected to the compressive earth stresses (closure stresses) once pressure is removed from the hydraulic fluid.

Under good circumstances a propped fracture can mitigate each of te principal causes of low productivity. If there is a zone of formation damage surrounding the well bore, the high-conductivity path provided by the fracture by passes the damaged zone. Since damaged zones are generally believed to extend only a few feet into the formation, the required size of racture is not great. If the high-conductivity path extends far enough into the formation, the basic flow pttern within thereservoir may be altered as in Figure 6. This new liner flow pattern can result in a many-fold increase in productivity. Also if the hydraulic fracture succeeds in intersecting a fracture that is part of the interconnected network forming the reservoir, a substantial increase in productivity can be expected.

Hydrofracturing techniques include the so called "planar frac" and the "_____" or "keil frac". In the planar frac technique there is basically one application of fracturing fluid and proppant. After fracturing is completed, usually in a few hours, production from the well may be slowly started. The fracture that is created is the one schematically shaven in Figure 7.

A quite different concept is used in Keil technique. The treatment is usually designed to use the highest possible flow rate that the equipment (pumps, well head equipment, casing, etc.) will sustain. As slick fluid, created by proper additives to a water based fracturing fluid is used to minimize tubular friction losses. Slugs of fine and coarse proppants are injected throughout the several stages of the treatment. As few as two or as many as 10 stages comprise the Kiel frac.

One stage consists of a pod of clear fluid followed by a slug of fine sand, followed by another pod to displace the sand into the formation. Another slug of fine sand is followed by a pod and then a slug of coarse sand and a pod. Then a shut-down, flow-book plied is used to allow the formation to close and to slough off into the fracture as stress is released. After a brief rest, injection is restarted and the spalled material moved outward where it blocks the leading edge of the fracture. A short shut down period follows to terminte the first stage. This sequence is repeated as specified. Under good conditions the second and subsequent stages will each form new fractures rather than merely increasing the dimensions of the original fracture. A change in fracture direction may be possible due to modification of the stress regime by previous stages.

PRODUCTION AND RESERVOIR ENGINEERING

Reservoir engineering encompasses the study of production characteristics of single wells and of the entire geothermal field and the design of production and fluid reinjection schemes so as to optimize energy produced against cost of production. As the scale broadens from that of a single production zone or feed zone in a well to that of the whole geothermal field, the reservoir engineer is asked different questions. For a single well, the engineer considers questions of optimum and maximum production rate, pressure and temperature decline under production, variation in the enthalpy of produced fluid, interference with nearby wells and predicted ultimate well lifetime and production scenario. This information, once integrated for all available wells, is used to predict ultimate development potential and longevity for the reservoir.

Before we consider how reservoir engineering measurements and predictions are made, let us turn to a discussion of the way the typical reservoir behaves under production.

Production from a Geothermal Reservoir

Donaldson and Grant (1981) give an excellent discussion of a single model first proposed by M. L. Nabb (1975) that explains many observed aspects of a geothermal reservoir under production. Figure 10 is a schematic representation of the model. The undisturbed reservoir is pictured as a vertical cylinder of thermal water surrounded by cold, non-thermal water. A lower hot water-saturated zone is overlain by a two-phase, steam and water zone, which is in turn overlain by a zone where cold groundwater _____ with the geothermal fluids beneath. In nature, the two-phase zone may or may not be present. Although boundaries between various elements of the model are shown with simple geometries, in nature they are much more complex, boundary irregularities being caused by the extreme variability in the permeability of rock.

The response of the reservoir to production is manifested by movement of these boundaries and by boiling in the two-phase zone. Let us see how such a simple model applies to some of the resource types previously discussed.

The Single-Phase Hot-Water Reservoir

We assume that in the pre-exploitation state the reservoir has reached a steady state with all rocks within the reservoir at the temperature of the fluid and that there is a steady inflow of thermal waters at the base, M_b, and a steady outflow at the surface, M_f and that $M_s = 0$. Under these conditions the reservoir will have much the same temperature from top to bottom, neglecting the very minor pressure effects because of the relative incompressibility of water.

If the reservoir is now topped through one or more wells at rate, Mw, all of the steady state parameters will change. It is probable that the base inflow, Mb, will not be altered significantly because Mb is determined by hydraulic conditions along a rather long recharge path. It is driven by the buoyancy effects of the rising hot water, which may be of order 30 MPa for a 300°C reservoir, whereas typical pressure variations due to production amount to only one-tenth of this value.

A side inflow, Ms, will be created, however, by the internal pressure drop due to production of the reservoir, and cold water at the boundaries of the field will begin to move inward. As these cold waters contact warmer

rock, they will be heated, and so the moving fluid front progresses at a rate much faster than the cool thermal front due to mining of heat from the rocks at the sides of the reservoir, and the temperature within the reservoir will remain constant at its steady state value.

The total well discharge, Mw, is controlled both by man and by nature. Although Mw depends on number of wells and valve settings, it depends to a greater extent on field pressure. Additional wells may increase production temporarily, but may cause pressure to fall more rapidly, thereby again constraining production. The field may set its own limits to Ms.

The surface discharge, M_f, will usually decrease, and M_f may even sign if surface water level drops where cold groundwater overlies the reservoir, thermal instability coupled with this dropping water level may lead to fingering of cold water downward into the reservoir.

Because fluid flow in the geothermal environment exhibits a greater or lesser degree of control by fractures these features become important to our model. Water flowing rapidly along a fracture has insufficient time to heat completely, if a fracture tapped by a well is connected to cold water at the sides or above a reservoir, this cold water can be drawn into the heart of the reservoir. This effect becomes more pronounced at higher production rates, alerting us to the potential for ruining a good well or even a field by production at too great a rate.

The Two-Phase Reservoir

Two-phase reservoirs behave very differently from single-phase reservoirs under production. These differences derive mainly from the effects of partitioning of fluid between the water and steam phases.

Because water is a relatively incompressible liquid, pressure pulses diffuse rapidly through a hot-water reservoir. A pressure pulse due to opening or closing a given production well may cross such a field in a few days. By contrast in a reservoir filled with a highly compressible fluid such as steam or a two-phase fluid such as a water-steam mixture it takes far longer, of the order of perhaps years, for a pressure pulse to cross the reservoir. If pressure is decreased at a specific point in a two-phase reservoir, as production from a well would case, the temperature also drops and boiling of the water phase is _____ with the steam produced attempting to increase pressure to restore original conditions. Energy is supplied to the boiling process from the rock. Thus, the pressure drop does not propogate for into the rock from the well because boiling makes up for the pressure loss, and no effect is seen at more distant points from the well. Figure 11 illustrates the variation in pressure and the rate of water loss from the rock matrix by boiling at a specific instant of time away from a well in a two-phase reservoir from which there is a constant mass withdrawal. Note that although the pressure increases rapidly away from the well, fluid is yielded to boiling with _____ variation over distances from the well. At a later instant of time the curves would move to the right, and the low pressure pulse would eventually propogate to the edges of the reservoir. Meanwhile, however, no effect of the withdrawal of fluid is felt at the boundaries, and so Ms remains zero. Both the fluid and the heat are being mined from the rock matrix itself.

There are two varieties of two-phase systems to be considered, the liquid-dominated and the vapor-dominated systems, according to which phase controls pressure within the reservoir. The basic model (Figure 10) may be viewed as illustrating liquid-dominated reservoir in which all wells top a

Because water is a relatively incompressible fluid, pressure pulses [illegible] rapidly through a [illegible]-water reservoir. [illegible]

[illegible]

two-phase zone which overlies a single-phase, hot water zone below. In such a field, it would take years for the low-pressure pulse due to onset of production to reach the reservoir boundaries. Until that time, both mass and heat would be extracted from the interior of the reservoir itself. The sideways propogation of the pressure front would eventually reach the cold boundary and inflow of cold fluid from the sides would be initiated. From this time onward there would also be a contribution of fluid from cold water inflow and of energy due to heating of the inflowing cold water, as was the case for a single-phase system. Downward propogation of the pressure move would eventually reach the base of the steam-water zone and would thence propogate quietly through the single-phase, water position of the reservoir. Boiling out the water both would be induced, with energy being supplied by the rock in the immediate vicinity and the interface level would progressively drop.

In a <u>vapor-dominated system</u> the water phase may be nearly or completely _____. Decrease in pressure through production will cause water to boil to contribute more steam, and both mass and heat are _____ from the rock in the vicinity of the well. Now variations in water saturation in the two-phase zone are directly related to variations in temperature and pressure. If production causes pressure (and thereby temperature) to decay enough, the water saturation may reach zero and a zone of dry steam may develop around the well. Within this dry steam zone there is then no mining of mass, and only a little mining of heat, which goes into superheatng the steam _____.

CONCLUSIONS

The porosity of crystalline rocks is small, typically 1 to 2%. More than 90% of the total porosity resides in unconnected pores, with the remainder in fractures and microcracks. Accordingly, more than 90% of the rock mass is not generally accessible to fluids either through advective or diffusional transport. Porosity can be measured 1) in the laboratory, 2) in situ by well logging techniques, or 3) calculated from well testing. Indirect methods do not uniquely distinguish between flow porosity, diffusion porosity and unconnected residual porosity. Flow porosity is critical to rock mass permeability, whereas residual porosity is the dominant form of porosity in crystalline rocks.

Permeabilities in crystalline rocks range over 12 orders of magnitude. Permeabilities in pristine, unfractured crystalline rock are commonly on the order of 10^{-6}d (1μd) or less. However, in situ measurements at individual sites may vary by as much as 4 to 6 orders of magnitude, and zones of >100 md are commonly encountered. These higher permeabilities are due to increased fracture density. Fracture permeability may be inferred provided information is available on the spacing, continuity, aperture and orientation of fractures. Thus, it is important to detect and characterize accurately fractures in targets for solution mining.

Permeability may decrease with depth, but its behavior is not systematic. Increased temperature has been shown experimentally both to increase and decrease permeability; these conflicting results arise from the complex interplay of thermal expansion, thermal stress cracking, and dissolution and redeposition of mineral phases.

RECOMMENDATIONS FOR RESEARCH

Research needs to be performed in several areas:

1) Establish the effects of pressure, temperature and chemical reactions upon permeability.
2) Characterize the fluid flow and diffusional pathways in crystalline rock such as tortuosity in microcracks and the degree of chemical equilibria between flow regimes and the host rock. Oxygen isotopes would be particularly useful in the latter and could be applied both to natural systems and laboratory experimental studies.
3) Develop improved indirect methods to determine permeability. It would be especially valuable to be able to utilize well logging methods to characterize fractures in terms of frequency, orientation and aperture.

BACKGROUND

Porosity and permeability are of fundamental importance to solution mining because they govern the fluid flow characteristics of the rock mass to a significant extent. Porosity is related to the fluid storage capacity of the system, and steady state fluid flow is simply a function of permeability and the pressure gradient. Accordingly, it is important to establish what the porosity and permeability are in potential targets for solution mining. Because these targets are often of difficult access, it is also important to know if porosity and permeability can be predicted, or if not, what kind of measurements are appropriate. Must porosity and permeability be measured in situ? Can they be measured or inferred remotely or by instrumental technique? Are laboratory measurements suitable?

EVALUATION OF STATE OF THE ART

Porosity

Formulations and models. Porosity is simply the ratio of pore volume to bulk volume. For crystalline rocks a convenient model is presented by Norton and Knapp (1977) whereby the total porosity, ϕ_T, is given by $\phi_T = \phi_F + \phi_D + \phi_R$ where ϕ_F is the flow porosity in which mass transport is dominated by fluid flow, ϕ_D is the diffusional porosity in which transport is by diffusion in the aqueous phase, and ϕ_R is the residual porosity composed of pores unconnected to ϕ_F or ϕ_D (Fig. 1.5-1). ϕ_F contributes significantly to the permeability of the rock along planar features such as joints, fractures or bedding planes, and ϕ_F = nd for any given parallel fracture array, where n is the fracture abundance (L^{-1}), and d is the fracture aperture. For the Mayflower pluton in the Park City district, Utah, $\phi_F = 2 \times 10^{-3}$ to 3×10^{-5} compared to ϕ_T of 0.01 to 0.03 (Villas, 1975; Villas and Norton, 1977). In the Sherman granite $\phi_F = 5 \times 10^{-6}$ and $\phi_T = 10^{-2}$ (Pratt et al., 1974). In general, ϕ_F constitutes a small proportion of the bulk total porosity; nonetheless, flow porosity is primarily responsible for the permeability of the rock mass.

Experimental evidence indicates that diffusion porosity in fractured crystalline rocks is on the order of 10^{-3} to 10^{-4} in rocks with total porosities of 10^{-2} to 10^{-1}. Because ϕ_F and ϕ_D are small, the bulk of the total porosity of crystalline rocks is in the residual porosity; $\phi_R > 0.9\phi_T$ (Norton and Knapp, 1977). This is significant for it means that more than 90% of crystalline bulk rock porosity is not accessible to fluids either through advective or diffusional transport.

Porosity Data. Davis (1969) has summarized porosity data for a wide range of rock types. The hydrologic characteristics of sedimentary rocks are so well known from research by the petroleum industry that they are not dealt with here; comprehensive studies are summarized by Davis (1969). Porosities of crystalline rocks are less well studied. Norton and Knapp (1977) have determined porosities on over 75 samples of fresh, altered and mineralized rocks. Unaltered igneous rocks should have finite porosity due to cooling effects. For particular geologic settings, total porosities exhibit moderate ranges: Chino, New Mexico skarns, 0.73×10^{-2} - 9.43×10^{-2}; San Manuel, Arizona altered quartz monzonite, 1.46×10^{-2} - 6.84×10^{-2}; Butte, Montana quartz monzonite, fresh = 1.01×10^{-2}, sericitized = 6.35×10^{-2}, argillized = 0.0753×10^{-2}; Bingham, Utah 2.64×10^{-2} - 12.5×10^{-2}; Ronda, Spain metamorphics, 0.66×10^{-2} - 8.42×10^{-2}. Norton and Knapp (1977) observed that the majority of the hydrothermally altered plutonic rocks and calc-silicates they studied had higher total porosities than their unaltered equivalents; the major component of the increased porosity is in larger residual porosity in the altered rocks. Fabric analyses indicate a correlation between mineral grain size and pore distribution in which pores occur predominately between grains and the more continuous pores are concentrated around larger mineral grains (Norton and Knapp, 1977).

Permeability.

Formulation and units. Permeability is a second rank tensor associated with a flux and a gradient:

$$q_i = -K_{ij} \frac{\partial P}{\partial X_j} \qquad \text{(Eq 1.5-1)}$$

where when $\partial P/\partial X_j$ is the pressure gradient, q_i is the volume rate of flow per unit time and K_{ij} are constants. This is an empirical relationship known as Darcy's Law. It has been shown experimentally that K_{ij} is a function of the characteristics of both the fluid and the medium. In order to separate the effects of the two and concentrate on characteristics of the medium, Darcy's

Law is often modified as follows (shown in scalar form):

$$q = \frac{k}{\mu} \left(\frac{\partial P}{\partial X}\right) \qquad \text{(Eq 1.5-2)}$$

where μ is the viscosity of the fluid, and k is the permeability with dimensions of L^2. For the purposes of this section we shall use the unit of the darcy (10^{-8} $cm^2 \approx 10^{-11}$ ft^2). Units of hydraulic conductivity also known as coefficient of permeability, common in ground-water studies, superimpose fluid properties on permeability and have dimensions of LT^{-1}, and 1 darcy $\approx$ 10^{-3} cm/sec = 10^{-5}m/s. In the U. S. the meinzer is a unit of hydraulic conductivity. For pure water at 15.6°C (60°F), 1 darcy = 18.2 meinzer = 18.2 gallon/day/ft^2.

Observations. There are three fundamental types of permeability determinations 1) laboratory measurements, 2) in situ measurements, and 3) inferred permeabilities from geological phenomena. Brace (1980) has summarized the various measurement techniques and compiled available permeability data on crystalline and argillaceous rocks. These data are illustrated in Fig. 1.5-2. The range in permeabilities spans more than 12 orders of magnitude, and wide ranges occur within similar rock types.

Laboratory measurements of granites and metamorphic rocks range from 10^{-4} to 10^{-12}d. In situ measurements vary from greater than 1 d to 10^{-10}d for granites and gneisses; in general k falls between 1μd and 100 md, and in situ values are often somewhat greater than laboratory values. Of particular significance is that in situ values in crystalline rock vary by over 4 orders of magnitude at a given site, and at most sites some volume of the rock has a permeability of >100 md.

Laboratory permeability determinations of well characterized crystalline rocks include the following: Westerly granite, 4 x 10^{-8}d (Heard et al., 1979), 3-5 x 10^{-9}d (Brace et al., 1968); Sherman granite, 40-100 x 10^{-6}d (Pratt et al., 1977); Climax granite, < 10^{-9}d (Ramspott, 1979); Barre granite, 10^{-6} - 10^{-7}d (Kranz et al., 1979); White Lake gneiss, 10^{-10} - 10^{-12}d (Heard et al., 1979). In situ measurements at the Stripa Mine, Sweden yield k = 10^{-5} - 10^{-6}d for low fracture-density rock, and k = 10^{-2}d for well fractured rock (Witherspoon et al., 1979a,b). Variable fracture permeability undoubtedly contributes to the wide range in values of k for other in situ measurements.

Estimates of permeabilities in the deeper crust can be made from inferred permeabilities (Fig. 1.5-1C.). Permeabilities may be inferred either from fracture spacing and aperture analysis (discussed below) or from geological phenomena involving time and distance controlled predominantly by hydraulic diffusion (induced seismicity, for example). As expected, jointed and fractured rock have high permeabilities, > 10^{2} - 10^{-2}d. Upper crustal seismic phenomena suggests k values of 10^{-1} - 10^{-4}d. Volcanic terranes will be heterogeneous in permeability; lithologies can vary from dense crystalline rock, highly fractured rock, porous but relatively impermeable pyroclastic material, and interbedded clastic sediments. The gross permeability will be anisotropic with greatest permeabilities parallel to volcanic flow directions. Oceanic crust, which may have significant through-going fractures, has inferred k of 10^{-1} - 10^{-3}d (Fenn and Cathles, 1979) to 4.5 x 10^{-4}d (Ribando et al., 1976). Inferred permeabilities for the Skaergaard intrusion (Norton and Taylor, 1979), which sample deeply into the crust, are 10^{-4} - 10^{-5}d in gabbro at 4-8 km, and 10^{-8}d in host gneiss at 7-10 km. Very low permeabilities (10^{-10}d) are necessary in order to attain pore pressures equivalent to lithostatic pressure at depth and to provide effective sealing in argillaceous rocks above deep oil and gas reservoirs.

Fracture Permeability. The common observation that laboratory measurements yield lower permeabilities than in situ measurements is generally attributed to fractures, faults, and joints in the rock mass not present in small laboratory samples. For example, the in situ permeability of the jointed Sherman granite was measured as 1.2×10^{-3}d compared to 10^{-4}d for an intact sample in the laboratory (Pratt et al., 1977).

In crystalline rocks in nature, fracture permeability is the dominant control on fluid flow. In the Skaergaard intrusion, for example, isotopic evidence demonstrates that fluid flow occurred primarily along fractures. Interstitial fluids maintained isotopic equilibrium by diffusion through microcracks to the fracture-flow system. Plagioclase $\delta^{18}O$ values are lowest near major fractures, indicating higher water/rock ratios (Taylor and Forester, 1979); $\delta^{18}O$ values increase away from major fracture zones.

Determination of the three-dimensional distribution of permeability in the target volume is important. The basic parameters of the natural fracture system that must be known are spacing, continuity, aperture and orientation. Mapping studies give good information on continuity and fracture system geometry, but do not define aperture well. In situ hydrologic experiments yield aperture information (Witherspoon et al., 1979a,b). When these are combined with fracture geometry directional permeabilities can be estimated. Fracture continuity is best determined by pressure measurements. Witherspoon et al., (1979a) have demonstrated that experimental laboratory measurements of fracture permeability on small diameter cores may be significantly less than in field tests. At the very least, they recommend that large sample sizes on the order of 1 m diameter be employed.

Data on the permeability of naturally fractured rock is sparse; in most cases estimates of permeability have been based on field observations combined with a simple planar fracture model. The model assumes that the fractures constitute a set of equidistant, parallel planes of infinite extent (Snow, 1968, 1970). For a single parallel plate opening, the volume rate of flow is a function of the aperture of the fracture, d, and the viscosity of the fluid, μ (Snow 1968, 1970; Norton and Knapp, 1977):

$$q = \frac{d^3}{12\mu} \left(\frac{\partial P}{\partial x}\right) \qquad \text{(Eq. 1.5-3)}$$

From (Eq. 1.5-2) and (Eq. 1.5-3) it follows that

$$k = \frac{nd^3}{12} \qquad \text{(Eq. 1.5-4)}$$

where n is the fracture abundance (in fractures-cm^{-1}). In ore bodies, fractures may be closely spaced. At Bingham, Utah, n = 0.5 fractures/cm, and in the Mayflower pluton, Park City, Utah, n = 6-21 fractures/m, with other stocks in the area ranging from 0.6 - 3.0 fractures/m (Villas, 1975; Villas and Norton, 1977; Norton and Knapp, 1977). Unaltered igneous bodies have $n \simeq 0.1$ fracture/m. Fracture apertures in the Mayflower pluton range from 40 to 270 μ m and average 145 μ m (Villas, 1975). These values yield wide ranges in permeabilities in the Mayflower Pluton from < 10^{-6} to 7d (Villas and Norton, 1977).

In addition to fracture geometry, fracture asperity or surface roughness contributes to permeability variation. A perfectly planar fracture may close completely upon application of normal stress, whereas surface roughness will affect permeability in two ways. Rough surfaces will inhibit closure of the fracture and will also change the path length or tortuosity of the flow path. Pratt et al. (1977) demonstrated that when a joint "closed" upon application of normal stress in situ, permeability remained 3 orders of rger than for the unjointed rock. In the laboratory, Kranz et al. rved a rapid decrease in permeability (from an initial 8×10^{-5}d) e granite as fractures closed, the rate of which was a function of ughness. However, they achieved low permeabilities equivalent to ctured rock (10^{-6} to 10^{-7}d) at confining pressures of 2 kb.

ssure effects. The data compilation of Brace (1980, Fig. 1.5-3) a subtle decrease in permeability with depth; however it is not tic, and formulations in which permeability decreases exponentially epth (c.f. Fenn and Cathles, 1979, p. 249) are not generally cable. Experiments demonstrate a decrease in permeability with asing pressure. The permeability of Westerly granite declined from 350 x d at 100 b pressure to 4×10^{-9}d at 4000 b. The permeability change was sely correlated with electrical resistivity, and led to an extrapolation of $\approx 0.5 \times 10^{-9}$d at 10 kb (Brace et al., 1968). Heard et al. (1979) produced a ctor of two reduction in permeability in Westerly granite with applied essure. However the permeability of the White Lake gneissic granite peared to be unaffected by either pressure or differential stress, even ough differential stress produced a factor of 2 variation in permeability in sterly granite. Similarly, Witherspoon et al. (1979a) produced a decrease hydraulic conductivity through a fracture with increased applied normal ess. In the Eleana argillite at the Nevada Test Site, permeability varies h a negative power of the effective pressure in experiments up to 24 MPa proximately equivalent to 2 km depth), and permeability ranged from 10^{-11} 10^{-8}d, depending upon the effective pressure. A single through-going cture increased k by 3 to 4 orders of magnitude at low pressure and 1 to 2 ers of magnitude at 24 MPa (Lin, 1978).

Temperature effects. There is little data on the effects of temperature permeability, nor is there consistent agreement as to its effects. In situ asurements on the Climax granite at the Nevada Test Site, at a depth of 425 recorded a systematic decrease in permeability as the rock mass was heated d an increase in k as temperature declined. Permeabilities were less than 1 anodarcy, with a minimum of 0.02 nd at the highest temperature of pproximately 325°C (Ramspott, 1979). The decrease in permeability is resumably due to closing of cracks accompanying thermal expansion of the rock nass. Heard (1980) heated Climax quartz monzonite to 300°C. In this instance the thermal expansion was accompanied by new microcrack formation which was inferred to enhance the permeability by a factor of 2 to 5. Laboratory measurements of Westerly granite (on cylinders 1.59 cm in diameter, 3.81 cm in length) demonstrate that the initial permeability at elevated temperatures is higher by 1 to 2 orders of magnitude than at room temperature. In this case, enhanced permeability is attributed to thermal stress cracking. However, permeabilities ultimately declined significantly due to dissolution and reprecipitation of plagioclase and quartz (Summers et al., 1978). Similar results were achieved by Morrow et al. (1981) in Westerly granite where permeability decreased by 1 to 2 orders of magnitude at elevated temperature, again as a result of dissolution and redeposition of quartz and feldspar along fractures.

Hydrology

Hydrology is the study of the movement of fluids in the earth. In this it is related to reservoir engineering, but these related disciplines have evolved along somewhat separate paths that are just now coming together. Hydrology has traditionally been concerned mainly with determination of the flow of rather shallow water (surface to a few thousand feet) ~~and how that flow can be determined from~~ as regards production of water from wells and springs for ~~[illegible]~~ domestic, agricultural and industrial uses. By contrast reservoir engineering began in the study of the movement of fluids, both brines and hydrocarbons, in petroleum reservoirs. Today we have two sets of terminology, units and characteristic ways of writing equations and formulating problems that describe much the same physical phenomena. Nevertheless, a person ~~[illegible]~~ educated and experienced along traditional hydrological lines is an important adjunct to an exploration program.

heat flow -
dispersion of zoning patterns
suppression of manifestations, such as springs
source areas for recharge

The insight

Knowledge ~~Study~~ of the ~~near surface~~ flow of normal groundwater in a geothermal prospect area is important for a number of applications. Perhaps most important is the profound adverse affect that ~~flow to~~ lateral or vertical flow of groundwater can have on other exploration methods, particularly the thermal and geochemical methods

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GEOTHERMAL ENERGY

AN OVERVIEW OF OCCURRENCE AND EXPLORATION

by

Phillip M. Wright
Director
Earth Science Laboratory
University of Utah Research Institute
Salt Lake City, Utah, USA

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INTRODUCTION

Development of geothermal resources if being aggressively pursued on a worldwide basis. Approximately 3 800 MW of electricity are currently being generated from geothermal energy, and about 10 000 thermal MW are being used for direct heat applications. While this may seem small compared to the estimated 8.4 x 10^6 MW of human use of fossil energy (Williams and Von Herzen, 1974), it nevertheless represents a savings in the consumption of about 77 million barrels of oil per year worldwide. It is very difficult to estimate the ultimate potential contribution of geothermal energy to mankind's needs for at least three reason: 1) long-range future energy costs, although generally predicted to be higher than today's levels, are uncertain, and a large number of lower-grade geothermal resources would become economic at higher energy prices; 2) only preliminary estimates of the worldwide resource base have been made, and; 3) technology for using energy in magma, hot rock and normal thermal-gradient resources, whose potential contributions are very large, is not yet available.

Geothermal energy is heat that originates within the earth. The earth is an active thermal engine, and 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 phenomena as motion of the earth's crustal plates, uplifting of mountain ranges, occurrence of earthquakes, eruption of volcanos and spouting of geysers all owe their origin to the transport of internal thermal energy.

Although the feasibility of use of geothermal energy has been known for many years, the total amount of application today is small compared with the potential for application. The present availability of less expensive energy from fossil fuels has suppressed use of geothermal energy at all but a few of the highest-grade resources. Research and development of new techniques and equipment is needed to decrease costs of exploration, drilling, reservoir evaluation and extraction to make the vastly more numerous lower-grade resources also economic.

The objective of this paper is to present an overview of the nature of and exploration for geothermal resources.

NATURE AND OCCURRENCE OF GEOTHERMAL RESOURCES

Although the distribution with depth in the earth of density, pressure and other related physical parameters is 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, 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. Using present technology and under favorable conditions, holes can be drilled to depths of about 10 km, where temperatures range upward from about 150°C in areas underlain by cooler rocks to perhaps 600°C in exceptional areas.

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 is 61 x 10^{-3} watts/m^2 for the continents and 92 x 10^{-3} watts/m^2 for the oceans, including effects of sea-floor spreading discussed below, 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 42 X 10^{12} watts (42 million megawatts), a very large amount indeed (Williams and Von Herzen, 1974). 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 82 milliwatts/m^2 is about 50,000 times smaller than the flux of heat from the sun (much of which is reflected or re-directed into space), 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 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. The relative contribution to the observed surface heat flow of these two mechanisms is not yet resolved. Some theoretical

TEMPERATURE VS DEPTH IN EARTH

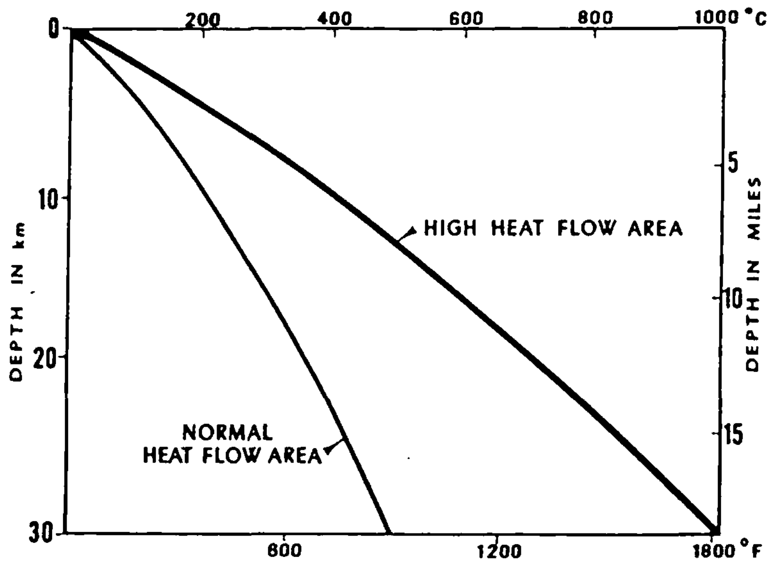


FIGURE 1

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 4.7 billion years ago. We do know that the thermal conductivity of crustal rocks is low so that heat escapes from the surface very 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. White (1965) has estimated that the total heat stored above surface temperature in the earth to a depth of 10 km is about 1.3×10^{27} J, equivalent to the burning of about 2.3×10^{17} barrels of oil. It is apparent that if even a small part of this heat could be made available, its contribution would be significant.

Geological Processes

The fundamental cause of geothermal resources lies in the transport of heat near to the surface through one or more of a number of geological processes. We have seen that the ultimate source of that heat is in the interior of the earth where temperatures are much higher than they are at the surface. Geothermal resources commonly have three components:

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

We will now discuss some of the geological aspects of each of these factors.

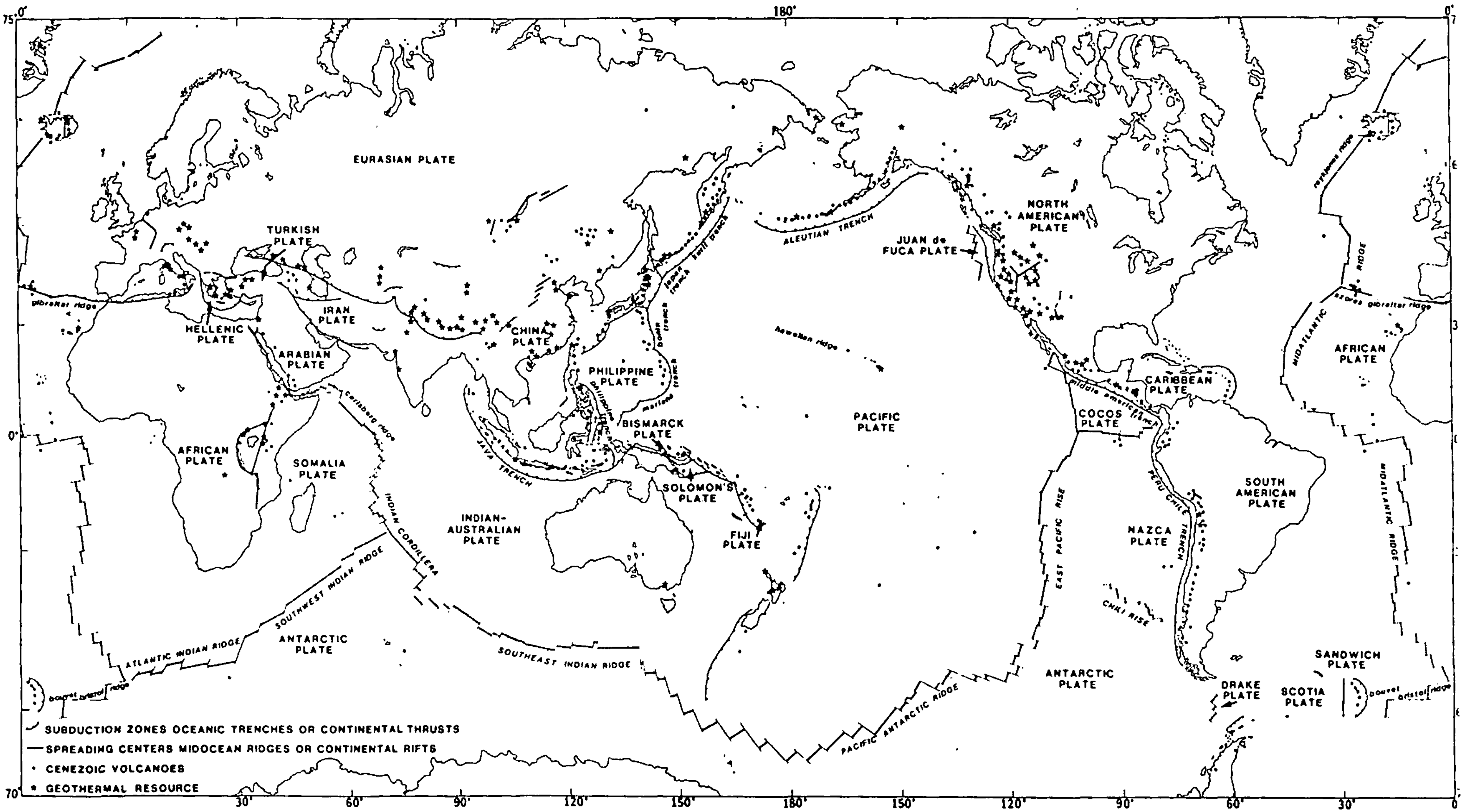
Heat Source. Geothermal resource areas, or geothermal areas for short, are generally those in which higher temperatures are found at shallower depths than is normal. This condition usually results from either 1) intrusion of molten rock to high levels in the earth's crust, 2) higher-than-average flow of heat to the surface with an attendant high rate of increase of temperature with depth (geothermal gradient) as illustrated in Figure 1, often in broad areas where the earth's crust is thin, 3) heating of ground water that circulates to depths of 2 to 5 km with subsequent ascent of the thermal water to the surface, or 4) anomalous heating of a shallow rock body by decay of an unusually high content of radioactive elements. In many geothermal areas, heat is brought right to the surface by circulation of ground water. If temperature is high enough, steam may be produced, and geysers, fumaroles, and hot springs are common surface manifestations of underlying geothermal reservoirs.

The distribution of geothermal areas on the earth's surface is not random but instead is governed by geological processes of global, regional and local scale. 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, our learning rate is high.

Figure 2 shows the principal areas of known geothermal occurrences on a world map. Also indicated are areas of young volcanos 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 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 3. 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 next layer, the mantle, which lies below the crust. Mantle material below the lithosphere is less rigid 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 mixture of the same metals.

One unifying geological process that generates heat sources is known 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 convec-



GEOTHERMAL RESOURCES AND PLATE TECTONIC FEATURES

Figure 2

INTERIOR OF THE EARTH

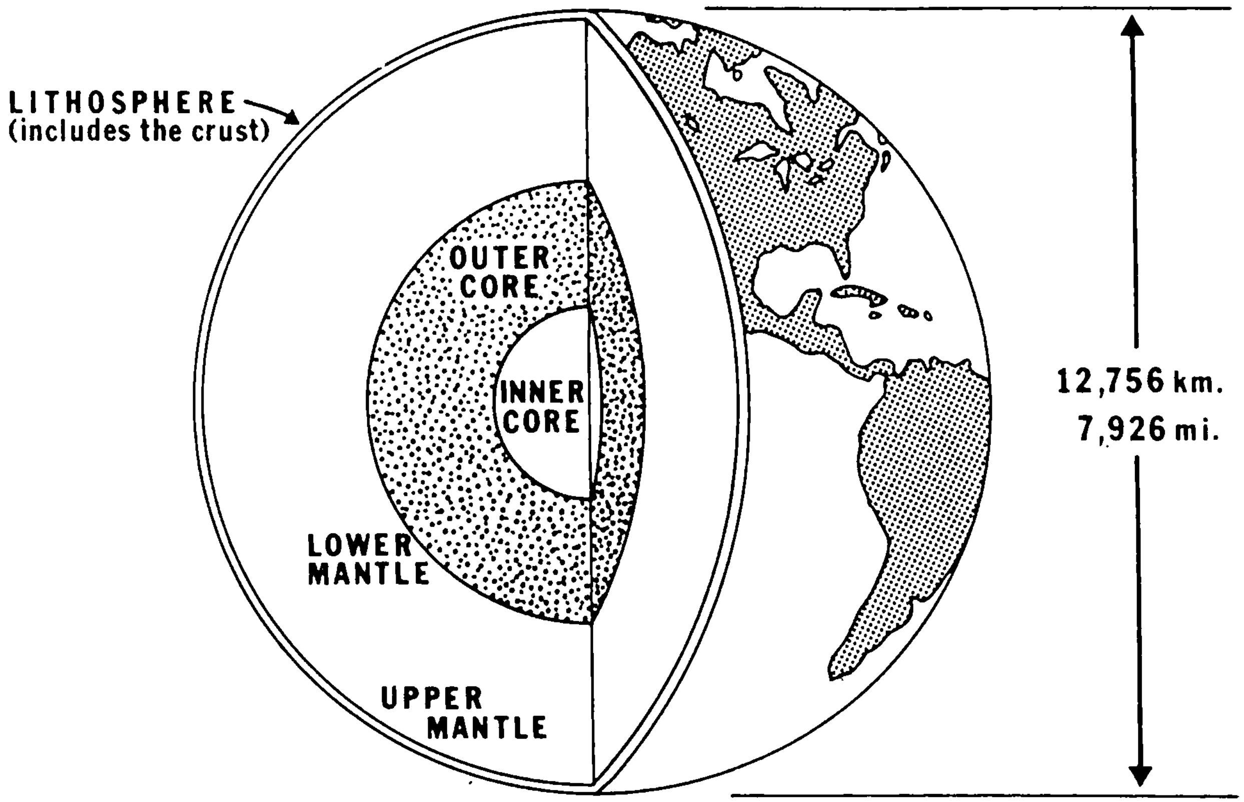


Figure 3

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CONCEPT OF PLATE TECTONICS

(NOT TO SCALE)

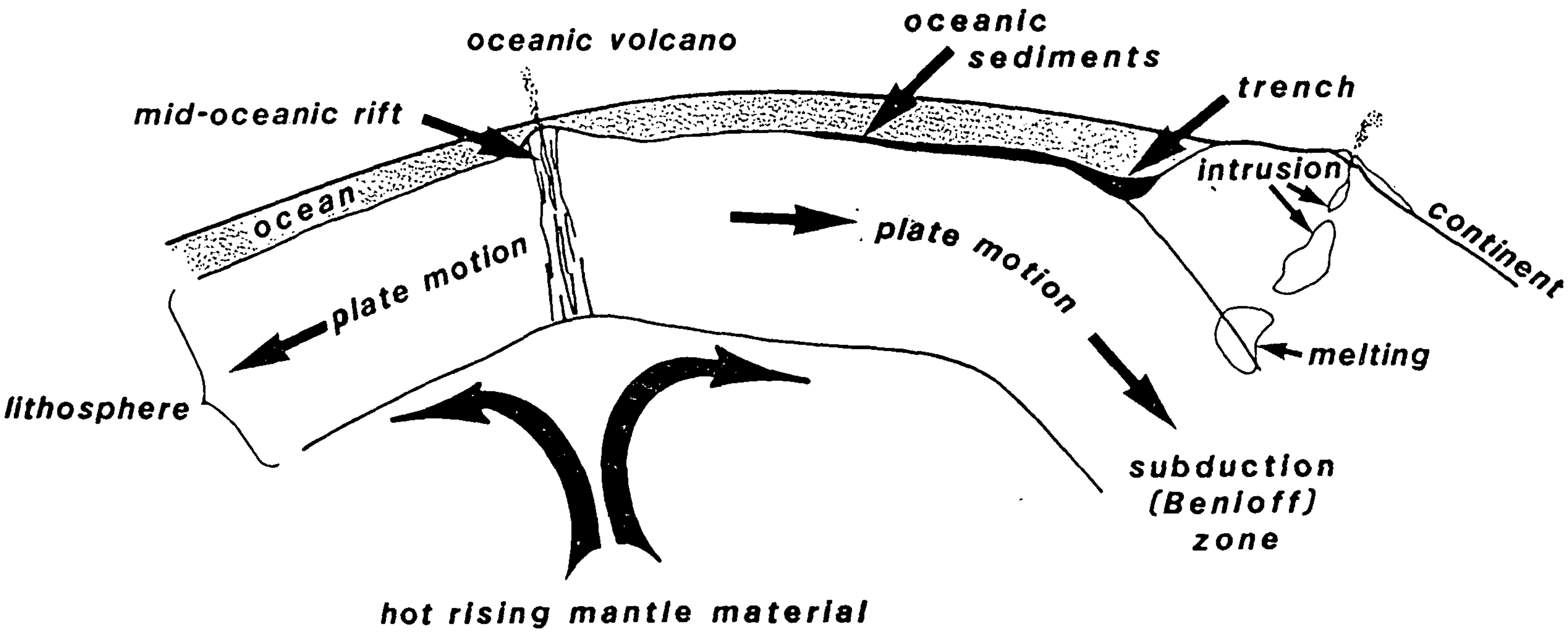

Figure 4

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tion 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 is dragged 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. 2 and 4), typified by the mid-Atlantic Ridge and the East Pacific Rise. 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. This intrusion of molten material brings large quantities of heat to shallow depths and is the heat source for the recently discovered oceanic hydrothermal systems. The laterally moving oceanic lithospheric plates impinge against adjacent plates, some of which 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 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. The upper layers of the descending plate often contain oceanic sediments rich in water, thus assisting in the melting process. The molten or partially molten rock bodies (magmas) ascend buoyantly through the crust (Figs. 4 and 5), probably along lines of structural weakness, and carry their contained heat to within 1.5 to 20 km of the surface. They may give rise to volcanos if part of the molten material escapes to the surface through faults and fractures in the upper crust. These shallow crustal intrusions occur on the landward side of oceanic trenches, usually 50 km to 200 km inland (Fig. 4). They are the cause of the volcanos in the Cascade Range of California, Oregon and Washington, for example, and of those of Central and South America. A number of these volcanic areas have geothermal systems associated with them.

Figure 2 shows where these processes of crustal spreading, formation of

CRUSTAL INTRUSION

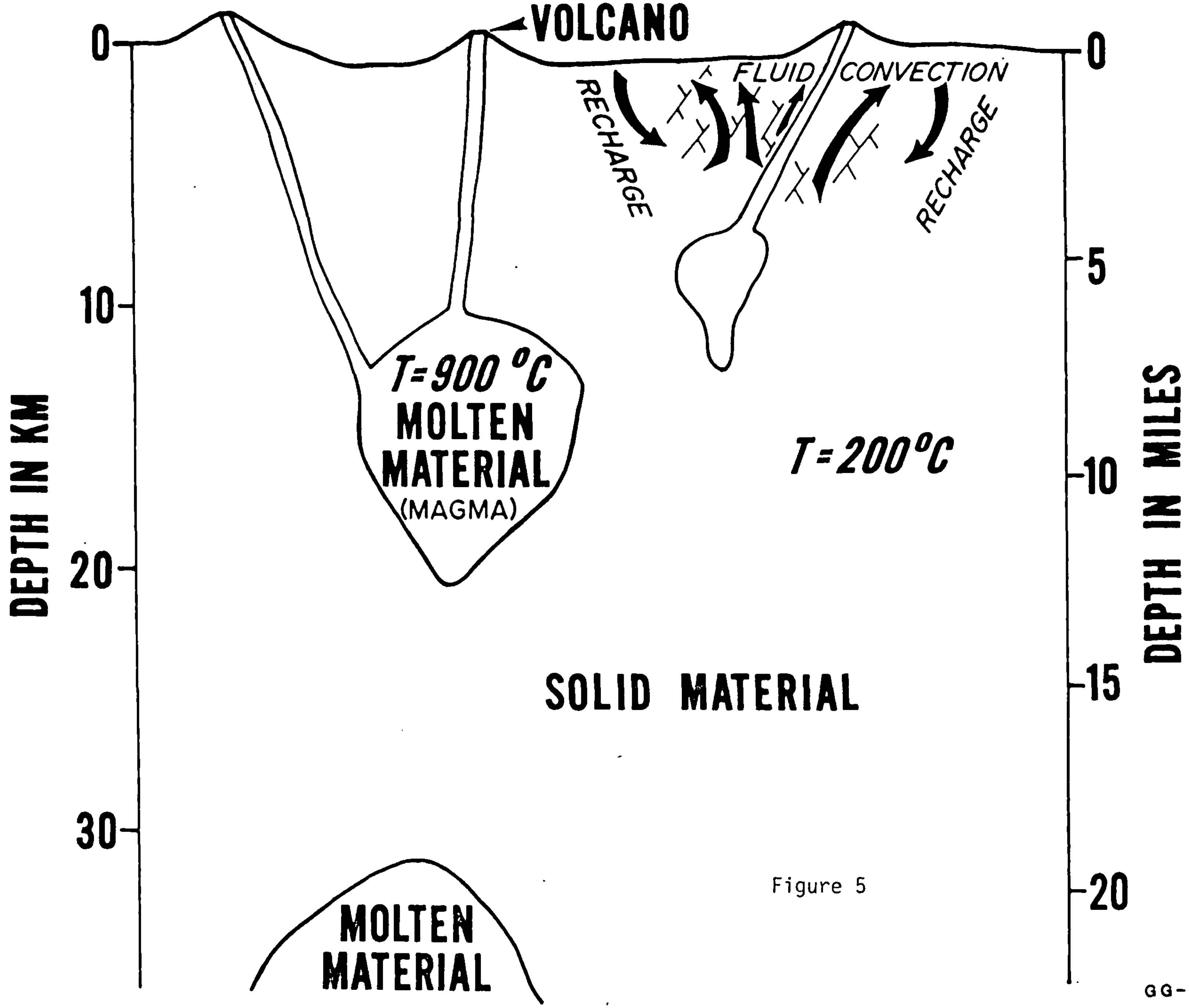

Figure 5

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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 consumed in the subduction zones, usually marked by trenches. 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 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 in somewhat more detail the development of crustal intrusions, 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. Intrusion of 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 (hundreds of square miles) 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 compositional 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 can be seen in the rocks that compose the Hawaiian Islands. The crystalline, plutonic variety of this rock that has cooled slowly and 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 also used for any crystalline igneous rock. Magmas that result in basalt or gabbro are termed "mafic" or "basic" whereas magmas that result in rhyolite or granite are termed "felsic" or "acidic".

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 seem to indicate a more or less direct pipeline for the magma from the upper mantle to the surface.

The origin of granites is a subject of some controversy. It can be shown that felsic magma may be derived by progressive segregation of the melt fraction of a basaltic magma as it cools and begins to crystallize. However, the chemical composition of granites is much like the average composition of the continental crust, and some granites probably also result from melting of crustal rocks by upwelling basaltic magmas. Basaltic magmas melt at a higher temperature and are less viscous (more fluid) than granitic magmas. Occurrence on the surface or in drill samples of felsic 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 they may indicate a large body of viscous magma at depth to provide a geothermal heat source. On the other hand, occurrence of young basaltic rocks 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 (Smith and Shaw, 1975). 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.

Another important source of volcanic rocks are the hypothesized point sources of heat in the mantle as contrasted with the rather large convection cells that drive plate motions. 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 volcanos is developed. Young volcanic rocks occur at one end of the chain with older ones at the other end. The Hawaiian Island chain is an example. The youngest volcanic rocks on the island of Kauai to the northwest end have been dated through radioactive means at about 4 million years, whereas the volcanos Mauna Loa and Mauna Kea on the island of Hawaii at the southeast end of the chain are forming today and are in almost continual eruptive activity. A new island southeast of Hawaii is being formed by suboceanic volcanic eruptions and at present is only a few hundred feet beneath the surface. To the northwest, the Hawaiian chain continues beyond Kauai for more than 2000 miles to Midway Island, where the last volcanic activity was about 16 million years ago. The trace of the island chain is consistent with the motions of the pacific plate postulated by geophysicists. Geologists also 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.

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 (40°C/km to 60°C/km). Geophysical and geological data indicate that the earth's crust is thinner than normal in the Basin and Range province and that the isotherms are warped upward beneath this area. Much of the western U. S. is geologically active, as manifested by earthquakes and active or recently active volcanos. Faulting and fracturing during earthquakes help to keep fluid pathways 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 the surface. Many of the hot springs and wells in the western United States and elsewhere owe their origin to such processes.

<u>Permeability</u>. 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 minerals themselves but rather it flows in pores between mineral grains and in open spaces created by fractures and faults. Porosity is the term given to the fraction or percent of void space in a volume of

rock.

Permeability and porosity can be primary or secondary, i.e. formed with the rock or subsequently. Primary permeability in sedimentary rocks originates from intergranular porosity and it usually decreases with depth due to compaction and cementation. In volcanic sequences, primary intergranular porosity and permeability exist, but primary permeability also exists in open spaces at contacts between individual flows and within the flows themselves. Secondary permeability occurs in open fault zones, fractures and fracture intersections, along dikes and in breccia zones produced by hydraulic fracturing (Brace, 1968; Moore et al., 1985). Changes in permeability may also come about through mineral deposition by leaching by the thermal fluids.

Regarding exploration for hydrothermal systems, the key problem appears to be more in locating permeable zones than in locating high temperatures. Fractures sufficient to make a well a good producer need be only a few millimeters in width, but must be connected to the general fracture network in the rock in order to carry large fluid volumes. Grindly and Browne (1976) note that of 11 hydrothermal fields investigated in New Zealand, all of which have high temperatures (230°C to 300°C), five are non-productive chiefly because of low permeability. Three of the eleven fields are in production (Wairakei, Kawerau and Broadlands) and in each of these, permeability limits production more than temperature does.

Permeabilities in rocks range over 12 orders of magnitude. Permeabilities in pristine, unfractured crystalline rock are commonly on the order of 10^{-6} darcy (1μd) or less. However, in situ measurements at individual sites may vary by as much as 4 to 6 orders of magnitude, and zones of >100 md are commonly encountered. These higher permeabilities are due to increased fracture density. Fracture permeability may be inferred provided information is available on the spacing, continuity, aperture and orientation of fractures. Permeability in igneous rocks may decrease with depth, but its behavior is not systematic. Increased temperature has been shown experimentally both to increase and decrease permeability; these conflicting results arise from the complex interplay of thermal expansion, thermal stress cracking, and dissolution and redeposition of mineral phases.

Most geothermal systems are structurally controlled, i.e. the magmatic heat source has been emplaced along zones of structural weakness in the crust. Permeability has usually been increased in the vicinity of the intrusion from fracturing and faulting in response to stresses involved in the intrusion process itself and in response to regional stresses. Thus, an understanding of the geologic structure of a resource area can lead not only to evidence for the most likely location of a subsurface magma chamber, but also to inferences about areas of higher permeability at depth. Such areas would be prime geothermal exploration targets.

Heat Transfer Fluid. The purpose of the heat transfer fluid is to remove heat from the rocks at depth and bring it to the surface. This fluid is either water (usually saline) or steam. Water has a high heat capacity (amount of heat absorbed or released when the temperature increases or decreases by 1°C, 80 cal/gm) and a high latent heat of vaporization (amount of heat needed to convert water to steam or released when steam condenses, 540 cal/gm). Thus water, which pervades fractures and other open spaces in rocks and so is available in nature, is an ideal heat transfer fluid because a given quantity of water or steam can carry a relatively large amount of heat to the surface where it is easily removed.

The density and viscosity of water both decrease as temperature increases. Thus, water heated at depth in the earth is lighter than is cold water in surrounding rocks, and is therefore subjected to buoyant forces. If the heating is great enough that the buoyant forces overcome the resistance to flow imposed by the rock, the heated water will rise toward the earth's surface. As it does so, cooler water will move in to replace it, be heated and also rise. Because heated water will flow along paths of least resistance, it may also move laterally in places, but its net flow will be upward. In this way, natural convection can be set up in the groundwater above a source of heat such as an intrusion. This convective process can bring large quantities of heat near enough to the earth's surface to be reached by wells, and is thus responsible for the most economically important class of geothermal resources, as we shall discuss below.

Water can also form steam in the earth, and the process can be quite complex. The transition of water to its gaseous form occurs at a temperature which is a function of pressure (Figure 6). This transition absorbs a considerable latent heat (540 cal/gm at 100°), and the latent heat appreciably diminishes when the temperature increases. The density of steam, like all

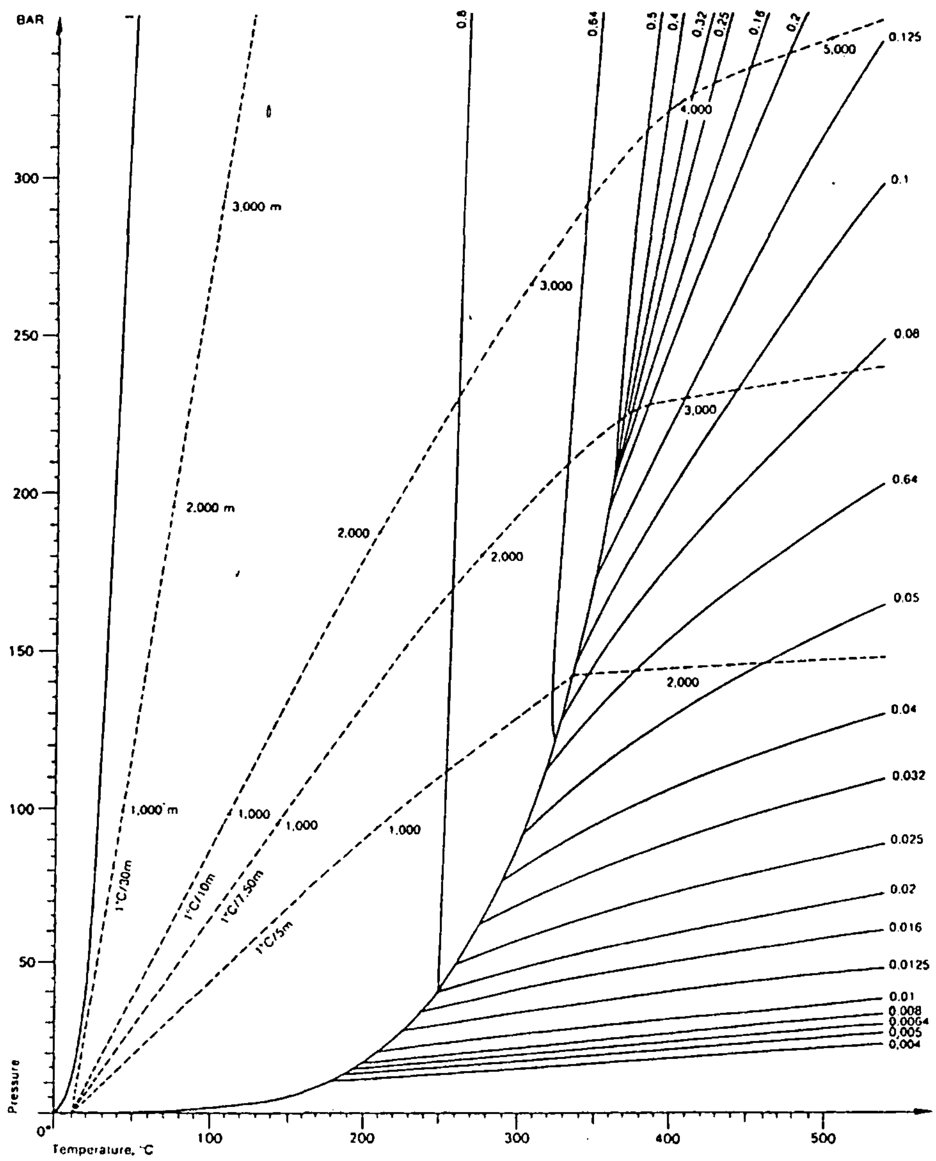

FIGURE 6. Graph of water density as a function of temperature and pressure. The dashed lines indicate the states as a function of the depth for a fluid phase in static equilibrium with a temperature of 11°C at the phreatic level and the different values of the thermal gradient, supposed uniform, taking account of the water density. Note that the critical state of water is reached for a gradient of about 1°C/7.5 m at a depth of 2,700 m. For a higher gradient, the hypothesis of a uniform thermal gradient becomes incompatible with the hypothesis of a static fluid equilibrium, this equilibrium being unstable.

gases, diminishes when the temperature increases and increases with pressure. At the critical point (T = 375°C, pressure = 221 bars), the specific volume of steam becomes equal to that of liquid water, the latent heat falls to zero, and there are no longer any differences between the two phases. For temperatures and pressures with higher values than the critical point, the supracritical domain, there exists only a single fluid phase whose density varies in a continuous manner, as shown in Figure 6. There are tables for water as well as for steam which furnish all the pertinent characteristics as functions of temperature and pressure. The viscosity values appear to be the least well known in the supracritical domain.

If the water contains dissolved gas, the gas will accumulate in the vapor phase, where the total pressure will be the sum of the partial pressures of the gases and the water vapor. On the other hand, dissolved salts are distributed very unequally between the two phases, nearly all being found in the liquid.

In some convective hydrothermal resources, the temperature never exceeds the boiling point for the pressure at any particular depth, and the system does not generate steam. However, in other systems the temperature can rise above the local boiling point, and steam is produced. The steam ascends and meets cooler rocks where it partially condenses while heating the rocks, and the pressure drop due to condensation brings up more steam. Cooler water descends alongside the steam cell and is heated by hotter rocks at depth until it finally vaporizes at the level of boiling. In this way, steam convection is set up. Several small steam plumes or convection cells may be present to start, but we predict the evolution toward a smaller number of important ascenting cells, each producing a substantial flow of vapor. If there is a sealed zone, a zone of low permeability, above the steam cells, steam will accumulate in the reservoir. The temperature and pressure in such a steam reservoir vary only slowly with depth. At Larderello, Italy, the reservoir temperature and pressure are 240°C and 35 bars, values that appear to be typical of other vapor-dominated systems.

With the foregoing material as background, we are now in a position to develop a classification for geothermal systems and to describe their workings in more detail.

Classification of Geothermal Resources

The classifications of geothermal resource types shown in Table 1 is modeled after one given by White and William (1975). Each resource type will be described briefly with emphasis on those that are presently nearest to commercial use in the U.S. In order to describe these resources, 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. In spite of disagreement over details, however, the models predicted below are generally acceptable and aid our thinking on the topic.

Geothermal resource temperatures range upward from the mean annual ambient temperature (usually 10-30°C) to well over 350°C. Figure 7 shows the span of temperatures of interest in geothermal work. It will be helpful for the reader to refer to this figure during the subsequent discussions.

Convective Hydrothermal Resources. Convective hydrothermal 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 (Fig. 5) that is 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 resource depends, among other less important variables, on temperature and pressure conditions at depth.

Figure 8 (after White et al., 1971) shows a simple conceptual model of a hydrothermal system where steam is the pressure-controlling fluid phase, a so-called vapor-dominated hydrothermal system. 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. This removes the latent heat of vaporization from this level, thereby cooling the rock and allowing more heat to rise from depth. Steam moves upward through fractures in the rock and is possibly superheated by the hot surrounding rock. At the top and sides of the system, heat is lost from the vapor to the

TABLE 1

GEOTHERMAL RESOURCE CLASSIFICATION
(Modified from White and Williams, 1975)

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

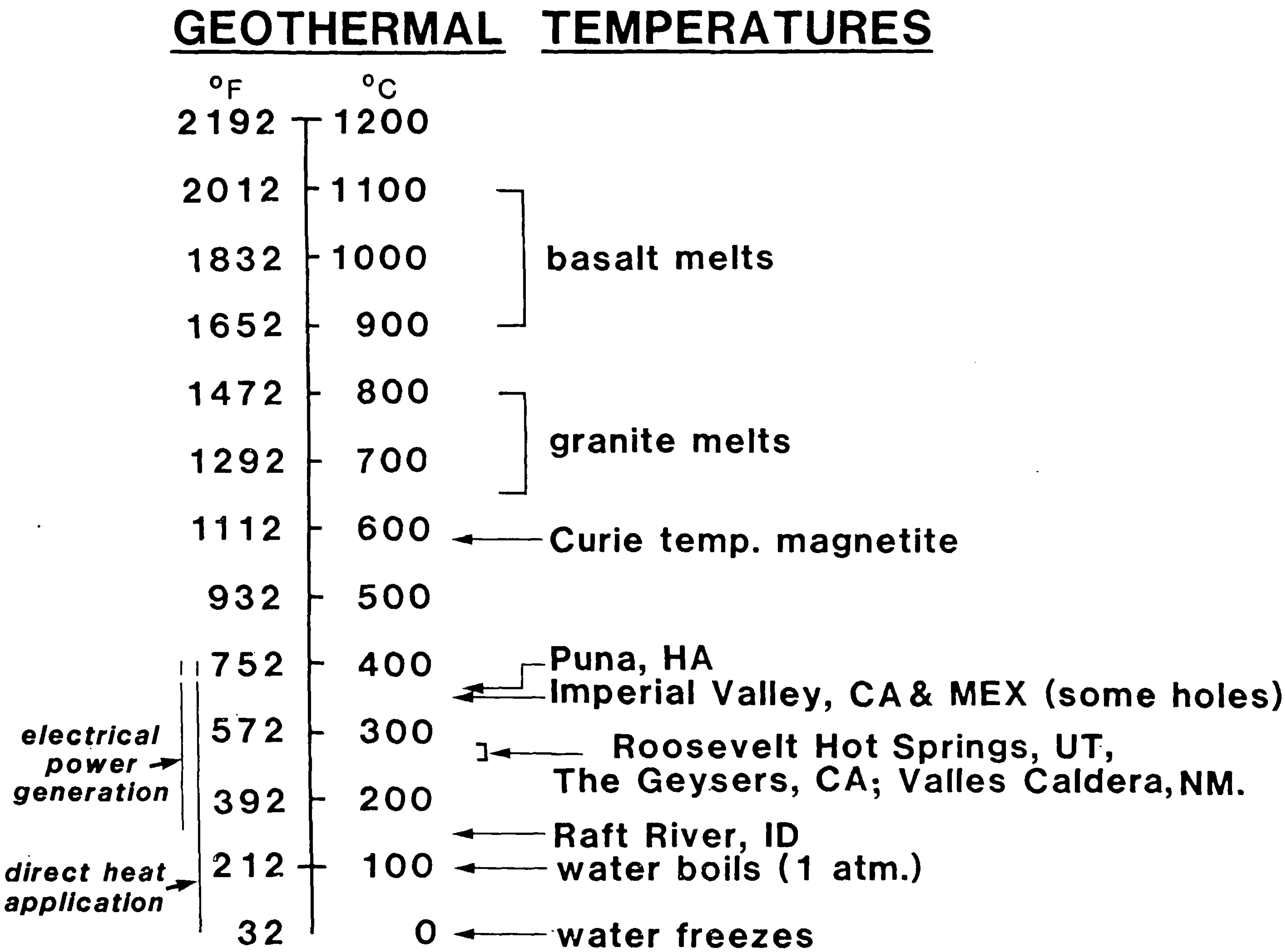

Figure 7

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VAPOR DOMINATED GEOTHERMAL RESERVOIR

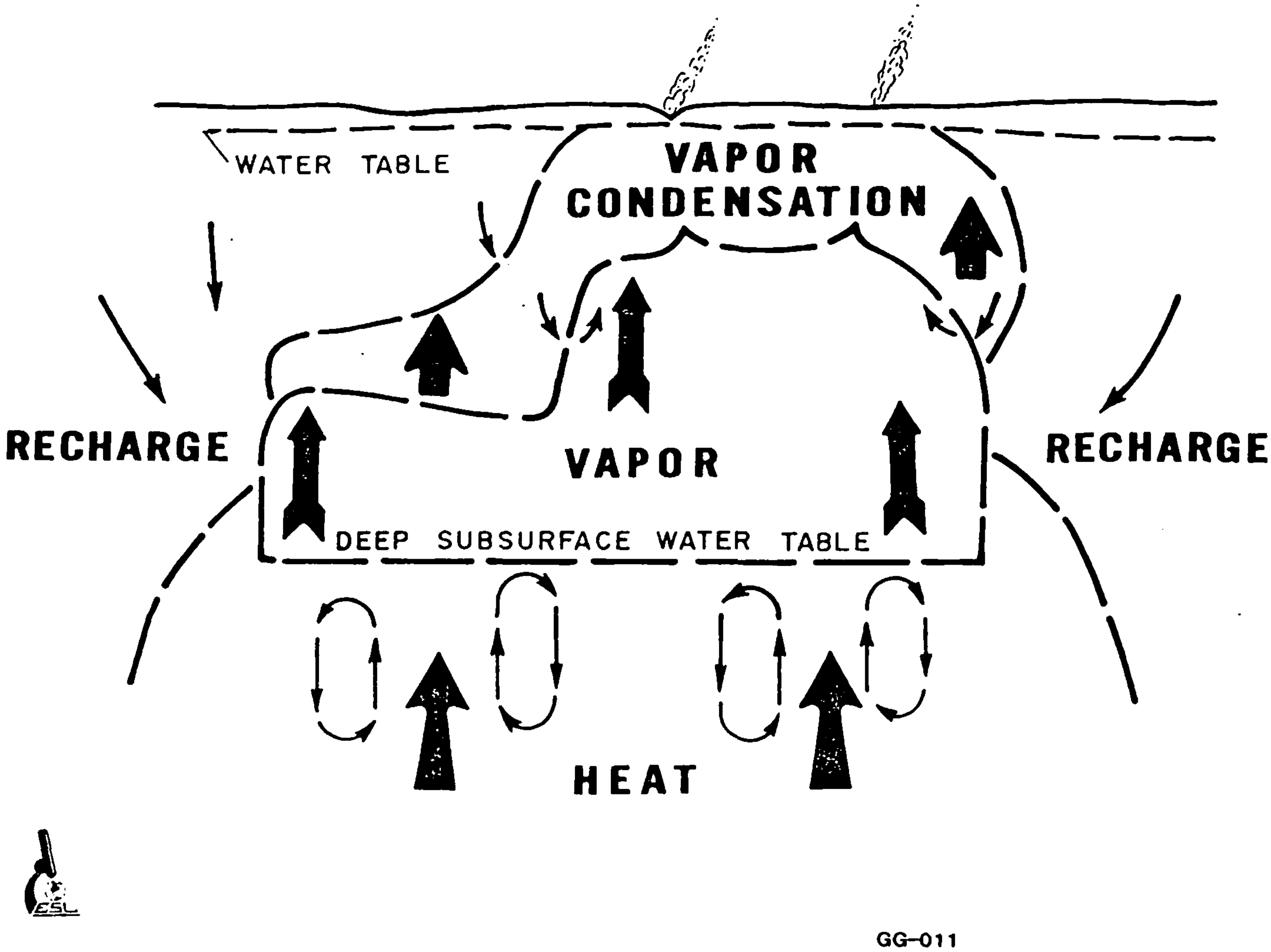


Figure 8

cooler 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. If an open fracture penetrates to the surface, steam may vent. Water lost to the system is 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 is controlled by the vapor phase and 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 steam 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 and above the reservoir is either naturally low or has been decreased by deposition in the fractures and pores of minerals from the hydrothermal fluid to form a sealed zone around the reservoir.

The Geysers geothermal area in California (see Figs. 20 and 21 and the discussion below) 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 about 1200 MWe (megawatts of electrical power, where 1 megawatt = 1 million watts). 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 vapor-dominated 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 by industry because they are generally easier and less expensive to develop than the more common water-dominated system discussed below.

Figure 9 schematically illustrates a <u>high-temperature, hot-water-dominated hydrothermal system</u>. Models for such systems have been discussed by White et al. (1971), Mahon et al. (1980), Henley and Ellis (1983), and Norton

HYDROTHERMAL SYSTEM IN VOLCANIC TERRANE

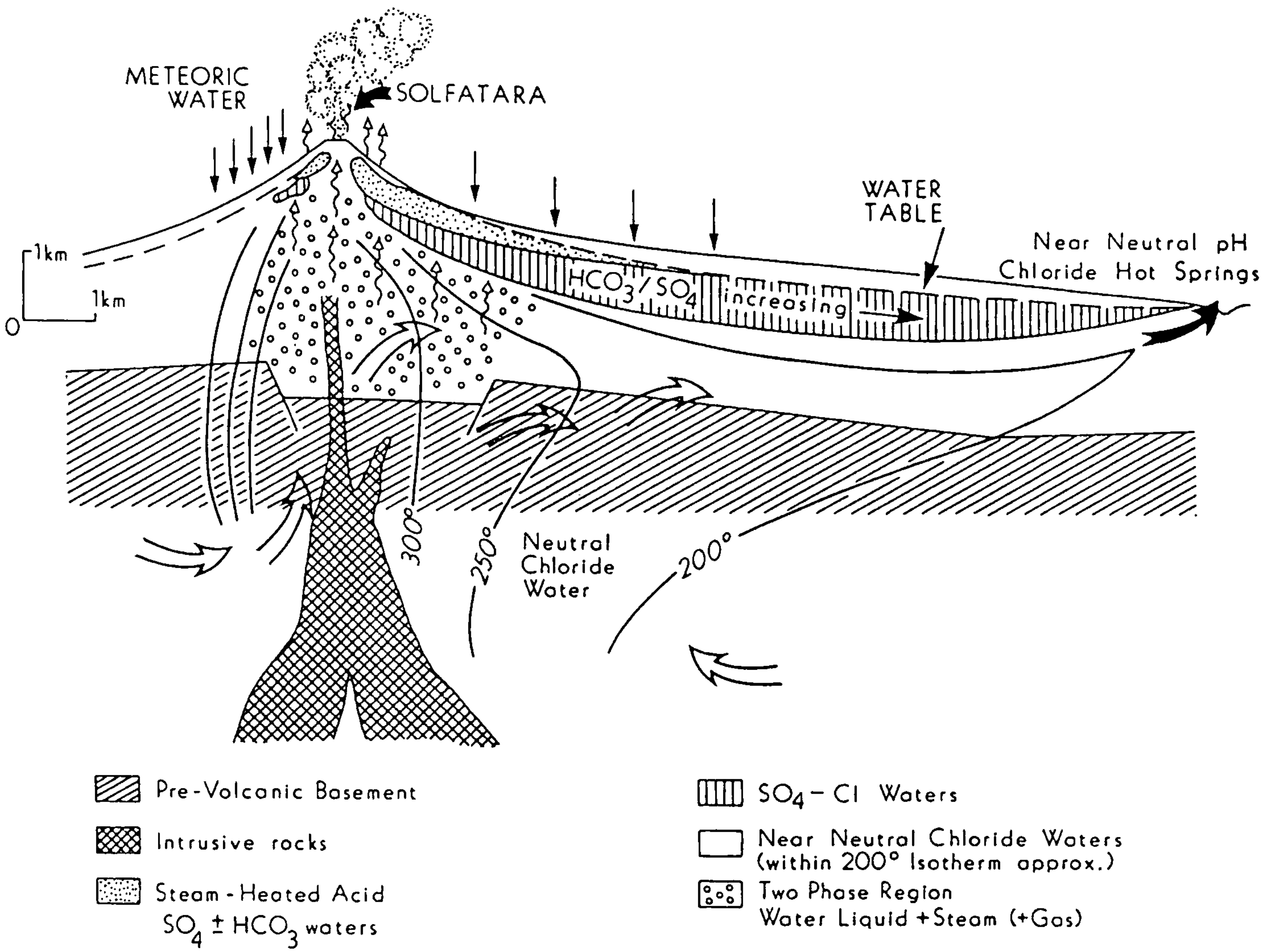


Figure 9

(1984), among others. 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. Rapid convection produces uniform temperatures over large volumes of the reservoir. In some parts of the system, boiling may occur and a two-phase region (water and steam) may exist, but the pressure in the system is controlled by the water. Pressure therefore increases much more rapidly with depth than it does in a vapor-dominated system. 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 geothermal fluids of minerals in fractures and pore spaces. Surface manifestations of such geothermal systems 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 encounter tight, hot rocks with hot water inflow from the rock into the well bore mainly along open fractures. Areas where different fracture or fault 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.

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

The vertical pressure and temperature gradients in most high-temperature (i.e. > 200°C) hydrothermal convection systems lie near the curve of boiling point versus depth for saline water, and sporadic boiling occurs in many systems. Because boiling concentrates such acidic gases as CO_2 and H_2S in the steam, the oxygenated meteoric fluids overlying a boiling reservoir are heated and acidified. This process may lead to the deposition of clays and the formation of fluids having a distinct $NaHCO_3(-SO_4)$ chemical character.

The general structure of high-temperature systems associated with andesitic stratovolcanos (e.g., the Cascade Range, U.S.A.; Ahuachapan, El Salvador), silicic or bimodal volcanic regimes (e.g., Coso, California; Steamboat Hot Springs, Nevada; the Taupo volcanic zone, New Zealand) and sedimentary basins (e.g., the Imperial Valley, California, and Mexicali Valley, Mexico) are shown in Figures 9, 10, and 11, respectively. The mineral assemblages produced by the thermal fluids significantly alter the physical properties of the reservoir rocks. The six factors temperature, fluid composition, permeability, and to a lesser extent, pressure, rock type, and time each control the distribution and type of hydrothermal alteration (Browne, 1978). The alteration minerals are strongly zoned in most systems. Beneath the water table, clay minerals, quartz and carbonate are the dominant secondary minerals below temperatures of about 225°C. Chlorite, illite, epidote, quartz and potassium feldspar are important at higher temperatures. In the highest-temperature fields (above 250°C), metamorphism to the greenschist or higher facies may occur, resulting in significant densification of the reservoir rocks. Precipitation of silica may occur through cooling of the hot brine. The porosity and permeability of the silicified rocks are thereby considerably reduced, which can effectively seal the sodium chloride reservoir and prevent its expansion or appearance at the surface. However, steam and gas may be able to move through the sealed boundary and to interact with meteoric water above. The product of this interaction is usually a near-neutral pH sodium bicarbonate-sulfate water that forms a hot, secondary geothermal reservoir. Although the bicarbonate-sulfate waters may constitute an exploitable resource, it is the deep chloride water that is the prime hydrothermal resource.

Fumaroles may vent CO_2 and H_2S at the surface, which interact with meteoric water to produce highly acidic waters that cause advanced argillic alteration of near-surface rocks. Intense alteration of this type may extend

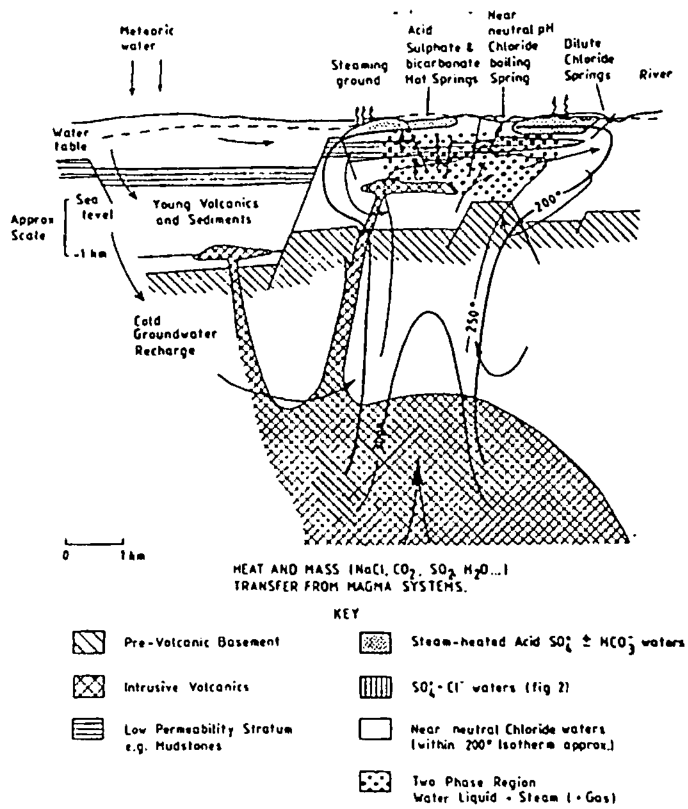

Meteoric water
Steaming ground
Acid Sulphate & bicarbonate Hot Springs
Near neutral pH Chloride boiling Spring
Dilute Chloride Springs
River
Water table
Sea level
Approx Scale
-1 km
Young Volcanics and Sediments
Cold Groundwater Recharge
200°
250°
0 1 km
HEAT AND MASS (NaCl, CO_2, SO_2, H_2O...) TRANSFER FROM MAGMA SYSTEMS.
KEY
Pre-Volcanic Basement
Intrusive Volcanics
Low Permeability Stratum e.g. Mudstones
Steam-heated Acid $SO_4^=$ ± HCO_3^- waters
$SO_4^=$-Cl^- waters (fig 2)
Near neutral Chloride waters (within 200° isotherm approx.)
Two Phase Region Water Liquid + Steam (+ Gas)

Figure 10

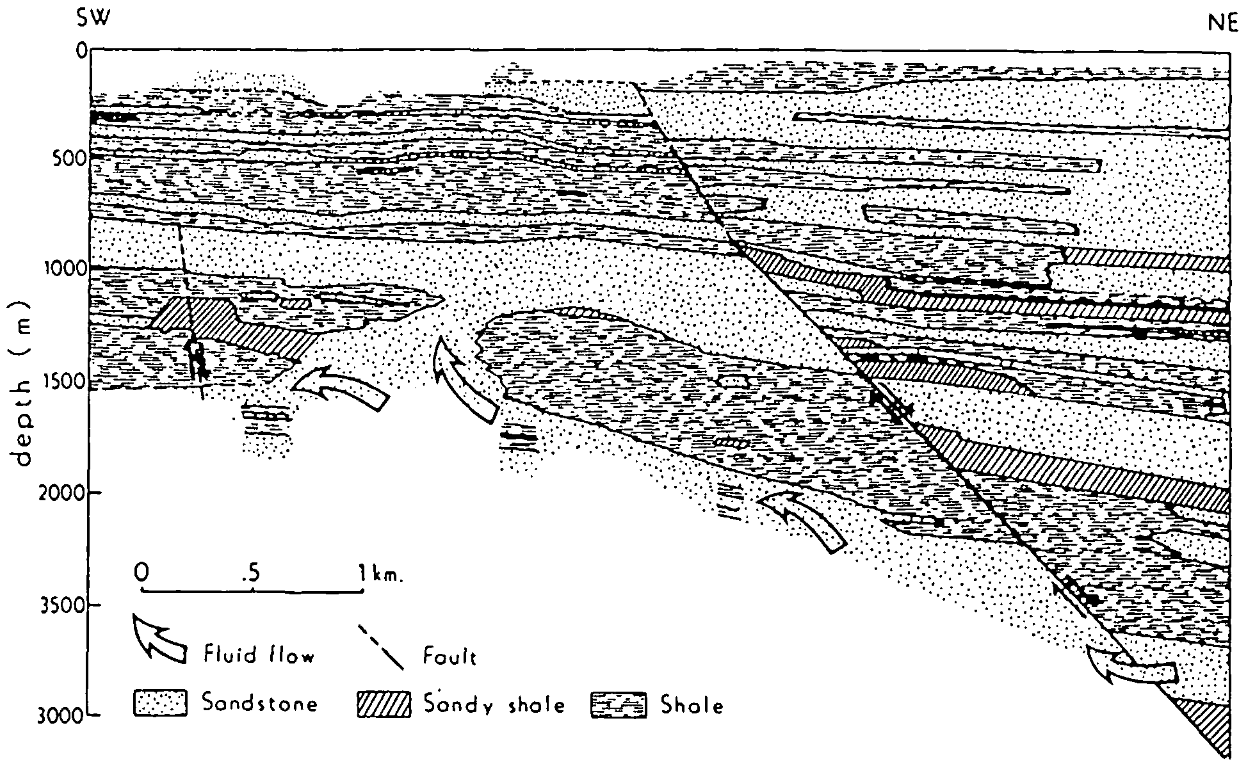


Figure 11. Fluid Flow Model of Cerro Prieto, Mexico.

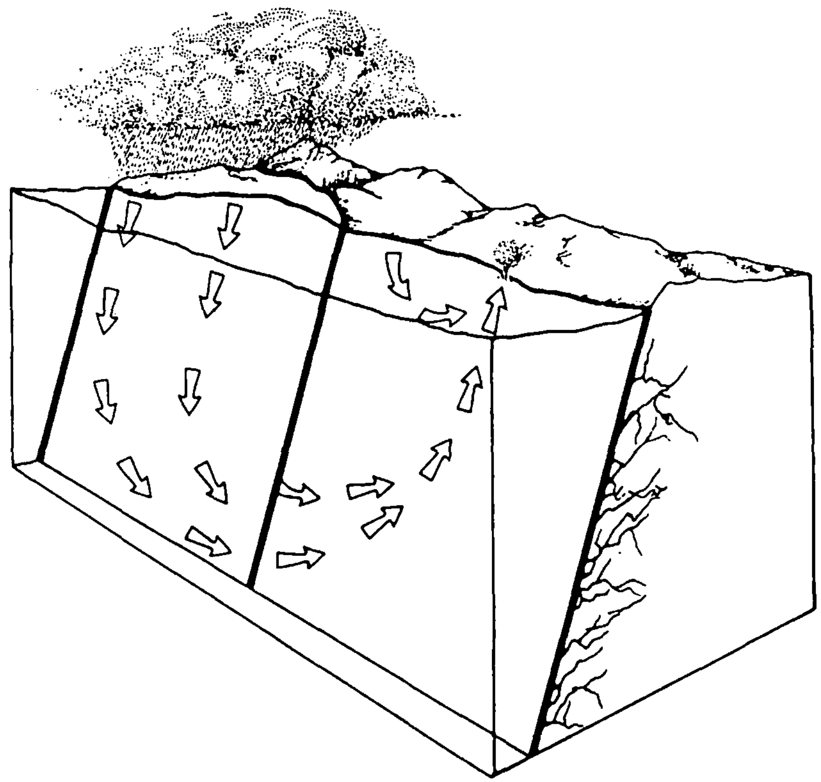
to depths of hundreds of meters below the surface in areas such as Cove Fort-Sulphurdale, Utah, where the water table is deep (Ross and Moore, 1985).

Outflow of the deep NaCl fluid may occur at a considerable distance from the hottest portion of a hydrothermal system. These chloride brines may emerge as boiling springs, frequently surrounded by silica deposits, or as a non-boiling mixture of local groundwater, with geothermal bicarbonate and chloride fluids. Because the solubility of calcite decreases with increasing temperature and decreases when CO_2 is released to the atmosphere, $CaCO_3$ in the form of travertine often precipitates where thermal waters mix with groundwaters and/or reach the surface.

Virtually all of industry's geothermal exploration effort in the United States is presently directed at locating vapor- or water-dominated hydrothermal systems of the types described above having temperatures above 200°C. A few of these resources are capable of commercial electrical power generation today. Current surface exploration techniques are generally conceded to be inadequate for discovery and assessment of these resources at a fast enough pace to satisfy the reliance the U.S. may ultimately put upon them for alternative energy sources. Development of better and more cost-effective techniques is badly needed.

The fringe areas of high-temperature vapor- and water-dominated hydrothermal systems often produce water of low and intermediate temperature. These lower-temperature fluids are suitable for direct-heat applications and may also be used for electrical power production with the newer binary technology typified by Ormat systems, for example. 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. 12). 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. This type of warm water resource is especially prevalent in the western U.S. where active faulting keeps conduits open to depth.

Sedimentary Basins/Regional Aquifers. Some basins are filled to depths



MODEL OF DEEP CIRCULATION HYDROTHERMAL RESOURCE

Figure 12

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of 10 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. The Madison group carbonate rock sequence of widespread occurrence in North and South Dakota, Wyoming, Montana, and northward into Canada contains warm waters that are currently being tapped by drill holes for space heating and agricultural purposes. In a similar application, 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 (Varet, 1982). Many other areas of occurrence of this resource type are known worldwide.

Geopressured Resources. Geopressured resources consist of deeply buried fluids 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 13 gives a few typical parameters for geopressured reservoirs and illustrates the origin of the above-normal fluid pressure. These geopressured fluids, found mainly in the Gulf Coast of the U.S. (Fig. 16), generally 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.

Industry has a great deal of interest in development of geopressured resources, although they are not yet economic. The U.S. Department of Energy, Geothermal Technology Division, is currently sponsoring development of exploitation technologies. This program is being managed by the Idaho National Engineering Laboratory of DOE in Idaho Falls, Idaho.

Radiogenic Resources. Research that could lead to development of radiogenic geothermal resources in the eastern U.S. has been done following ideas developed at Virginia Polytechnic Institute and State University (Costain et al., 1980). The eastern states coastal plain is blanketed by a layer of thermally insulating sediments. In places beneath these sediments, rocks are

GEOPRESSURED GEOTHERMAL RESOURCE

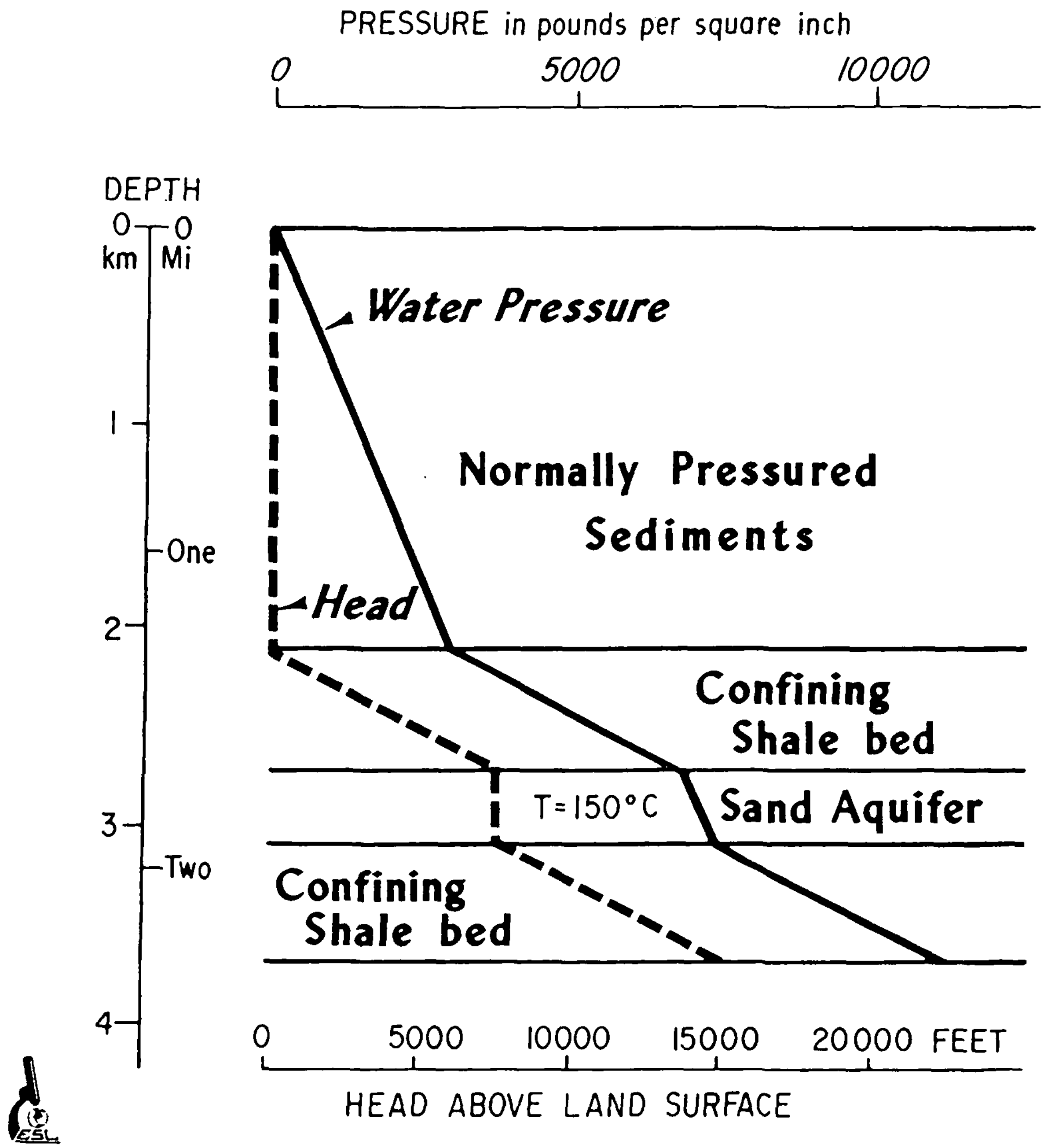


Figure 13

GG-001

believed to occur that have an anomalously high heat production due to high natural content of radioactive elements. These rocks represent old intrusions of once molten material that have long since cooled and crystallized. Geophysical and geological methods for locating such radiogenic rocks beneath the sedimentary cover have been partly developed, and very limited drill testing of the geothermal target concept (Fig. 14) has been completed under DOE funding. These resources may ultimately yield low- to intermediate-temperature geothermal water suitable for space heating and industrial processing. Availability of such a resource could mean a great deal to the eastern U.S. where energy consumption is high and where no shallow, high-temperature hydrothermal convection systems are known or expected to occur. Geophysical and geological data indicate that radiogenically heated rock bodies may be reasonably widespread.

Hot Dry Rock Resources. Hot dry rock resources 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 few pore spaces or fractures, and therefore contain little water and no interconnected permeability. The feasibility and economics of extraction of heat for electrical power generation and direct uses from hot dry rocks is presently the subject of a $150 million research program at the U.S. Department of Energy's Los Alamos National Laboratory in New Mexico (Smith and Ponder, 1982). Batchelor (1982) describes similar successful research conducted in England. Both projects indicate 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. During formation of the fracture system, its orientation and extent are mapped using passive seismic geophysical techniques. A second borehole is located such that it intersects the fracture system. Water can then be circulated down one hole, through the fracture system where it is heated, and up the second hole (Fig. 15). Fluids at temperatures of 150°C to more than 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.

RADIOGENIC GEOTHERMAL RESOURCE

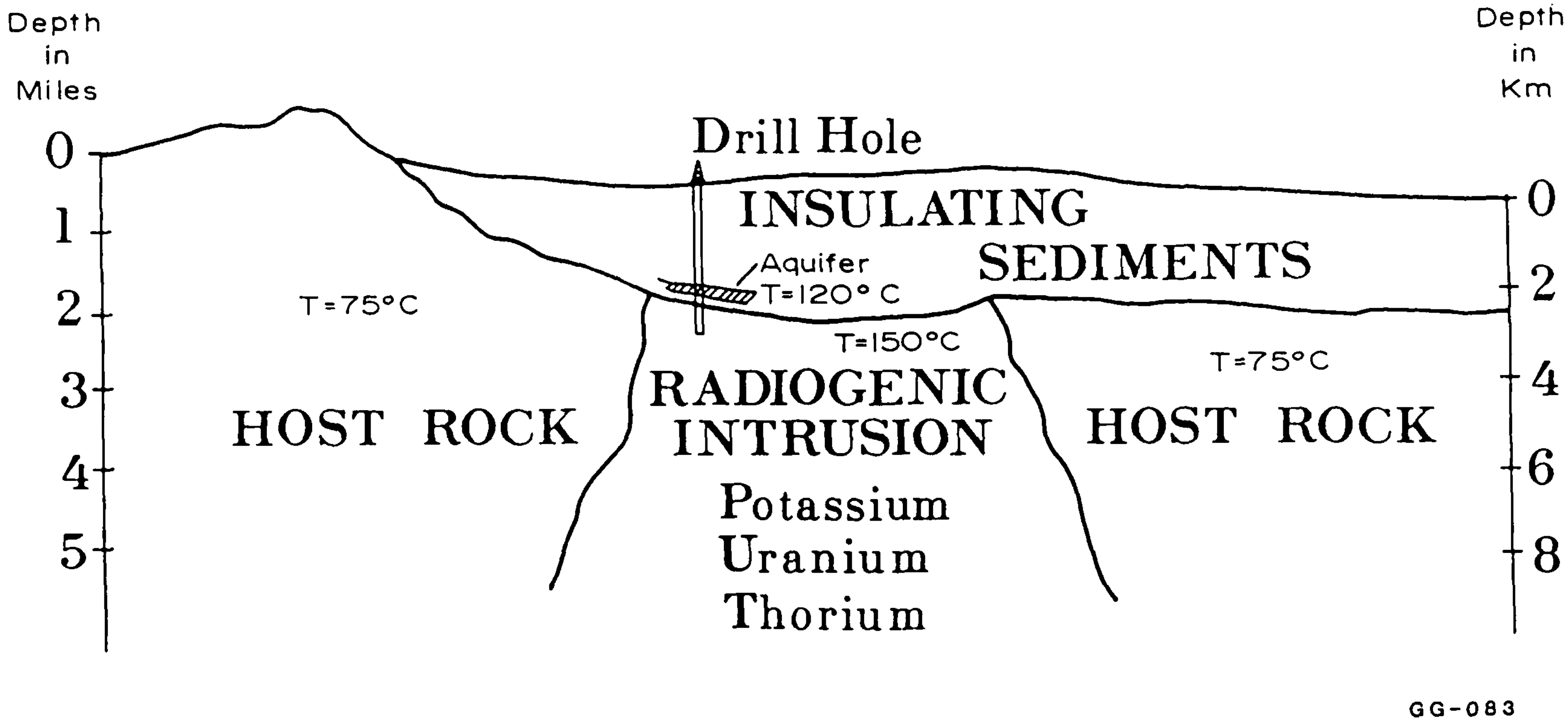
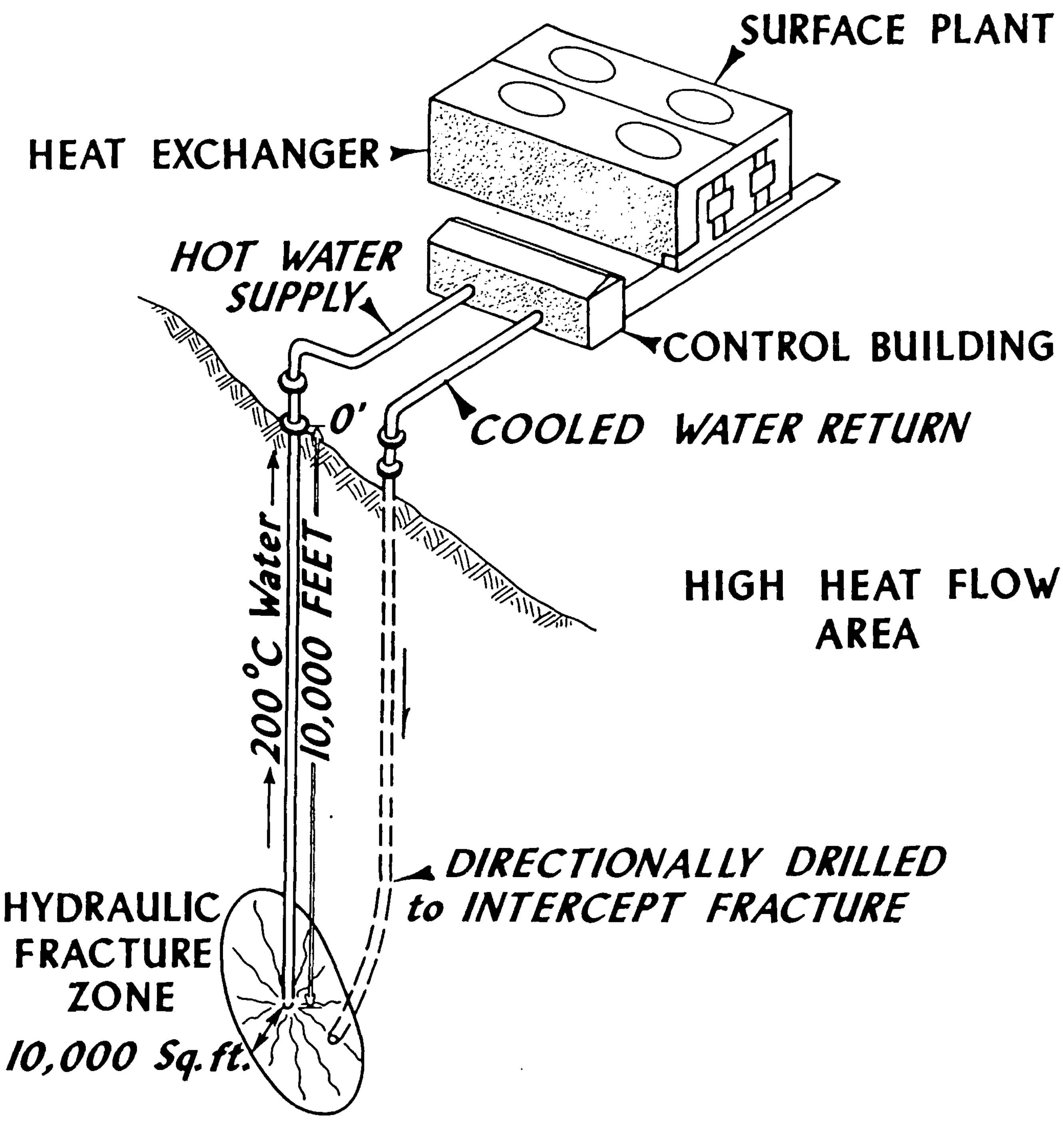


Figure 14



HOT DRY ROCK GEOTHERMAL RESOURCE

Figure 15 GG-081

Molten Rock. Experiments are underway at the U. S. Department of Energy's Sandia National Laboratories in Albuquerque, New Mexico to learn how to extract heat energy directly from molten rock. Techniques for locating a shallow, crustal magma body, drilling into it and implanting heat exchangers or possibly direct electrical converters remain to be developed (Carson and Allen, 1984).

Neither these experiments nor those of the hot, dry rock type described above are expected to result in economic energy production in the near future. In Iceland, however, where geothermal energy was first tapped for space heating in 1928, economic technology has been demonstrated for extraction of thermal energy from young lava flows (Björnsson, 1980). A heat exchanger constructed on the surface of the 1973 lava flow on Heimaey in the Westman Island group, recovers steam which results from downward percolation of water applied at the surface above hot portions of the flow. The space heating system which uses this energy has been operating successfully for over eight years.

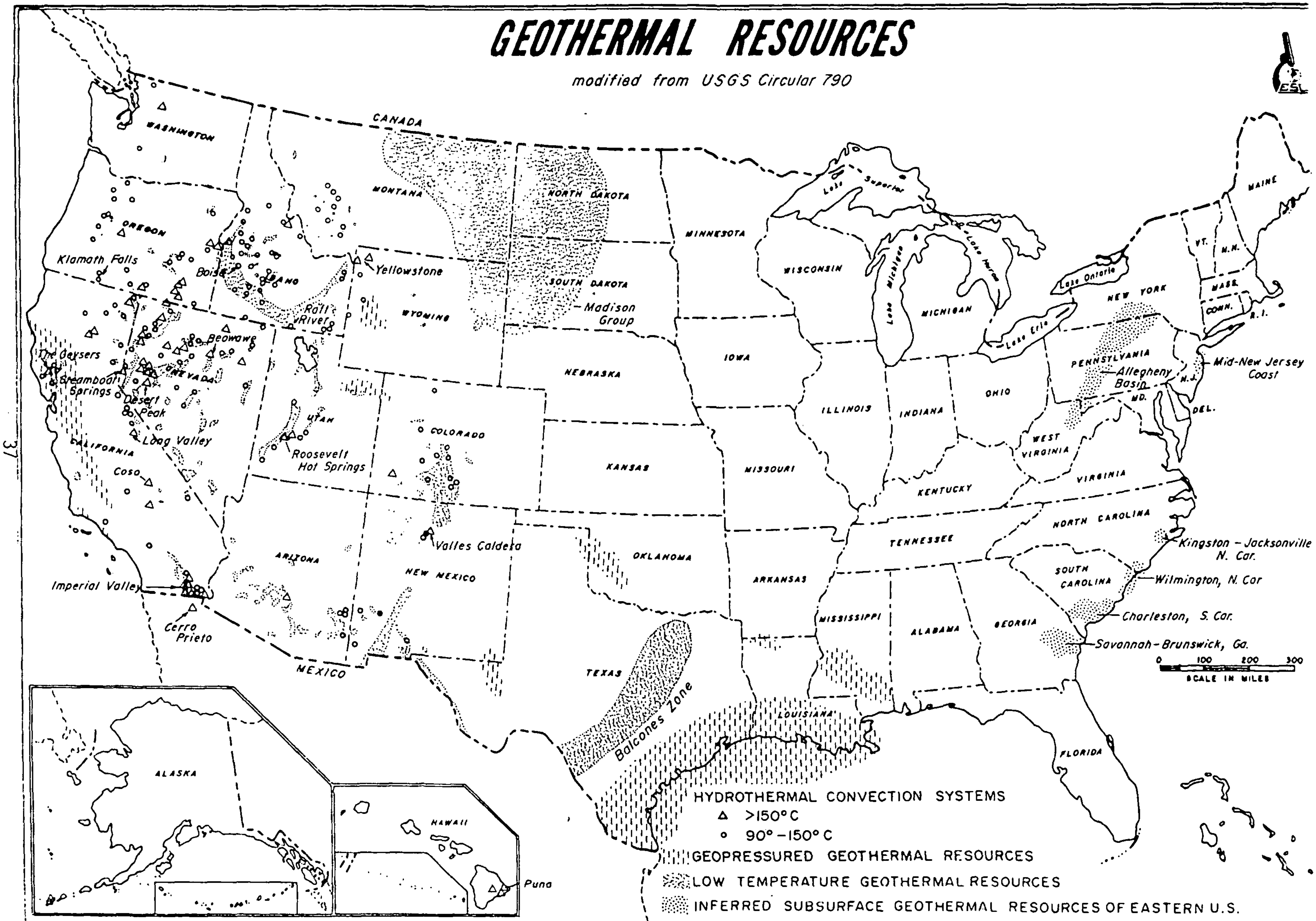
GEOTHERMAL RESOURCES IN THE CONTINENTAL UNITED STATES

Figure 16 displays the distribution of various resource types in the 48 contiguous states. Information for this figure was taken mainly from Muffler et al. (1978) and Reed (1982), where more detailed discussions and more detailed maps can be found. Not shown are locations of hot dry rock or magma resources because very little is known. In addition, it should be emphasized that the present state of knowledge of geothermal resources of all types is poor. Because of the very recent emergence of the geothermal industry, insufficient exploration has been done to define properly the resource base. Each year brings more resource discovery, so that Figure 16 will rapidly become outdated.

Figure 16 shows that most of the known hydrothermal resources and all of the presently known sites that are capable or believed to be capable of electric power generation from hydrothermal convection systems are in the western half of the U. S. The preponderance of thermal springs and other surface manifestations of underlying resources is also in the west. Large areas underlain by warm waters in sedimentary rocks exist in Montana, North and South Dakota, and Wyoming (the Madison Group of aquifers), but the extent and potential of these resources is poorly understood. Another important large area, much of which is underlain by low-temperature resources, is the northeast-trending Balcones zone in Texas. The geopressured resource areas of the Gulf Coast and surrounding states are also shown. Resource areas indicated in the eastern states are highly speculative because only one site has been drill tested to actually confirm their existence, which is only inferred at present.

Regarding the temperature distribution of geothermal resources, low- and intermediate-temperature resources are much more plentiful than are high-temperature resources. There are many, many thermal springs and wells that have water at temperature only slightly above the mean annual air temperature, which is the temperature of most non-geothermal shallow ground water. Resources having temperatures above 150°C are infrequent, but represent important occurrence. Muffler et al. (1978) show a statistical analysis of the temperature distribution of hydrothermal reosources and conclude that the cumulative frequency of occurrence increases exponentially as reservoir temperature decreases (Fig. 17). This relationship is based only on data for known

Figure 16



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FREQUENCY OF OCCURRENCE VS TEMPERATURE

FOR GEOTHERMAL RESOURCES

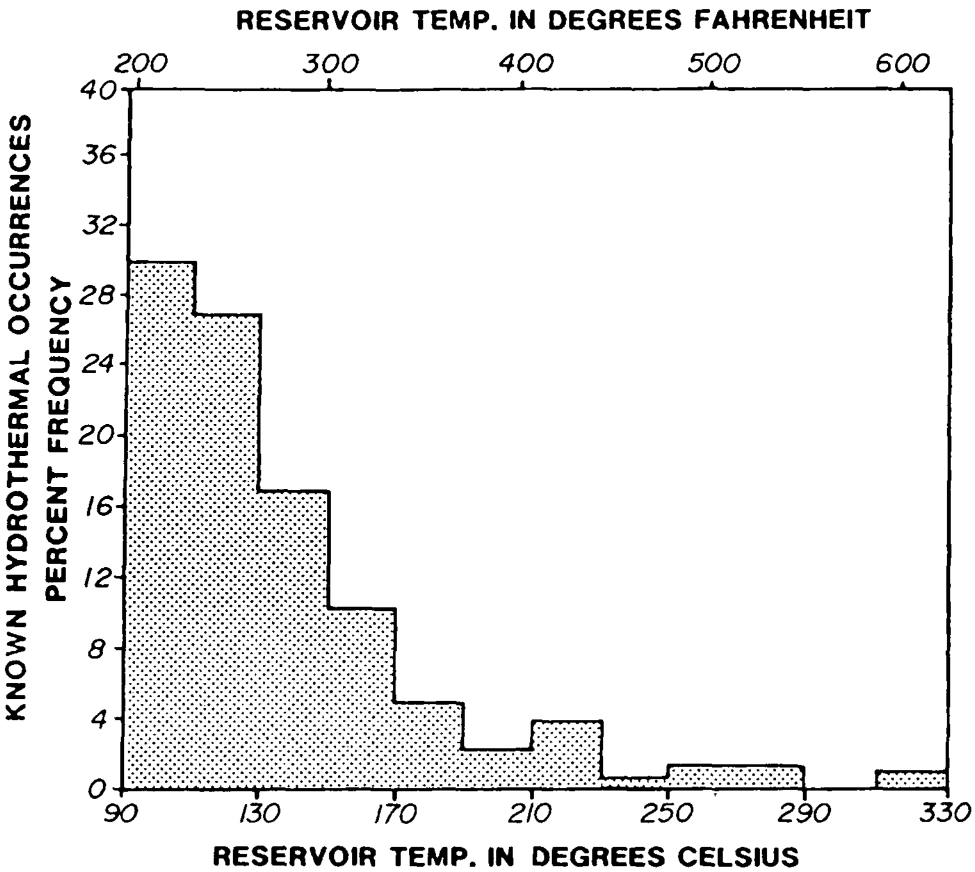


Figure 17

GG-007

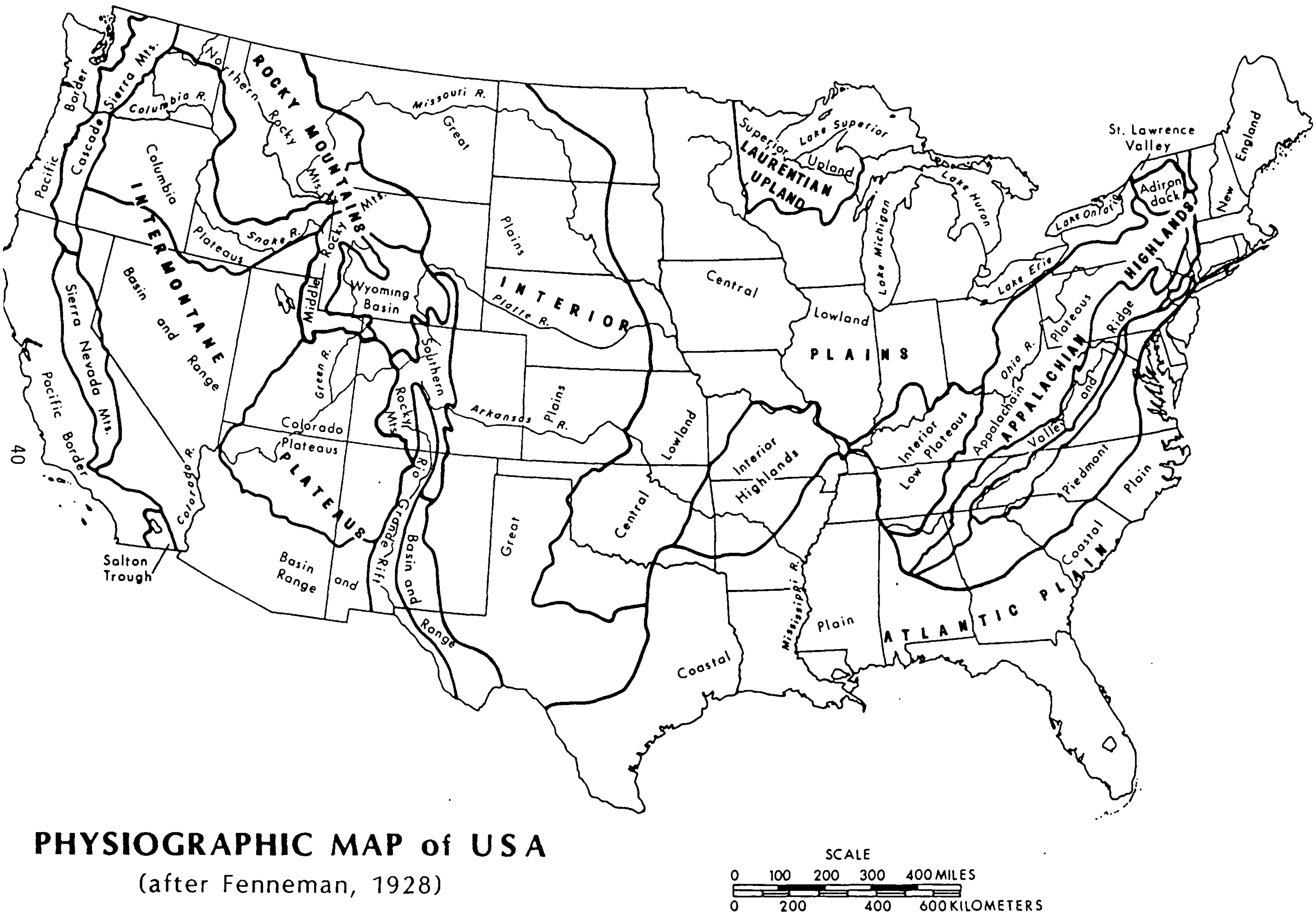
occurrences having temperatures 90°C or higher. It is firmly enough established, however, that we can have confidence in the existence of a very large low-temperature resource base, most of which is undiscovered.

Let us consider the known geothermal occurrences in a bit more detail, beginning in the Western U. S. The reader should refer to Figure 18 for locations of some of the geologic provinces discussed.

Salton Trough/Imperial Valley, CA

The Salton Trough is the name given an area along the landward extension of the Gulf of California. It is composed of the Imperial Valley in the U.S. and the Mexicali Valley in Mexico. The Salton Sea Trough is an area of complex, currently active plate tectonic geologic processes. As shown on Figure 2, the crest of the East Pacific Rise spreading center is offset repeatedly northward up the Gulf of California by transform faulting. Both the rise crest and the transform faults come onto the continent under the delta of the Colorado River (Fig. 19) and the structure of the Salton Trough suggests that they underlie the trough. The offsetting faults show right-lateral movement and trend northwestward, parallel to the strike of the well-known San Andreas fault. Elders (1979) and the contributing authors for his guidebook give summaries of the geothermal systems that occur in the Salton Trough.

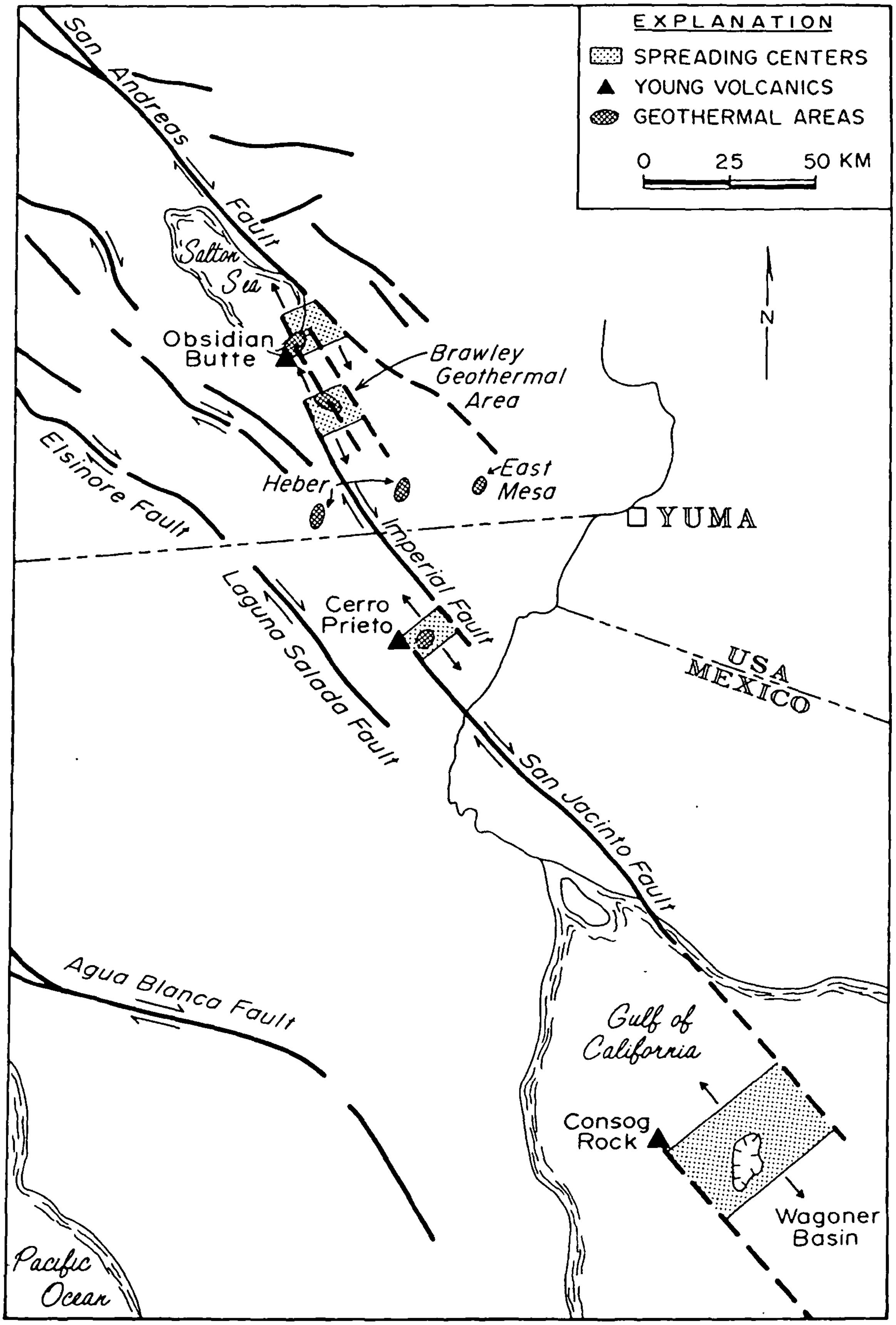
The Salton Trough has been an area of subsidence since Miocene times. During the ensuing years sedimentation in the trough has kept pace with subsidence, with shallow water sediments and debris from the Colorado River predominating. At present, 3 to 5 km of poorly-consolidated sediment overlie a basement of Mesozoic crystalline rocks that intruded Paleozoic and Precambrian sedimentary rocks. Detailed analysis of drilling data and of surface and downhole geophysics indicates that at least some of the known geothermal occurrences (Cerro Prieto, Brawley and the Salton Sea) are underlain by "pull-apart basins" apparently caused by crustal spreading above a local section of the East Pacific Rise crest (Elders, 1979). Very young volcanic activity has occurred at Cerro Prieto where a rhyodacite volcanic cone is known, and along the southern margin of the Salton Sea where rhyolite domes occur. The Salton Sea domes have an approximate age of 60,000 years (Muffler and White, 1969). The Cerro Prieto volcano has been difficult to date but may be about 10,000 years old (Wollenberg et al., 1980). Faulting is occurring at the present



PHYSIOGRAPHIC MAP of USA
(after Fenneman, 1928)

Figure 18

USA-002



MAJOR STRUCTURES OF SALTON TROUGH

(after Palmer et al., 1975)

Figure 19

GG-086

time as evidenced by the many earthquakes and earthquake swarms recorded in the Salton Trough.

The Cerro Prieto field is the best understood geothermal occurrence in the Salton Trough because of the drilling done there and its history of production. We may take it has an example of a Salton Trough resource type (refer to Fig. 11). The field is water-dominated and the more than 60 wells produce from depths of 1.5 to over 3 km. Fluid temperatures range from about 200°C to over 350°C (Alanso et al., 1979). The rocks are composed of an upper layer of unconsolidated silts, sands and clays, and a layer of consolidated sandstones and shales overlying the crystalline basement (Puete Cruz and de la Pena, 1979). Two principal reservoir horizons occur in sandstones within the consolidated sequence and enhanced production has been noted in the vicinity of faults, indicating that fracture permeability is important, although inter-granular permeability due to dissolution of minerals by the geothermal fluids is believed to be important also (Lyons and Van de Kamp, 1980). Reservoir recharge is apparently from the northeast and east and consists, at least partly, of Colorado River water (Truesdell et al., 1980).

The geothermal fluid from Cerro Prieto, after steam separation, contains about 25,000 ppm total dissolved solids. This figure is much lower than some of the other resources in the Salton Trough. For example, the Salton Sea area contains 20 to 30 percent by weight by solids (Palmer, 1975).

The heat source(s) for the several Salton Trough resources have not been found by drilling, although basalt dikes have been intersected in several areas, leading credence to the idea of upwelling basalt in pull-apart zones. Presumably, intrusion brings magma up into the depth range 5-10 km beneath the thermal anomalies.

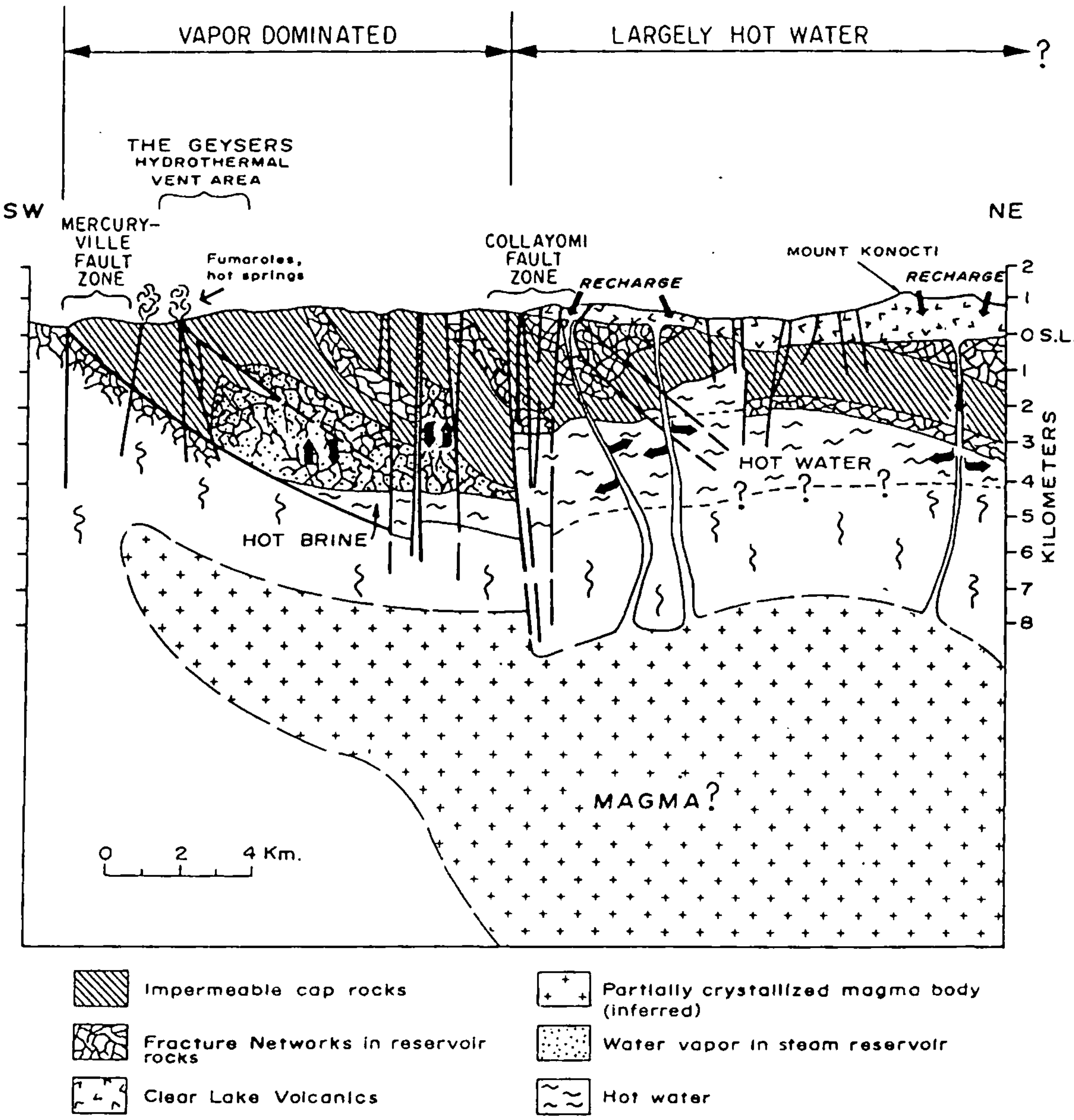
The Geysers, CA

The Geysers geothermal area is the world's largest producer of electricity from geothermal fluids with more than 1200 MWe on line and an additional several hundred scheduled. This area lies about 150 km north of San Francisco. The portion of the resource being exploited is a vapor-dominated field having a temperature of 240°C. The ultimate potential of the vapor-dominated system is presently believed to be around 2000 MWe. Associated with the vapor-dominated field are believed to be several unexploited hot water-dominated reservoirs whose volumes and temperatures are unknown (Fig. 20).

The geology of The Geysers area is complex, especially structurally. Reservoir rocks consist mainly of fractured greywackes, sandstone-like rocks consisting of poorly sorted fragments of quartzite, shale, granite, volcanic rocks and other rocks. The fracturing has created the permeability necessary for steam production in quantities large enough to be economically exploitable. Overlying the reservoir rocks, as shown in Figure 20, is a series of impermeable metamorphosed rocks (serpentinite, geenstone, melange and metagranite) that form a cap on the system. These rocks are all complexly folded and faulted. They are believed to have been closely associated with and perhaps included in subduction of the eastward-moving Pacific plate (Fig. 2) under the continent. This subduction apparently ended 2 to 3 million years ago.

As shown in Figure 21, the presently known steam field is confined between the Mercuryville fault zone on the southwest and the Collayomi fault zone on the northeast. The northwest and southeast margins are not definitely known. To the east and northeast lies the extensive Clear Lake volcanic field composed of dacite, rhyolite, andesite and basalt. The interval of eruption for these volcanics extends from 2 million years ago to 10,000 years ago, with ages progressively younger northward (Donnelly, 1977). The Clear Lake volcanics are very porous and soak up large quantities of surface water. It is believed that recharge of a deep, briny hot-water reservoir comes from water percolating through the Clear Lake volcanics, and that this deep reservoir may supply steam to the vapor-dominated system through boiling (Fig. 20) although these ideas are not universally supported by geologists and the deep water table has never been intersected by drilling.

The postulated water-dominated geothermal reservoirs do not occur everywhere in the Clear Lake volcanics. At several locations drill holes have found temperatures of 200°C at depths of only 2000 m, but the rocks are tight and impermeable (Goff, 1980). Fractured areas apparently host the water-dominated reservoirs at the Wilbur Springs district (Thompson, 1979), the Sulphur Bank Mine (White and Roberson, 1962) and other smaller occurrences. Potential in The Geysers area for discovery of additional exploitable resources is good.

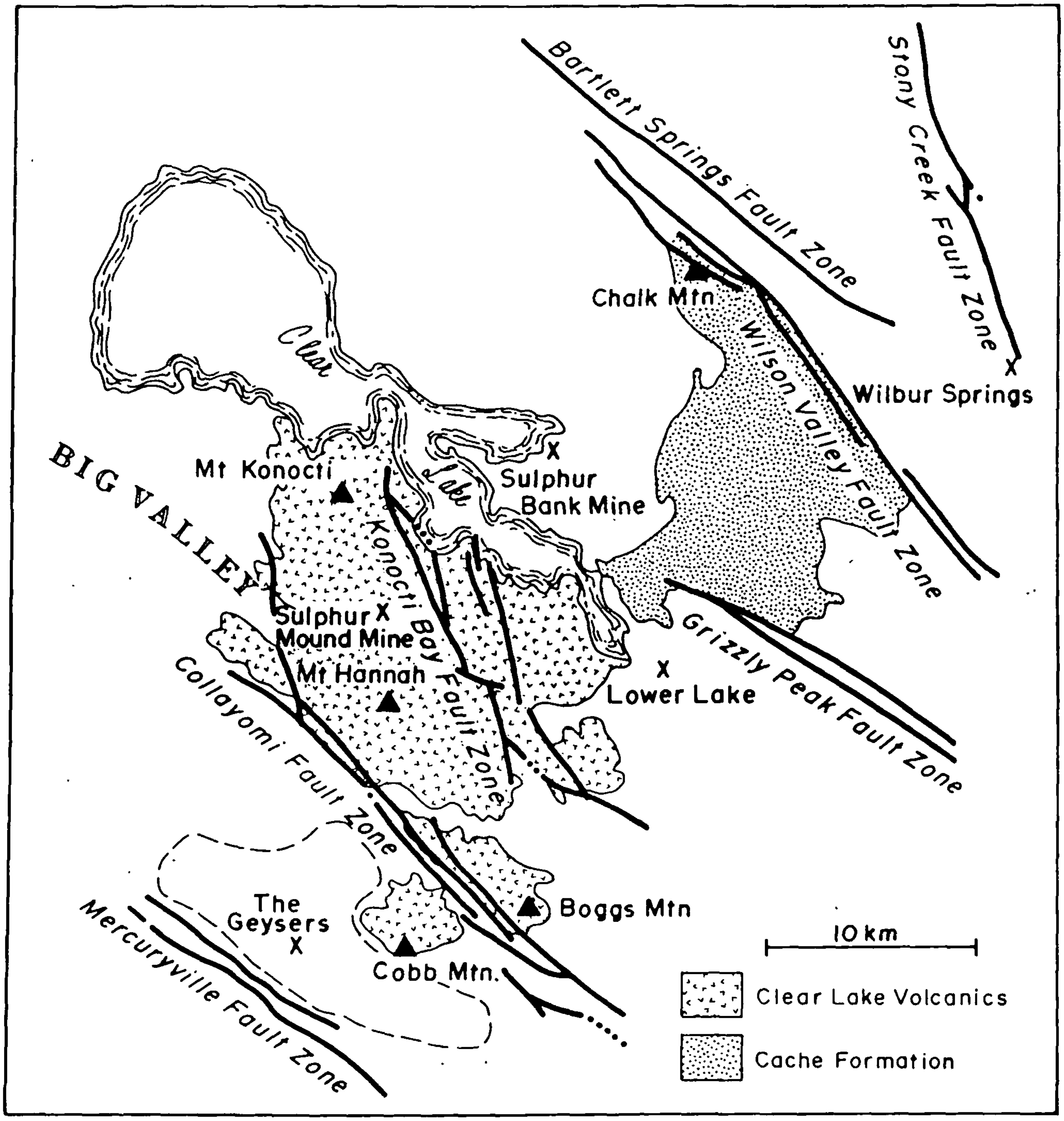


CRUSTAL MODEL FOR THE GEYSERS - CLEAR LAKE AREA, CA.

(after McLaughlin, 1977)

Figure 20

Ca/Ge-001



MAJOR STRUCTURES in THE GEYSERS-CLEAR LAKE AREA

(After Goff, 1980)

Ca/Ge-002

Figure 21

Basin and Range

The Basin and Range province extends from Mexico into southern Arizona, southwestern New Mexico and Texas on the south, through parts of California, Nevada and Utah, and becomes ill-defined beneath the covering volcanic flows of the Columbia Plateau on the north (Fig. 18). This area, especially the northern portion, contains abundant geothermal resources of all temperatures and has been one of the several areas of active exploration in the U. S. Resources along the eastern and western margins of the province appear to be both more abundant and of higher temperature. Electrical power is presently being generated from Roosevelt Hot Springs (20 MWe) and Cove Fort/Sulphurdale (3.2 MWe) in Utah. Candidate sites include Steamboat Springs, Dixie Valley, Desert Peak and Beowawe in Nevada and Coso, California. At both Desert Peak and Beowawe, plant construction is underway. Exploration is being or has been conducted at probably 20 or more sites in the Basin and Range, including, in addition to those named above, Tuscarora, McCoy, Baltazor, Leach Hot Springs, San Emidio, Soda Lake, Stillwater, and Humboldt House, Nevada; and Surprise Valley, and Long Valley Caldera, California. Direct application of geothermal energy for industrial process heating and space heating are currently operating n this area at several sites including Brady Hot Springs (vegetable drying), Reno (space heating) and Salt Lake City (greenhouse heating).

The reasons for the abundance of resources in the Basin and Range seem clear. This area, especially at its margins, is an active area geologically. Volcanism only a few hundred years old is known froms tens of areas, including parts of west central Utah on the east (Nash and Smith, 1977) and Long Valley caldera on the west (Rinehart and Huber, 1965). The area is also active seismically and faulting that causes the uplift of mountain ranges also serves to keep pathways open for deep fluid circulation at numerous locations. Rocks in the Basin and Range consist of Paleozoic and Mesozoic sandstones, limestones and shales that lie on Precambrian metamorphic and intrusive rocks. These rocks were deformed, complexly in some places, during the Nevadan and Laramide orogenies, and some base and precious metal deposits were formed. Beginning in mid-Tertiary times volcanic activity increased many fold with both basaltic and rhyolitic rocks being erupted. Extentional stresses also began to operate and a sequence of north-south mountain ranges were formed which separate valleys that have been filled with erosional debris from the mountains (Eardley, 1951). In some places more than 2 km offset has occurred along range-front faults, and the valleys may contain a hundred to as much as 3,000 m of unconsolidated erosional debris. This activity persists to the present time.

As an example of a Basin and Range hydrothermal system we will discuss Roosevelt Hot Springs, although it should not be supposed to be typical of all high-temperature occurrences in this province. This geothermal area has been studied in detail for the past six years (Nielson et al., 1978; Ward et al., 1978). The oldest rocks exposed (Figs. 22 and 23) are Precambrian sedimentary rocks that have been extensively metamorphosed. These rocks were intruded during Miocene time by granitic rocks (diorite, quartz monzonite, syenite and granite). Rhyolite volcanic flows and domes were emplaced during the interval 800,000 to 500,000 years ago. The area has been complexly faulted by north- to northwest-trending high-angle faults and by east-west high-angle faults. The Negro Mag fault is such an east-west fault that is an important controlling structure in the north portion of the field. The north-trending Opal Mound fault apparently forms the western limit of the system. The oldest fault system is a series of low-angle denudation faults (Fig. 23) along which the upper plate has moved west by about 600 m and has broken into a series of discrete blocks. Producing areas in the southern portion of the field are located in zones of intersection of the upper plate fault zone with the Opal Mound and other parallel faults. Producing zones in the northern part of the region are located at the intersection of north-south and east-west faults. The permeability is obviously fracture controlled.

Seven producing wells are shown on Figure 22 and more have been drilled recently. Fluid temperature is up to 260°C and the geothermal system is water-dominated. Average well production is perhaps 318,000 kg/hr (700,000 lbs/hr). Plans call for building up a base of experience with the 20 MWe power plant currently being operated there by Phillips Geothermal and Utah Power and Light, with two 50 MWe plants to be installed as knowledge of reservoir performance increases.

Cascade Range and Vicinity

The Cascade Range of northern California, Oregon, Washington and British Columbia is comprised of a series of volcanos, 12 of which have been active in

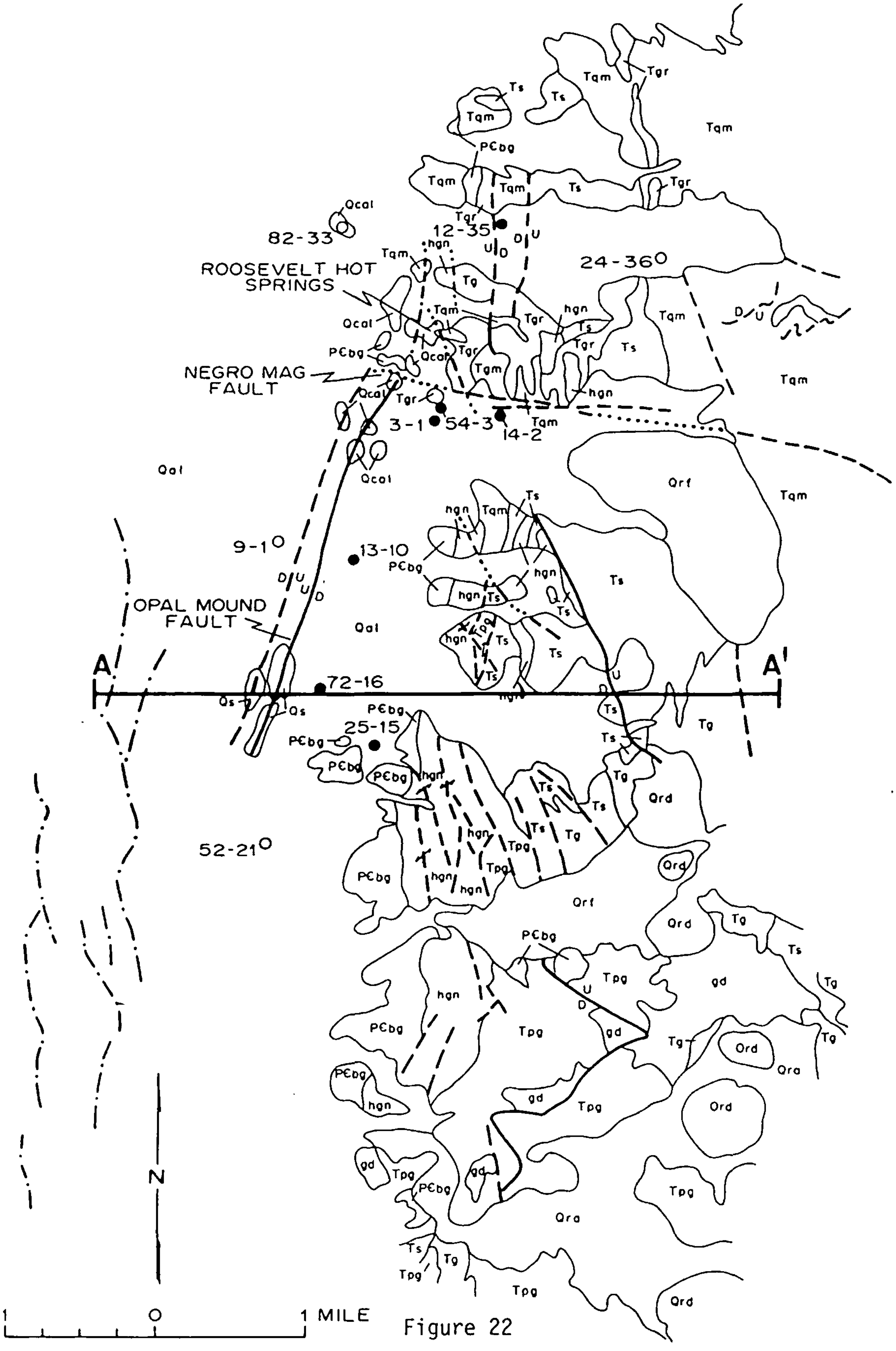


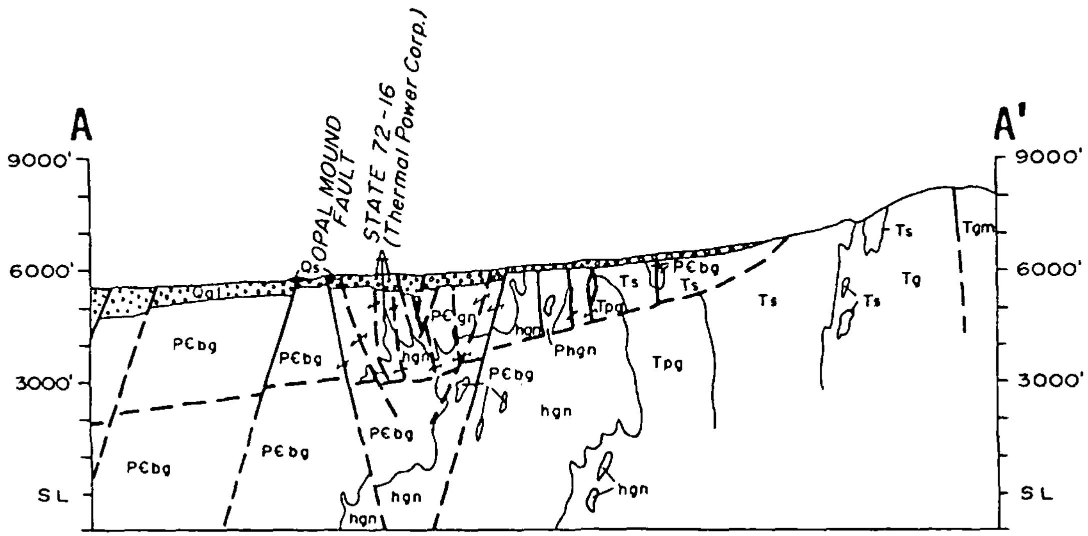
Figure 22

GEOLOGIC MAP ROOSEVELT HOT SPRINGS, UTAH

(from Nielson et al., 1978)

GEOLOGIC CROSS SECTION
ROOSEVELT HOT SPRINGS, UTAH

(from Nielson et al., 1978)



E X P L A N A T I O N

Qal	- alluvium	Tg	- granite
Qcal	- silicified alluvium	Ts	- syenite
Qs	- siliceous sinter	Tpg	- porphyritic granite
Qrd	- rhyolite domes	Tqm	- quartz monzonite
Qra	- pyroclastic deposits	gd	- biotite diorite
Qrf	- rhyolite flows	hgn	- foliated hornblende granodiorite
Tgr	- fine-grained granite	PЄbg	- banded gneiss

Ut/R-005a

Figure 23

historic times. The May 18, 1980 eruption of Mount St. Helens attests to be the youth of volcanic activity here. The Cascade Range lies above the zone of subduction of the Juan de Fuca plate beneath the North American plate, (Fig. 2) and magma moving into the upper crust has transported large amounts of heat upward. In spite of the widespread, young volcanism, however, geothermal manifestations are not as plentiful as one would suppose they should be. The high rainfall and snowfall in the Cascades are believed to suppress surface geothermal manifestations through downward percolation of the cold surface waters in the highly permeable volcanic rocks. In the absence of surface manifestation, discovery becomes much more difficult.

No producible high-temperature hydrothermal systems have yet been located in the Cascades. Geological and geochemical evidence indicates that a vapor-dominated system is present at Lassen Peak in California, but it lies within a national park, and will not be developed. A hydrothermal system having temperatures greater than 200°C has been located at Newberry Caldera in Oregon through research drilling sponsored by the U. S. Geological Survey (Sammel, 1981), but the known the portion of the system lies within the caldera will not be exploited for environmental reasons.

Industry's exploration effort so far in the Cascades has been minimal, but has increased somewhat in the last several years as leases have been issued. The Department of Energy is currently sponsoring a cost-shared drilling program with industry to encourage more subsurface exploration in the Cascades. To date, one hole has been drilled in the south flank of Newberry volcano by GeoOperator, and results of that drilling will be made public soon.

The use of geothermal energy for space heating at Klamath Falls, Oregon is well known (Lund, 1980), and numerous hot springs and wells occur in both Oregon and Washington. Potential for discovery of resources in all temperature categories is great. Priest (1983) evaluated the geothermal geology of the Oregon Cascades in a very useful publication.

Columbia Plateaus

The Columbia Plateaus area is an area of young volcanic rocks, mostly basalt flows, that cover much of eastern Washington and Oregon and continue in a curved pattern into Idaho, following the course of the Snake River.

There are no hydrothermal resources having temperatures greater than 90°C known through drilling in this area. However, there are numerous warm springs and wells that indicate the presence of geothermal resources potentially suitable for direct heat uses.

Snake River Plain

The basalt flows and other volcanic deposits of the Snake River Plain are an extension of the Columbia Plateau eastward across southern Idaho to the border with Wyoming. The plain is divided into a western part and an eastern part. Thermal waters occur in numerous wells and springs in the western portion, especially on or near the edges of the plain. Geochemically indicated resource temperatures exceed 150°C at Neal Hot Springs and Vale, Oregon and Crane Creek, Idaho, but indicated temperatures for most resources are lower. Younger volcanic rocks occur in the eastern part of the plain, but no high-temperature resources (T>150°C) are yet identified, although numerous areas have warm wells and springs. This part of the plain is underlain by a high-flow cold-water aquifer that is believed to mask surface geothermal indications.

Direct use of hydrothermal energy for space heating is famous at Boise, Idaho, where the Warm Springs district has been heating homes geothermally for almost 100 years (Mink et al., 1977). Also in this area is the Raft River site where the Idaho National Engineering Laboratory of DOE constructed and operated a 5 MWe binary demonstration plant on a hydrothermal resource whose temperature is 147°C. This project is currently inoperative and the plant has been sold.

Rio Grande Rift

The Rio Grande Rift is a north-trending tectonic feature that extends from Mexico through central New Mexico and ends in central Colorado. It is a down-dropped area that has been filled with volcanic rocks and erosional debris from the bordering plateaus and mountains. The rift began to form in late Oliogocene times, and volcanic and seismic activity have occurred subsequently to the present. Young volcanism, faulting and high heat flow characterize the area today.

There are several low- and intermediate-temperature hydrothermal convection systems in this area, but the only high-temperature system that has been drill tested to any significant extent and where production is proven is a hot water-dominated system in the Valles caldera (Dondanville, 1978; Nielson and Hulen, 1984). Surface manifestations at the Baca No. 1 location in the caldera include fumaroles, widely distributed hot springs and gas seeps. Hydrothermal alteration extends over 40 km^2. Deep drilling has encountered a hydrothermal convection system in fractured Tertiary volcanic, Paleozoic sedimentary and Precambrian granitic rocks at an average depth of 2 to 3 km.. Temperatures as high as 300°C have been recorded. An attempt by DOE, Union Geothermal and Public Service Company of New Mexico to build a demonstration plant at that location failed when the steam supply proved to be inadequate. Also located near the caldera is the site of Los Alamos National Laboratory's hot dry rock experiment at Fenton Hill. Both the hot dry rock site and the hydrothermal convection system(s) probably derive their heat from magma that has provided the material for the several episodes of volcanism that created the caldera structure.

Elsewhere in the Rio Grande Rift, there are numerous hot springs and wells. Discovery potential appears to be high, although there are no known sites where fluids in excess of 150 to 170°C is indicated by present data (Harder et al., 1980).

Madison and other Aquifers

Underlying a large area in western North and South Dakota, eastern Montana and northeastern Wyoming are a number of aquifers that contain thermal waters. These aquifers have developed in carbonates and sandstones of Paleozoic and Mesozoic age. The permeability is both intergranular and fracture controlled in the case of the sandstones (e.g. the Dakota Sandstone) and fracture and solution cavities in the carbonates (e.g. the Madison Limestone). At least some of the aquifers will produce under artesian pressure. Depths to production vary widely but average perhaps 2,000 ft. Temperatures are 30-80°C (Gries, 1977) in the Madison but are lower in other shallower aquifers such as the Dakota. Direct use of the thermal water is being made at a few locations today, and it is evident that the potential for further development is substantial.

Balcones Zone, Texas

Thermal waters at temperatures generally below 60°C occur in a zone that trends northeasterly across central Texas. Many of the large population centers are in or near this zone, and there appears to be significant potential for geothermal development in spite of the rather low temperatures.

An initial assessment of the geothermal potential has been documented by Woodruff and McBride (1979). The thermal waters occur in a band broadly delimited by the Balcones fault zone on the west and the Luling-Mexia-Talco fault zone on the east. In many locations the thermal waters are low enough in content of dissolved salts to be potable, and indeed many communities already tap the warm waters for their municipal water supplies.

The geothermal aquifers are mostly Cretaceous Sandstone units, although locally thermal waters are provided from Cretaceous limestones and Tertiary sandstones. The thermally anomalous zone coincides with an ancient zone of structural weakness dating back more than 200 million years. The zone has been a hinge line with uplift of mountain ranges to the north and west and downwarping to the south and east. Sediments have deposited in the area of downwarping, and the rate of sedimentation has kept pace with sinking, keeping this area close to sea level. Structural deformation of the sediments, including faulting and folding, and interfingering of diverse sedimentary units have resulted in the complex aquifer system of today.

The source of the anomalous heat is not known with certainty but several postulates are (Woodruff and McGride, 1979): 1) deep circulation of ground waters along faults; 2) upwelling of connate waters, originally trapped in sediments now deeply buried; 3) stagnation of deep ground waters owing to faults that retard circulation; 4) local hot spots such as radiogenic heat sources (intrusions) within the basement complex, or; 5) other loci of high heat flow.

A minor amount of direct use is being made of these waters at present, and potential for further development is good.

Eastern Half of U. S.

Hydrothermal resources in other areas of the continental U. S. besides those mentioned above are very poorly known. There is believed to be potential for thermal waters of about 100°C at a number of locations along the Atlantic Coastal plain associated with buried intrusions that are generating anomalous heat through radioactive decay of contained natural uranium, thorium and potassium. Examples of such areas are shown on Figure 16 at Savannah-Brunswick, Charleston, Wilmington, Kingston-Jacksonville and the mid-New Jersey Coast. One drill test of such an area (Delmarva Peninsula near Washington, D. C.) has been conducted by DOE with inconclusive results regarding amount of thermal water that could be produced. Less than a dozen warm springs and wells are known at present. The Allegheny Basin is outlined on Figure 16 because it has potential for thermal fluids in aquifers buried deeply enough to be heated in a normal earth's gradient. Parts of Ohio, Kansas, Nebraska and Oklahoma as well as other states are believed to have potential for low-temperature fluids. No geothermal drill tests have been conducted, however.

Hawaiian Islands

The chain of islands known as the Hawaiian archipelago stretches 2500 km in a northwest-southeast line across the Pacific Ocean from Kure and Midway Islands to the Big Island of Hawaii. Built of basaltic volcanic rocks, this island chain boasts the greatest volcanic masses on earth. The volcano Kilauea rises 9800 m above the floor of the ocean, the world's largest mountain in terms of elevation above its base. The Kilauea, Mauna Loa and other vents on the big island are in an almost continual state of activity, but by contrast volcanos on the other islands have shown little recent activity. Haleakala on the island of Maui is the only other volcano in the state that has erupted in the last few hundred years, and the last eruption there was in 1790 (MacDonald and Hubbard, 1975).

Several of the Hawaiian islands are believed to have geothermal potential. The only area where exploration has proceeded far enough to establish the existence of a hydrothermal reservoir is in the Puna district near Kapoho along the so-called "East Rift", a fault zone on the east flank of Kileaua. Here a well was completed to a depth of 1965 m (Helsley, 1977) with a bottom-hole temperature of 358°C. Little is known in detail of the reservoir(s) at present, but they are believed to be fracture-controlled and water-dominated. A 3 MWe generator is currently being operated at the site. Exploration is currently underway by several companies in areas adjacent to the operating plant.

Elsewhere on the islands potential for occurrence of low- to moderate-temperature resources has been established at a number of locations on Hawaii, Maui and Oahu, although little drilling to prove resources has been completed (Thomas et al., 1980).

Alaska

Little geothermal exploration work has been done in Alaska. A number of geothermal occurrences are located on the Alaska Peninsula and the Aleutian Islands and in central and southeast Alaska. The Aleutians and the Peninsula overlie a zone of active subduction (Fig. 2), and volcanos are numerous. A hydrothermal system was located at Makushin volcano on the island of Unalaska (Reeder et al., 1985) and the island of Adak is also believed to have good discovery potential.

Low- and moderate-temperature resources are indicated in a number of locations in Alaska by occurrence of hot springs (Muffler et al., 1978). One area that has been studied in more detail and has had limited drilling is Pilgrim Hot Springs (Turner et al., 1980). This site is 75 km north of Nome, Alaska. Initial drilling has confirmed the presence of a hot water reservoir about 1 km^2 in extent that has artesian flow rates of 200-400 gallons/minute of 90°C water. Geophysical data suggest that the reservoir is near the intersection of two inferred fault zones. Further exploration work will be required to determine the potential of this reservoir.

Potential for Geothermal Development

Muffler et al. (1978) have dealt with the problem of how much accessible resource exists in the U. S. both at known sites and those that are undiscovered. They conclude that the undiscovered resource base is on the order of 3 to 5 times greater than the resources known today. These figures do not include possible hot dry rock or other more speculative resources. Table 2 is a summary of the current estimate of the geothermal resource base as taken from Muffler et al. (1978). This table demonstrates our lack of resource knowledge through the ranges and relative amounts of undiscovered resources and through the many missing numbers. We can conclude, however, that the geothermal resource base is large in the U. S.

TABLE 2

GEOTHERMAL ENERGY OF THE UNITED STATES
After Muffler et al. (1978) Table 20

RESOURCE TYPE	ELECTRICITY (MWe for 30 yr)	BENEFICIAL HEAT (10^{18} joules)	RESOURCE (10^{18} joules)
Hydrothermal			
Identified	23,000	42	400
Undiscovered	72,000-127,000	184-310	2,000
Sedimentary Basins	?	?	?
Geopressured (N. Gulf of Mexico)			
Thermal			270-2800
Methane			160-1600
Radiogenic	?	?	?
Hot Rock	?	?	?

EXPLORATION AND RESOURCE EVALUATION

Geothermal exploration may be divided into two types:

1) Exploring for geothermal resource areas, that is, locating geothermal resource areas, and
2) Exploring within geothermal resource areas, that is, defining the lateral and vertical boundaries and the properties of the actual reservoir(s).

In each case, the central problem of the geoscientific work is to site wells that intersect the resource and learn as much as possible about it. The main difference between the two problems is one of scale since many of the techniques used are common to both.

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 geologies are 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 led 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, California, where the experience of locating and drilling hundreds of wells is available, the success rate for production is only about 80 percent. For wildcat geothermal drilling in relatively unknown areas the success rate is much lower -- about 15 percent for the Basin and Range Province of the western United States. The problem 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 system installation and maintenance. In many geothermal reservoirs, this means drilling into one or more fractures that are connected to the source area for the geothermal fluids. Although large blocks of rock in nature are 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 mostly indirect and provide only circumstantial evidence of the existence and location of the reservoir.

Geology

Collection of geologic data through surface geologic mapping and through logging of drill cuttings and core provides the basic data required for interpretation of all other exploration data. Often ignored or shortchanged in geothermal exploration, 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 (sedimentary rocks, plutonic rocks, volcanic rocks), (2) maps the structure within and among rock units (faults, fractures, folds, rock contacts), (3) studies the age relationships amoung rock units as shown by their mutual field relationships, (4) searches for evidence of geothermal activity, which evidence may range from obvious thermal springs, geysers and fumaroles to very subtle indications such as hydrothermal alteration of rocks, 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 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 1 million years old) volcanic rocks in the area that would indicate an underlying molten mass that could provide a source of heat?, (3) are there porous and permeable rock units or are there active faults or open rock contacts that could constitute a plumbing system?, and (4) is this a viable geothermal prospect area and if so what exploration techniques should be used next?

Stratigraphic Analysis. A thorough knowledge of the rock types in the prospecting area is fundamental. The geologist analyzes both surface outcrops and samples from drilling. He strives to identify rocks in the area that would make a good reservoir rock at depth, i.e. one that has adequate permeability or in which permeability may be developed. In a volcanic sequence, for example, sequences of young flows often are highly permeable whereas air fall or water-laid tuffs are easily altered to clay minerals and become impermeable. On volcanic islands, the portion of the rocks that were extruded in the atmosphere (subaerial) may be more permeable than those extruded under water (subàqueous). A volcanic island typically sinks during its formation, so the subaerial-subaqeous volcanic contact is found below sea level. The geologist will try to determine the affect that this change may have on permeability for a particular island and the expected depth of the contact. In areas like the Salton Trough, permeability is controlled by the type of rock (permeable sandstone or impermeable shale) and by its degree of metamorphism (high-temperature metamorphism causes the rocks to be brittle and to fracture whereas low-temperature metamorphism does not induce brittleness). It is obvious that an understanding of effects such as these is important to the success of a geothermal project.

Structural Analysis. A thorough knowledge of the structure of an area is important. Moore and Samberg (1979), for example, showed that at Cove Fort/Sulphurdale, Utah, much of the surface is covered by rock units that have slid into place from the east along an underlying nearly horizontal fault. Subsequent faulting has occurred along vertical faults, and the area now consists of separate fault blocks. One obvious implication from this discovery was that surface geology can be projected to depth only with great care. Faults can form zones of permeability if they fracture rock and create open spaces, or alternatively they can be filled with gouge, a rock flour that is quite impermeable. Gouge developed along faults can isolate the aquifers in individual fault blocks and decrease hydrologic communication across an area. Fluids trapped in sandstone aquifers between impermeable shale beds in isolated fault blocks can become highly pressured as the unit sinks through geologic time due to deposition of new sediments above. In such isolated blocks are found the geopressured resources of the Gulf Coast and elsewhere.

In places where faults intersect, permeability may be especially enhanced. It is important to determine the relative ages of faults and especially to be able to distinguish young faults and fractures from older ones. Older fualts are more likely to have had their open spaces filled by deposition of minerals. Relative ages of faults can sometimes be determined through detailed geologic mapping.

Age Dating. Certain minerals contain potassium, and a small percentage will be the naturally radioactive isotope K^{40}. This isotope decays to argon, $\nabla\theta^{40}$, with a half-life of about 1.2 billion years. By measuring the amount of A^{40} relative to the amount of K^{40} in a mineral, the time since A^{40} began to accumulate can be determined. In this way certain rocks can be dated. There are also other radioactive isotopes that can be used for dating. One must be careful about interpretation of the dates derived by these methods. In the case of K-Ar dating, for example, if the mineral being used for the dating has been heated sufficiently by a thermal event subsequent to its formation, the gaseous Ar may escape, thus resetting the radioactive clock to the date of the thermal event. Age dating has obvious use in geothermal exploration in terms of helping to locate young igneous rocks.

Geochemistry

A number of important exploration and reservoir production questions can be answered from studies of the chemistry of geothermal fluids and reservoir rocks, and so geochemistry plays a relatively important role in geothermal exploration (Ellis and Mahon, 1977). Geochemical reconnaissance involves sampling and analyzing waters and gases from hot springs and fumaroles in the area under investigation. The data obtained are then used to determine whether the geothermal system is hot-water or vapor-dominated, to estimate the minimum temperature expected at depth, to estimate the homogeneity of water supply, to infer the chemical character of the waters at depth, and to determine the source of recharge water. We will discuss some of the more important geochemical applications.

The processes causing many of today's high-temperature geothermal resources consist of convection of hot saline 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 (White, 1981).

Geothermal fluids contain a wide variety and concentration of dissolved constituents (Table 3). 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.

As geothermal fluids move through rocks, they interact chemically with the rocks, which themselves are usually chemically complex. Certain minerals in the reservoir rocks may be selectively dissolved by the fluids while other minerals may be precipitated from solution or certain chemical elements from the fluid may substitute for certain others within a mineral. These chemical/mineralogical changes in the reservoir rocks may or may not cause volume changes. Obviously, if the rock volume increases it must be at the expense of open space in the rock, which decreases 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 plugging. Thus, some hydrothermal systems form a sealed cap or a self-sealed zone above the reservoir and perhaps on the lateral boundaries also. In this self-sealing process, silica and calcium carbonate are the principal phases involved. The solubility of SiO_2 decreases with a decrease in temperature, with very little pressure effect. Calcite has a retrograde solubility, i.e., it is more soluble at low temperatures than at high temperatures. However, calcite solubility does increase rapidly with an increase in the partial pressure of carbon dioxide. Thus, as fluids which are saturated with calcium carbonate approach the surface, $CaCO_3$ is deposited as a result of the loss of CO_2. Other carbonate species such as dolomite ($MgCO_3$), as well as sulfates such as anhydrite ($CaSO_4$), show solubility relationships similar to those of calcite.

TABLE 3

REPRESENTATIVE ANALYSES OF GEOTHERMAL FLUIDS

Sample #	1	2	3	4	5	6	7	8	9	10	11	12
Temp °C	42	47	44.5	60		89	96		255	<260	292	316
pH		7.1	7.3	3.4	7.9	7.9	9.5		8.4			
SiO_2 (ppm)	52	11.3	157.3	136	289	293	373	400	690	563	705	400
Ca (ppm)	257	88.2	117.9	11.5	2.6	5.0	.8	10	17	8	592	28,000
Mg (ppm)	17	20.8	106	4.9	1.3	.8	.0	37	.03	<2	.6	54
Na (ppm)	578	11.3	228.9	5.7	247	653	230	117	1,320	2,320	6,382	50,400
K (ppm)		5.4	39.2	3.6	12.9	71	16	86	225	461	1,551	17,500
Li (ppm)	.5					.7	1.3		14.2	25.3	14.5	215
HCO_3 (ppm)		39.7	391		377	305	116	12.0		232	28	7,150
SO_4 (ppm)	932	57.3	748	126	340		89	414	36	72	<3.5	5
Cl (ppm)	625	7.4	110	7.4	9.6	865	30.	10	2,260	3,860	11,918	155,000
F (ppm)	2.8	.25	.59	.5	.8	1.8	15	8	8.3	6.8		15
B (ppm)	2.6					4.9	2.0	24.1			13.4	390
As (ppm)						2.7			4.8	4.3		12

Sample Descriptions:

1. Hot spring; Monroe Hot Springs, Utah (Mundorff, 1970). Actively depositing travertine.
2. Hot spring; Yunotani geothermal field, Japan (Parmentier and Hayashi, 1981).
3. Hot spring; Yunotani geothermal field, Japan (Parmentier and Hayashi, 1981).
4. Acid sulfate water; Yunotani geothermal field, Japan (Parmenteir and Hayashi, 1981).
5. Water discharged from well; Yunotani geothermal field, Japan (Parmenteir and Hayashi, 1981).
6. Hot spring; Steamboat Hot Springs, Nevada (White et al., 1971).
7. Hot spring; Beowawe, Nevada, includes 149 CO_3 (Roberts et al., 1967).
8. Water discharged from well; The Geysers steam field (Frye; in Geothermal Resources Council Technical Session 5, 1980).
9. Well 44, Wairakei, New Zealand (Ellis and Mahon, 1977); pH measured at 20°C.
10. Brine discharged from well 54-3, Roosevelt Hot Springs, Utah (Capuano and Cole, 1981).
11. Analyses calculated from flashed brine, well M-26, Cerro Prieto (Fournier, 1981).
12. Brine discharged from well 11D, Salton Sea Geothermal Field (Palmer, T.D., 1975).

Other factors may also affect the deposition of carbonate and sulfate and other minerals, such as variations in pH, total pressure and partial pressure of oxygen. For example, subsurface boiling, accompanied by loss of CO_2, may cause the deposition of calcite, while the deposition of anhydrite may reflect the occurrence of locally oxidizing conditions produced when upwelling fluids contact aerated non-thermal groundwater.

This chemically complex hydrothermal system is dynamic through time, that is, for any given volume element in the reservoir, the fluid composition varies slowly with time, bringing about variation in the rock composition, porosity and permeability. However, because the rate of fluid circulation is perhaps only a few centimeters per year, in most hydrothermal systems a state of chemical equilibrium or near-equilibrium is observed to exist between reservoir fluid and reservoir rocks (Capuano and Cole, 1981; Helgeson, 1969). Lack of equilibrium could be evidence for rapid movement of fluid through the reservoir.

Chemical and Mineral Zoning. The hydrothermal mineral assemblages of active geothermal systems are dominated by clays or zeolites at relatively low temperatures, and by chlorite, illite, K-feldspar and epidote (or wairakite) at higher temperatures (Table 4). Quartz, calcite, pyrite and anhydrite are frequently associated with these minerals, and appear to form readily at both high and low temperatures. As expected, the distributions of the clay and silicate minerals is strongly temperature-dependent. At the lowest temperatures, below about 180°C, the stable assemblage consists of dolomite, kaolinite, montmorillonite and interlayered illite/montmorillonite. With increasing temperature and depth, montmorillonite, dolomite, kaolinite, and interlayered illite/montmorillonite disappear, and at temperatures above about 150°-180°C, the typical assemblage is illite, chlorite, potassium-feldspar and quartz. The calcium-aluminosilicates, wairakite and epidote appear only in rocks above 230-250°C. Prehnite, actinolite, diopside and biotite characterize the highest temperature assemblages associated with temperatures above about 300°C. One very important result of this mineral zoning is that the higher-temperature mineral assemblages cause the rocks to become brittle, and they fracture easily under the influence of tectonic movement and stress. This creates and renews fracture permeability in the higher-temperature parts of some Imperial Valley systems. In systems where base temperature is below

TABLE 4

SOME HYDROTHERMAL MINERALS IN SELECTED GEOTHERMAL FIELDS[1]

	Imperial Valley, California	Yellowstone, Wyoming	The Geysers, California	Pauzhetsk, Kamchatka	Matsukawa, Japan	Otake, Japan	Tongonan, Philipines	Kawah Kamojang, Java	N. Z. Volcanic Zone	El Tatio, Chile	Low temp Iceland	High temp Iceland	Larderello, Italy
Quartz	x	x	x	x	x	x	x	x	x	x	r?	x	x
Cristobalite		x		x	x	x	x	x	x	x			
Kaolin group	d	x	x	x	x	x	x	x	x	x			
Montmorillonite	d	x		x	x	x	x	x	x	x			
Interlayered illite-mont.	x			x	x	x	x	x	x	x	x	x	x
Illite	x	x	x	x	x	x	x	x	x	x			
Biotite	x			x					x				
Chlorite	x	x	x	x	x	x	x	x	x	x	?	x	x
Celadonite		x		x						x	x		
Alunite			x	x	x	x	x		x				
Anhydrite	x		x	x	x	x	x	x	x	x		x	x
Sulfur			x	x	x		x		x				
Siderite			x	x			x		x	x			
Ankerite	x			x								x	
Analcime		x		x					x		x	x	
Wairakite	x		x	x		x	x	x	x			x	x
Laumontite		x		x	x	x			x	x	x	x	
Heulandite		x		x		x	x		x		x	x	
Mordenite		x		x					x		x	x	
Prehnite	x			x					x		r?	x	
Amphibole	x			x	x	x			x			x	
Epidote	x			x		x	x	x	x		r	x	x
Sphene	x			x			x	x	x				
Adularia	x	x	x	x		x	x	x	x	x			x
Albite	x			x			x	x	x	x		x	
Rutile				x	x	x							
Leucoxene			x	x	x		x		x				
Magnetite									x				
Hematite		x		x			x	x	x	x	x		
Pyrite	x	x	x	x	x	x	x	x	x	x	x	x	x
Pyrrhotite	x						x		x				x
Base-metal sulfides	x	x		x		x							x
Fluorite		x						x					

(1) From Browne, 1978

Note: d = detrital, r = relict.
[a] includes Cerro Prieto, Baja, California, Mexico.
[b] deposited in discharge pipes and channels.

about 200°C, this development of enhanced permeability may not be possible.

Although the metal contents of many geothermal brines are significant, with the exception of pyrite and, in places, pyrrhotite, base metal sulfides are relatively uncommon at depth even in the deeper parts of the explored systems which deposit metal sulfides at the surface. The more commonly observed base metal sulfide minerals found at depth include sphalerite and galena, although chalcopyrite, arsenopyrite, nickel glaucodot, cobaltite and silver telluride also occur in rocks of the Broadlands field in New Zealand. In general the base metal sulfides in the Broadlands are present in rocks whose temperature range from 265°-300°C, whereas pyrrhotite is present above about 150°C (Browne and Ellis, 1970). The distribution of pyrite is not sensitive to temperature.

The interpretation of the mineral assemblages found in many thermal systems is complicated by the presence of minerals formed during earlier, frequently unrelated, hydrothermal events. The Roosevelt Hot Springs thermal system provides a situation where at least two distinct hydrothermal events can be recognized; an earlier event related to intrusion of the Tertiary Mineral Mountains pluton, and the present hydrothermal system (Nielson et al., 1978). Cross-cutting veins, identified in drill chips suggest that the depositional histories of these events was complex. The reservoir rocks consist of Tertiary granitic rocks and Precambrian gneiss and schist containing potassium feldspar, quartz, plagioclase, biotite and hornblende. The hydrothermal minerals include clays, illite, chlorite, calcite, pyrite, quartz, hematite, epidote and anhydrite.

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

The acid sulphate springs are typically a surficial feature produced by the oxidation of hydrogen sulfide to sulfuric acid. Altered ground surrounding the acid springs and fumaroles provides a striking example of reactivity of the waters. The altered areas are typically bleached and converted to a siliceous residue containing native sulfur, cinnabar, yellow sulfate minerals, and clay minerals including kaolinite and alunite. Similar acid alteration can, however, also be formed at depths where steam heating of groundwaters occur.

Chemical Geothermometry. 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 dissipated, and (2) fluids encountered during drilling may indicate that higher temperatures may be found elsewhere.

In the usual reconnaissance application, water samples are taken for analysis from springs and wells in the vicinity of the prospect. Proper sampling technique is very important. The samples must be filtered and properly acidified for preservation until analysis. At each sample location, pH and temperature are measured at the time of collection.

Both quantitative and qualitative geothermometers are available. The basic assumptions in application of quantitative geothermometers are that equilibrium has been reached in temperature-dependent chemical reactions between the reservoir rock and the fluid, and that no changes occur in the fluid after it leaves the reservoir and is sampled by a well or at the site of a natural spring. In this case, the chemistry of the fluid will reflect the chemical equilibrium at reservoir temperature, and analysis with subsequent interpretation yields an estimate of this temperature.

Several major-element geothermometers have been used successfully for estimating subsurface temperature, and a review of the geothermometers was given by Fournier (1981). For example in certain geothermal areas, the silica content of geothermal fluids appears to be limited above about 180°C by the solubility of quartz (SiO_2) and below 180°C by the solubility of amorphous silica, and both solubilities are temperature dependent. We thus can write:

$$T(^{0}C) = \frac{1309}{5.19 - \log_{10}C} - 273 \qquad 180^{0}C < T < 250^{0}C$$

(quartz geothermometer, no steam loss)

and

$$T(^{0}C) = \frac{731}{4.52 - \log_{10}C} - 273 \qquad 100^{0}C < T < 180^{0}C$$

(amorphous silica geothermometer)

where C = silica concentration in the geothermal fluid in mg/kg.

Other silica geothermometers are based upon equilibrium with chalcedony, α-cristobalite, or β-cristobalite, and it is obviously of importance to know which silica minerals exist in the reservoir rocks. If drill information is not available on this point, as it usually is not early in an exploration program, one must rely on the geologic mapping and inference to provide this information.

A second system of geothermometers is based upon the equilibrium reached among sodium (Na), potassium (K) and calcium (Ca) where reservoir rocks contain abundant quartz and feldspar (Fournier and Truesdell, 1973). One common geothermometer of this class is the sodium-potassium-calcium geothermometer:

$$T(^{0}C) = \frac{1647}{\log_{10}(Na/K) + \beta[\log_{10}(\sqrt{Ca}/Na) + 2.06] + 2.47} - 273$$

where $\beta = 4/3$ if < 100 and [1 cg ($\sqrt{Ca}$/Na) + 2.06] > 0, but if

T with $\beta = 4/3$ is > 100 or if [1 cg ($\sqrt{Ca}$/Na + 2.06] < 0, use $\beta = 4/3$.

Calculations are in mg/kg.

Different geothermometers frequently give different results when applied to the same fluid. Care must be taken in interpretation, and in this matter there is no good substitute for experience. Use of other data may help shed light on the relative reliability of the various geothermometers in specific geologic situations. For example, silica concentration can be affected by pH, and temperatures calculated from the sodium-potassium-calcium geothermometer may be in serious error if the CO_2 or magnesium concentrations are too high or if there has been addition of any of these elements through interaction of the fluid with sedimentary rocks or ion-exchanging minerals such as montmorillonite.

Mixing of the thermal reservoir waters with normal ground water can also change concentrations of the critical elements in a geothermometer, and can result in a calculated temperature that is either too high or is to low.

Isotope Geochemistry. There is strong evidence, based on stable isotope analysis, that the fluids which form hydrothermal convection systems are largely meteoric in origin (Craig, 1963). The evidence is based on an observed positive δO^{18} shift of geothermal fluids relative to meteoric fluids in the vicinity of the system (Figure 24). The relative abundances of the isotopes of oxygen is well known. The shift in geothermal waters is due to the interaction of the fluids with the heavier δO^{18} characteristic of host rocks. Since rocks contain negligible hydrogen, there is little change in deuterium, δD. Reviews of this topic may be found in Ellis and Mahon (1977). The δO^{18} and δD values of magmatic fluids are also shown in Figure 24, and it can be seen that for some systems the fluid may contain some component of magmatic water, but this is generally not thought to be large.

Isotopic studies can help answer questions on reservoir recharge system permeability. Isotopes have also been used to attempt to quantify the age of geothermal systems. The most successful has been tritium (^3H) which has a half life of 12.26 years. Minor amounts of tritium are produced by cosmic radiation in the stratosphere. However, major amounts have been put into the atmosphere by tests of thermonuclear weapons. Tritium concentration is expressed in terms of the Tritium Unit (T.U.) which is equivalent to T/H of 1 x 10^{-18}. In continental climates in the temperate zone cosmic radiation produces about 10 T.U. Up to 10,000 T.U. were measured in 1963 following extensive atmospheric testing of nuclear weapons. This decreased until about 1968, and since then as remained fairly constant. The following generalizations can be made concerning the age of water in the absence of mixing. A T.U. of less than 3 indicates that no water younger than 25 years is present. Values of 3 to 20 T.U. suggest that a small amount of thermonuclear tritium is present, which suggests that the fluids entered the groundwater environment in the 1954-1961 time frame. If greater than 20 T.U. are found, the water is younger than 1963.

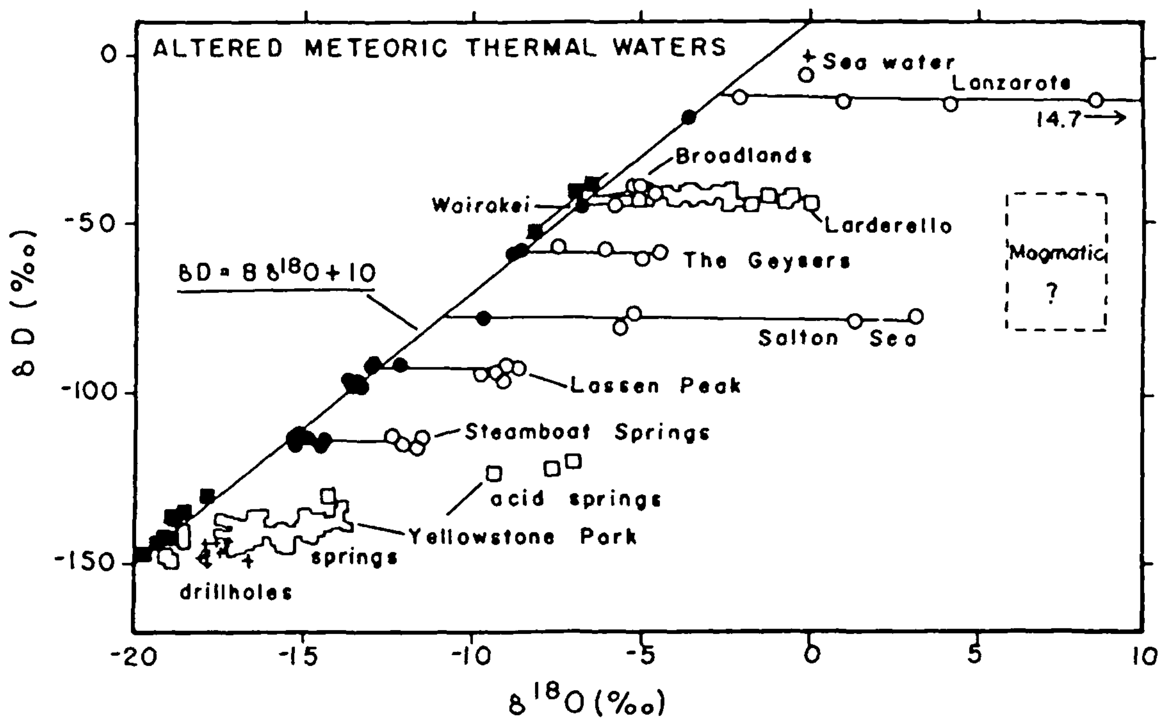


Figure 24. Oxygen-18 and Deuterium compositions of hot spring, fumarole, and drill hole thermal fluids derived from meteoric waters (o) and of meteoric waters local to each system (o).

Geophysics

Geophysical prospecting is the use of physical techniques and measurements either to detect directly the item sought or to provide indirect evidence of its existence and location. Such physical parameters as the distribution of temperature over the surface and at depth, the rate of heat flow to the surface, the electrical conductivity, magnetic susceptibility, density and elastic wave parameters of rocks can all respond in their own way to the presence of a geothermal resource (Ward, 1983; Wright et al., 1985) to an ore body, coal deposit or petroleum reservoir. We will discuss each of these methods briefly.

Thermal Methods. Under suitable circumstances, geothermal resources can be detected directly by application of one or more of the thermal methods. One basic parameter of interest is the heat flow, the rate at which heat flows upward toward the surface. We have seen previously that the outward flow of heat is a worldwide phenomenon (Sass et al., 1981), but in geothermal areas the heat flow is higher than this ubiquitous background amount, and so anomalous heat flow values may be clues to underlying geothermal resources.

The (vertical) heat flow is given as

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

where k = thermal conductivity (W/M-°C),

T = temperature (°C)

and z = the vertical coordinate in meters.

The quantity dT/dz is, of course, the vertical temperature gradient, the geothermal gradient, and in practice it is approximated by measuring temperature down a borehole and forming ratios $\Delta T/\Delta Z = (T_{z2} - T_{z1})/(z_2 - z_1)$ for various depth intervals. A typical value for the geothermal gradient is 30°C/km, or 0.03°C/m which is equivalent to 1.6°F/100 ft. 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. This need for samples of subsurface rocks exists in application of many geological, geochemical, geophysical and engineering techniques, and will be dealt with more fully below.

An often applied but dangerous shortcut to heat flow surveys is to forgo measurement of thermal conductivity, perhaps to save the cost of good sample collection and laboratory analyses, and to use only the thermal gradient data, $\Delta T/\Delta z$. Obviously, lateral as well as vertical variation of $\Delta T/\Delta z$ could be due to genuine changes in the heat flow field or simply to changes in lithology that are unrelated to any geothermal resource. And although it is often done, extrapolation of an observed temperature gradient to levels below the borehole is generally not justified and usually leads to disappointment.

Drilling can be expensive, and so the natural tendancy is to use thermal gradient or heat flow holes that are as shallow as possible. It is desirable to make the temperature measurements below the level affected by seasonal air temperature variations, and one is usually safe on this account with holes that are deeper than about 30 meters. Perhaps the biggest problem with shallow gradient holes, and deep holes in certain geologic environments, is movement of ground water. In some areas of sufficient topographic relief and abnormal precipitation, 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 pattern over the resource. It is imperative that one understands complications likely to be introduced into a heat flow or thermal gradient survey program before one embarks on use of this expensive technique.

Several workers have shown the utility in a few geothermal areas of the use of very shallow, say 2 meters deep, holes for heat flow studies or simply to measure anomalous near surface temperature. Such surveys at the Coso Hot Springs area in California show a +2C° anomaly over the reservoir in 2 m holes (LeShack and Lewis, 1983). Careful corrections must be applied for slope of the land, surface soil or rock type and vegetation (which affect reflectivity), surface hydrology, topography and other factors if these data are to be useful.

Existence of such shallow temperature anomalies implies that airborne or even satellite surveying in the thermal infrared region of the spectrum may be helpful. In practice, these methods have not been widely applied to date. Soil temperature fluctuations induced by sun angle variations, vegetation, ground slope and water table variations, to name a few variables, cause a high level of background "geologic" noise against which one must try to resolve the rare geothermal anomaly. Of course in specific areas, depending upon the geologic situation, infrared airborne surveying may be very helpful, but they would probably not constitute a first step in any exploraton program.

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 what they are associated can be detected.

With the exception of a few ore minerals, rock-forming minerals do not conduct electricity well. Electricity is conducted in the earth within waters that occupy the pore spaces in the rock. As we have seen, these waters invariably contain dissolved chemical species in the ionized state and the ions respond to an applied voltage difference between two points by moving through the water, thus sustaining a current. Two primary factors affect the conductivity of the groundwater: the concentraton of dissolved constituents and (2) the temperature. As we would expect, the higher the concentraton of dissolved species, the more ions in solution and the higher the rock conductivity (the lower the resistivity) will be. As temperature increases the activity of the ions increases, and so increasing temperature also causes increasing conductivity (decreasing resistivity). A third important cause of variation in rock conductivity is the amount, location and type of certain minerals that are capable of adsorbing ions from the solution or of ion exchange with the solution. If pore spaces or fractures are lined with minerals such as clays or zeolites, they may be holding loosely bound ions that are mobile enough to move along and within the layers of the clay structure in response to an electric field. Presence of clays and/or zeolites can greatly increase electrical conductivity (Moskowitz and Norton, 1977).

On the basis of the foregoing and from what we already know about the temperature, salinity and hydrothermal alteration within hydrothermal systems, one would expect such systems to be good electrical conductors. Indeed, low resistivity (high conductivity) has been discovered by surface surveys over many geothermal systems, and geophysical techniques that measure resistivity are in use worldwide in geothermal exploration (Ward and Sill, 1983). Most electrical techniques can be classified into one of those discussed below.

Galvanic Resistivity Surveys. In this technique two grounded electrodes are used to introduce a current flow in the earth, and voltage is measured between two separate grounded electrodes. There are several ways to deploy the electrodes, but perhaps the most useful configuration is the dipole-dipole array. Using this technique an effective depth of exploration of approximately two times the electrode separation can be achieved, and because the maximum practical value for separation is perhaps 500-750 meters, the resistivity method can detect low resistivity zones to depths of 7500 to 1500 meters. Volcanic areas often have high electrode contact resistance, causing low transmitted current and precluding deep exploration. Computer-aided interpretation methods are available and are easily applied. The method can be very useful for obtaining detail on a geothermal system.

Magnetotelluric (MT) Surveys. In this method, natural magnetic and electrical signals are used (Ward and Wannamaker, 1983). It can be shown that a measure of resistivity is given by the ratio of the electric field to the perpendicular magnetic field. Now, an electromagnetic field will penetrate into the electrically conducting earth to a depth dependant on its frequency. One can define a "skin depth" as that depth at which the electromagnetic field is attenuated in strength by the factor 1/e from its value at the surface, where e is the base of the natural logarithms, and equals approximately 2.72. Thus, lower frequency waves penetrate to deeper depths than do higher frequency waves, and by making simultaneous measurements of Ex and Hy for a range of frequencies, a depth sounding may be effected, the lower frequencies yielding information from deeper depths. Magnetotelluric surveys have been used with some success over a number of geothermal systems.

Magnetotelluric instrumentation incorporates the capability to make a tensor measurement, that is, to measure simultaneously both orthogonal electric field components (E_x, E_y) and all three orthogonal magnetic field components (H_x, H_y, H_z). This method is generally considered to be capable of exploration to depths of tens of kilometers, and to be capable of detecting magmas directly. Neither of these attributes is true in all cases. Although magma is conductive due to mineral semiconduction, the amount of contained water substantially affects the conductivity, dry magmas being much less conductive than wet ones (Lebedev and Khitarov, 1964; Wannamaker, in press). In geothermal exploration, it is possibly the wet magmas that we seek, however, because they have enough volatile content to produce the fracturing needed for hydrothermal convection. Depth of exploration depends to a certain extent on the near-surface resistivity structure. Also of great importance is the size and other characteristics of the magma body. Newman et al. (1985) have explored conditions under which crustal magma bodies can be detected. They conclude that if the body is isolated, i.e. has broken off from conductive magma at depth, it is more easily detected than if it maintains connective roots to the mantle.

The MT method has been used a great deal in geothermal exploration with generally disappointing results (Ward, 1983). By far the biggest problems appear to be misapplication and inadequate interpretation. Most MT data have been interpreted using one-dimensional inversion to a layered-earth resistivity structure. This method is totally inadequate in most geothermal exploration and usually produces misleading results. Full three-dimensional modeling is needed. The MT method has many subtleties, and must be applied with a great deal of care by geophysicists who are well experienced.

MT equipment can be considerably simplified if its range of operation is restricted to frequencies between about 1/10 Hz to 10,000 Hz, loosely called the audio range. This covers the depth range of usual interest in geothermal exploration. Therefore the AMT, or audiomagnetotelluric method, has seen some geothermal exploration. Most reported AMT surveys are scalar AMT, that is, only one component of electric field and one of magnetic field are measured at once. It can be demonstrated that in layered terrains this scheme is adequate for obtaining resistivity structure, but if resistivity also varies in either or both horizontal dimensions, as it invariably will in volcanic areas, scalar AMT is inadequate and is not recommended for exploration. For this task, tensor AMT measurement is needed. AMT has the advantage over conventional MT that it is less expensive both for surveying and interpretation.

CSEM. Controlled-source electromagnetic methods have been used as alternatives to galvanic resistivity or AMT surveying (Keller and Rapolla, 1974; Keller et al., 1982). A high-powered CSEM system has been developed and reported by workers at Lawrence Berkeley Laboratory (Wilt et al., 1981). The primary limitation of these techniques to date has been that interpretation methods have been limited to the one-dimensional case. Two- and three-dimensional algorithms are now becoming available, but further development is needed.

SP. Spontaneous-potential anomalies over convective hydrothermal systems arise from the electrokinetic and thermoelectric effects, which couple the generation of natural voltages with the flow of fluids and the flow of heat, respectively (Corwin and Hoover, 1979; Sill, 1983). SP surveys have been used successfully in certain volcanic terrains. On Hawaii, Zablocki (1976) found a large SP effect over the East Rift zone. Although these surveys are relatively inexpensive to run, they are also difficult to interpret in terms of the nature and location of the source area.

Seismic Methods

Elastic waves are transmitted through rocks and their measurement can be used to help determine the structure and properties of rock bodies. Two types of waves are most useful: (a) the compressional or primary (P) wave in which the particle motion is back and forth along the direction of travel of the wave, and (b) the shear secondary (S) wave in which the partial motion is perpendicular to the direction of travel of the noise. P-waves are ordinary sound-waves in rocks, and travel with typical velocities of 3-6 km/sec. S-waves have no analog in the air because fluids (liquids and gasses) do not support shear. In rock, S-waves travel at velocities about 70% of those of the P-wave.

Active Seismic Surveys. In this application, seismic waves are introduced into the earth, generally by explosion of a charge of dynamite or other explosive in a shallow borehole. Returns of seismic waves are measured at the surface. Coherent groups of waves are produced when a downgoing seismic wave reflects from a contact between two bodies of rock having different seismic velocities and/or densities and such reflections can be recognized in the surface seismic seconds and the depth to and attitude in space of the reflectors can be determined. In this way, much can be added to the knowledge of subsurface geology (Applegate et al., 1981). However, where the structure becomes complicated, diffraction of seismic waves occurs and makes the task of interpreting structure difficult. At Beowawe, Nevada, extensive and varied digital processing was ineffective in eliminating the ringing due to a complex near-surface intercalated volcanic-sediment section (Swift, 1979). This problem is typical in volcanic areas. 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 and for obtaining other structural-stratigraphic information. Because of access problems with large equipment, interpretational difficulties and the expense of reflection surveys, we see only limited use of this method in most volcanic exploration situations.

Passive Seismic Techniques. Seismic waves also occur naturally in the earth and such natural waves can be detected at the surface. There is limited evidence (e.g. Liaw and Suyenaga, 1982) that hydrothermal processes can generate seismic body waves in the frequency band 1 to 10 Hz. Noise also arises in such sources as traffic, trains, rivers, canals, wind, etc. Liaw and McEvilly (1978) have demonstrated that field and interpretive techniques for earth noise surveys require a great deal of understanding and care. These surveys can provide a guide to hydrothermal processes provided tha data quality is good and careful interpretation is done.

Microearthquakes frequently are closely related spatially to major geothermal systems. Accurate locations of these earthquakes can provide data on the locations of active faults that may channel hot water toward the surface. Microseismic activity is generally episodic rather than continuous, and this characteristic by provide a basic limitation to the technique in searching for or prioritizing geothermal prospect areas.

If a sufficiently distant earthquake is observed with a closely spaced array of seismographs, changes in P-wave traveltime from station to station can be taken to be due to velocity variations near the array. Traveltime residuals are computed as the observed arrival time minus that calculated for a standard earth. A magma chamber beneath a geothermal system would give rise to low P-wave velocities and hence to late observed traveltimes (Iyer and Stewart, 1977). While one can speculate that relative P-wave delays are caused by partial melts or magmas, they can also be caused by alluvium, alteration, compositional differences, lateral variations in temperature or locally fractured rock. Iyer et al. (1979) shows evidence for detection of magma in The Geysers using this technique.

Magnetic Methods

The earth has a main magnetic field that is similar in geometry to that of the classical bar magnetic and which is believed to arise from electrical currents flowing deep within the earth, in the electrically conducting, fluid core. This main field induces a magnetic response in certain magnetic minerals at and near the earth's surface and by detecting spacial variations in the total field the variations in distribution of magnetic minerals may be deduced and, through the process of interpretation, related to geology. The earth's field can be mapped on the ground, but it is much faster and cheaper to map this field from the air.

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 (Wright, 1981). Magnetic susceptibility often varies from 0 to 7000 μcgs units in these rock types and provides major magnetization changes which delineate geologic units. The scale of many geothermal systems is also similar to porphyry-type mineral occurrences.

Regional aeromagnetic data are often available as part of state or nationally sponsored surveys. These data often show major structural features and aid in forming a generalized geologic model for otherwise covered geology prospect areas. These regional data are generally too widely spaced and/or too high in altitude however, to warrant detailed quantitative model interpretation.

The locations of geologic structures (faults, fracture zones), intrusives, silicic domes and possibly major alteration areas (speculative) are often apparent on aeromagnetic data. Kane et al. (1976) show how gravity and magnetic data can be used to interpret structure at Long Valley, California. One other application of aeromagnetic surveying is in determining the depth to the so called Curie point isotherm. The Curie point of a rock is the temperature above which the rock ceases to be magnetic. Because of the increase in temperature with depth in the earth, the geothermal gradient, rocks below a certain depth will be non-magnetic because they will be above their Curie point, which for the mineral magnetite is about 585°C. In areas where the geothermal gradient is high, higher temperatures exist at shallower depths than elsewhere, and the Curie point isotherm is shallower. Through suitable interpretation techniques, usually computer based, the depth to the Curie point can be determined for an area from the aeromagnetic data from that area (Bhattacharyya and Leu, 1975). This gives one an idea of regions where the gradient is high, which would presumably be regions of greater geothermal potential. However, numerous limitations and assumptions apply to application of this interpretation technique, and such work must be under the direction of someone well experienced in it.

The Gravity Method

Minute variations in the earth's gravity field are caused by variations in density of subsurface rocks. In order to detect these variations very delicate instruments are required. The modern gravity meter measures 1 part in 10^9 of the earth's gravity field, and are among the most sensitive mechanical instruments yet made by man.

Gravity data are often acquired or compiled in the early stages of an exploration program. Regional data, with station densities of 1 station per sq km to 1 station per 25 sq km, may be available as the result of surveys by governments or universities. These data are generally 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, lithologic and other information. Isherwood (1976) interpreted the gravity survey of The Geysers area in terms of location of a subsurface magma chamber. In the Imperial Valley, California, gravity surveys have proved useful in locating areas where hydrothermal alteration and metamorphism have caused the sediments to become densified (Rex et al., 1971). Grannell (1980) showed how gravity could be used to monitor changes in the reservoir at Cerro Prieto due to production.

DRILLING AND LOGGING

Introduction

Geothermal drilling is generally done with somewhat modified, conventional drilling equipment. For wells that are expected to produce large quantities of hot water and/or steam for electric power production, large rotary drilling rigs of the type used for oil and gas wells are used, but if smaller quantities of lower temperature fluids for use in space heating are sought, a conventional water well rig might suffice. Because drilling is one of the most expensive steps in geothermal development and at the same time is typically fraught with problems and setbacks, it is important to choose the correct drilling contractor, equipment and techniques for the job at hand and not to compromise the drilling program for the sake of attempting to save money. One will usually come out ahead cost-wise by going at this job right in the first place rather than skimping and then running into expensive and time consuming difficulty. Even drilling contractors well experienced in other types of drilling will have trouble at first with geothermal drilling because there is not a direct transfer of oil field, water well or mining technology. The geothermal environment is characterized by temperatures above those in which most drilling equipment and muds were designed to operate. In addition, geothermal brines are especially corrosive due not only to their chemical composition but also to their elevated temperature. Special equipment and procedures are necessary to ensure the safety of the drilling crew and to minimize the possibility of blow-out. Besides all of these extra considerations, a good geothermal environment will have had a history of complex geological and geochemical processes that vastly increase the chances for difficulty during drilling attributable to zones of lost circulation, high angle faults or fractures that deflect the hole, or caving or sloughing of wall rock into the borehole.

During the drilling operation, the geologist, geochemist and geophysicist will be collecting exploration data important to their jobs. The geologist will log (record) information derived from examination of the drill chips or core. The geochemist will attempt to obtain uncontaminated samples of the subsurface fluids for analysis and the geophysicist will be concerned with geophysical well logging of the hole. We will consider these topics separately in what follows.

Lithologic and Mud Logging

Rotary drilling produces cuttings from the rock at the bottom of the hole which are brought to the surface by the drilling mud. These cuttings are collected at the shaker table on the rig, which separates them from the mud. It is the job of the geologist to specify sampling interval, sample size, collection method, washing procedures and packaging and labeling specifications. Grab samples from the shaker table at 3-meter intervals are usually recommended, and a minimum of 500 gm should be collected. When there is a great deal of lost circulation material in the mud, sample size should be increased to maintain the specified amount of rock chips. If the rock is hard, sample collection at the shaker table is no problem, but soft sands and especially clays may be lost through mixing with the mud. If high clay content is suspected, samples of the mud can be taken before the shaker table and analyzed. If the mud is oil-based, the cuttings must be washed immediately, otherwise washing is very difficult later. For water-based muds the geologist has options. Dried, unwashed samples are difficult to disarticulate, which argues for immediate washing. However, washing under laboratory conditions may produce better results.

As washed samples become available, the geologist logs these samples by examining them under a binocular microscope (Hulen and Sibbett, 1982). He identifies the rock type (lithology), notes evidence of hydrothermal alteration, looks for evidence of faults (development of fault gouge and slickensides), notes occurrence of vein minerals such as calcite or quartz and identifies material caved from the hole above the bottom. He records his observations on a standard log form.

If the drilling produces core, much more geologic information can be generated than from chips. The geologist estimates percentage core recovery by measuring the actual core and comparing to the footage drilled. The core is washed and a geologic log made by recording rock type, planar features, alteration, mineralization and other important features on a standard log form. The core may be photographed box by box. Some studies, such as core fluid analysis, require special handling and preservation. To preserve original fluids requires that the core be sealed in polyethylene tubes.

Fluid Sampling

It may be difficult and costly to obtain an uncontaminated sample of the subsurface fluids. Drilling muds are designed to hold back reservoir fluid pressure, and so there is usually a migration of mud filtrate into the rocks adjacent to the drill hole. In order to obtain an uncontaminated sample, enough fluid must be removed from the hole to eliminate contamination effects. There are several ways to do this, all of which require appropriate equipment and permits to handle large volumes of fluid at the surface. In the so-called "drill-stem test", the aquifer to be tested in packed off above and below and then flowed through the drill pipe until the chemistry of the produced fluid becomes stable. This is taken as an indication that uncontaminated reservoir fluid is being sampled. An alternative is to flow the entire uncased portion of the well, which may be cheaper because packers do not have to be set, but the procedure produces a mixed sample from several aquifers, and the interpretations that can be made from chemical analysis becomes more ambiguous and misleading. A third method of sampling is to case the entire hole, then perforate the casing at selected locations and obtain samples by flowing or bailing the hole.

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. Probes are lowered into the borehole to make these measurements. 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 understand fully the responses of various well logs in geothermal reservoirs and their typically fractured, altered, commonly igneous and metamorphic host rocks (Sanyal et al., 1980). In spite of the relative lack of knowledge of well-log reponse 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 a well (Hearst and Nelson, 1985).

A second, important problem in geothermal well logging is general lack of probes that will work in an environment where temperature exceed 200-250°C. Electric components and cables have generally not been available for temperatures this high and indeed have not generally been required by the petroleum industry, which makes the most use by far of well logging. Sandia National Laboratories has had an active and successful research program sponsored by the U.S. Department of Energy, to develop electronic components and logging tools for use in the geothermal environment, and so appropriate logging equipment is now becoming available.

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 may represent the only indication of permeable zones since production and injection tests can not, of course, be performed for cased intervals without perforation of the casing. 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 cross-calibration of logs that may be made with different instruments and different calibrations on successive logging runs. Few developers or drilling contrctors offer logging services themselves. Geophysycal logging of the well is almost alway done by a separate contractor.

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 herewith.

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

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

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 of detailed logging to detect zones of fluid flow or perform heat flow studies. One generally needs a calibrated log with a sensitivity of $\pm$ 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. 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 induction log also measures the electrical resistivity or conductivity of surrounding rocks, but requires no electrode contact with the borehole wall, as the conventional resistivity logs do. Magnetic fields generated by coils in the probe are used to induce currents to flow in the rock and the response of the material surrounding the borehole is measured by other coils in the probe.

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 K^{40} to Ar^{40}. 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.

The gamma-gamma 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. Owing to lack of calibration, gamma density logging may not be as useful in igneous and metamorphic rocks as in sedimentary rocks. Densities of certain igneous and metamorphic rocks, for example, may exceed the calibration range of commercially available logging tools. Additionally, gamma-gamma 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. Hydrogen nuclei in the formation 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 and rock quality designation or intensity of fracturing. Fractures can be located by analyzing the full wave form of the incoming acoustic velocity signal.

The acoustic televiewer, also known as the borehole televiewer or seisviewer, provides, through complex instrumentation described by Hearst (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 geothermal system (Keys and Sullivan, 1979).

Well Logging Example. We have selected only one of numerous examples to illustrate the application of well logging to geothermal exploration. Murawato and Elders (1984) discuss well logs in the Salton Sea and Westmorland geothermal fields iN California. Figure 25 shows gamma ray, gamma-gamma density, spontaneous potential and resistivity logs for a hole that intersects sediments typical of the Salton Trough and basalt dikes. Note the expression of the dikes as opposed to the sediments on these data.

Cross plots of one type of borehole data vs. another can greatly facilitate data interpretation, particularly for boreholes in complex igneous and metamorphic terrain (Ritch, 1975; Glenn and Hulen, 1979). As an example of the utility of these plots, bulk density is plotted against neutron porosity in Figure 26 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, 1979). 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.

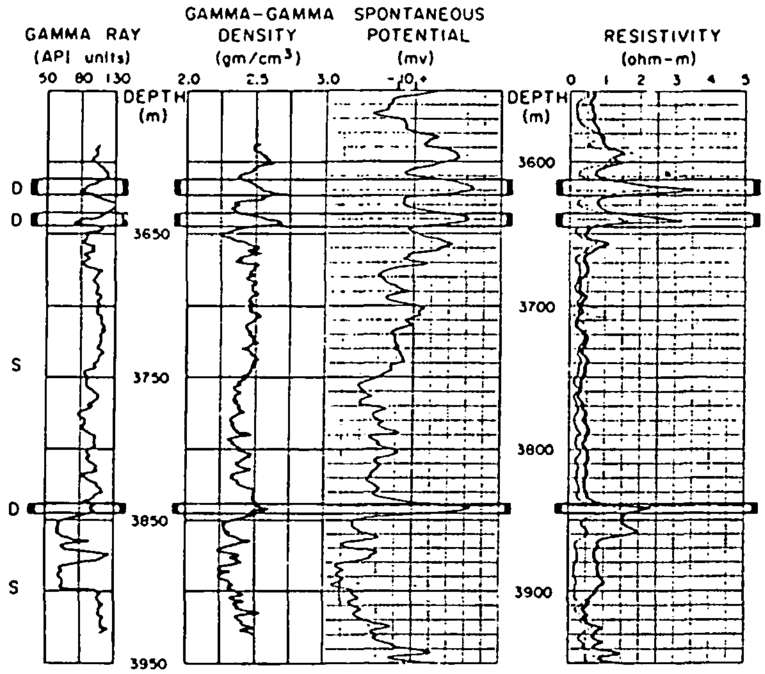


Figure 25. Geophysical well log from Magmamax 3, Imperial Valley, CA (from Muramoto and Elders, 1984).

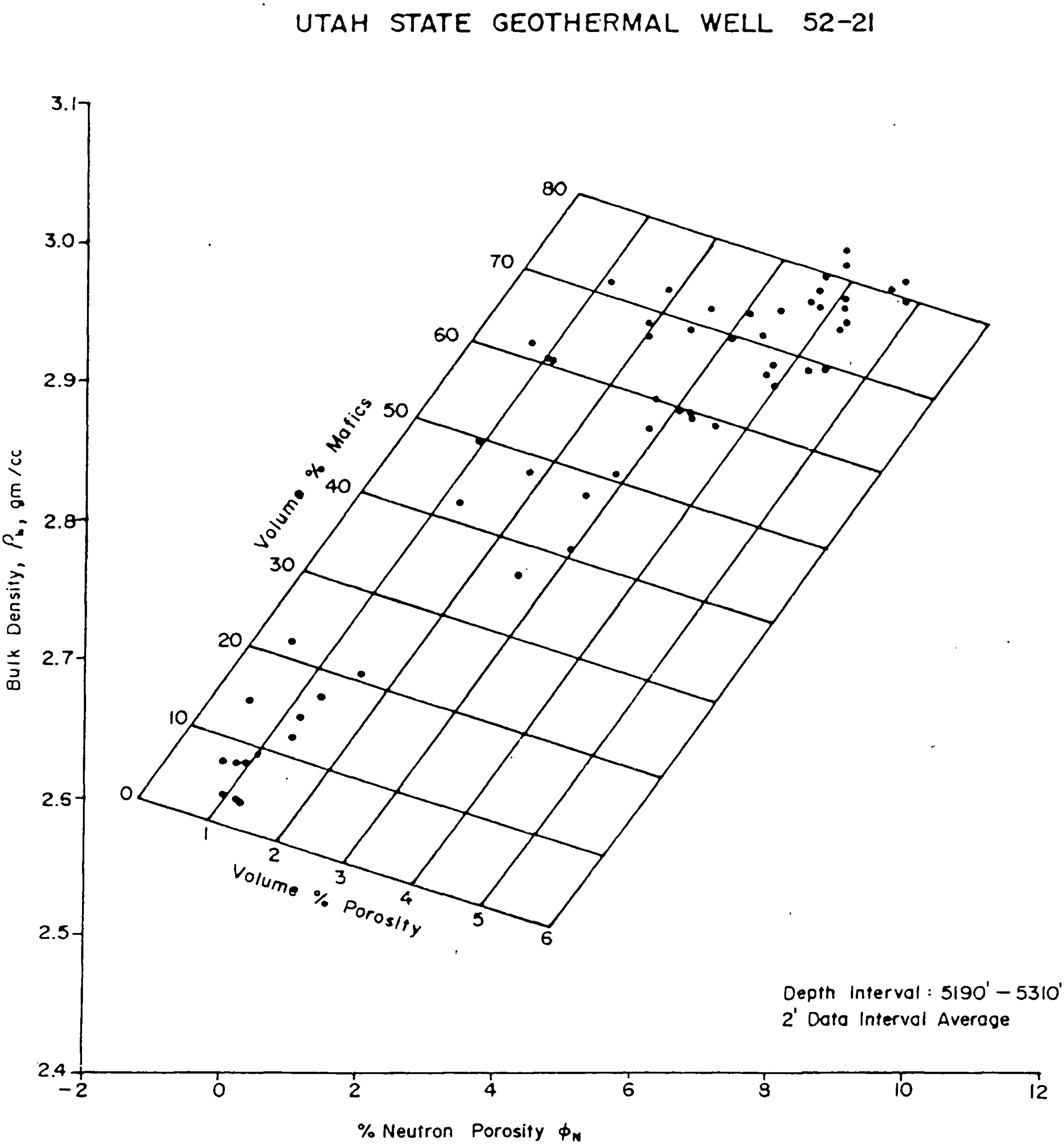

UTAH STATE GEOTHERMAL WELL 52-21
Bulk Density, ρ_b, gm/cc
% Neutron Porosity ϕ_N
Volume % Mafics
Volume % Porosity
Depth Interval : 5190' — 5310'
2' Data Interval Average
3.1
3.0
2.9
2.8
2.7
2.6
2.5
2.4
−2
0
2
4
6
8
10
12

Figure 26

EXPLORATION STRATEGIES

Geothermal development is an interdisciplinary endeavor. Figure 27 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, there are simply no tools or techniques to solve a particular problem.

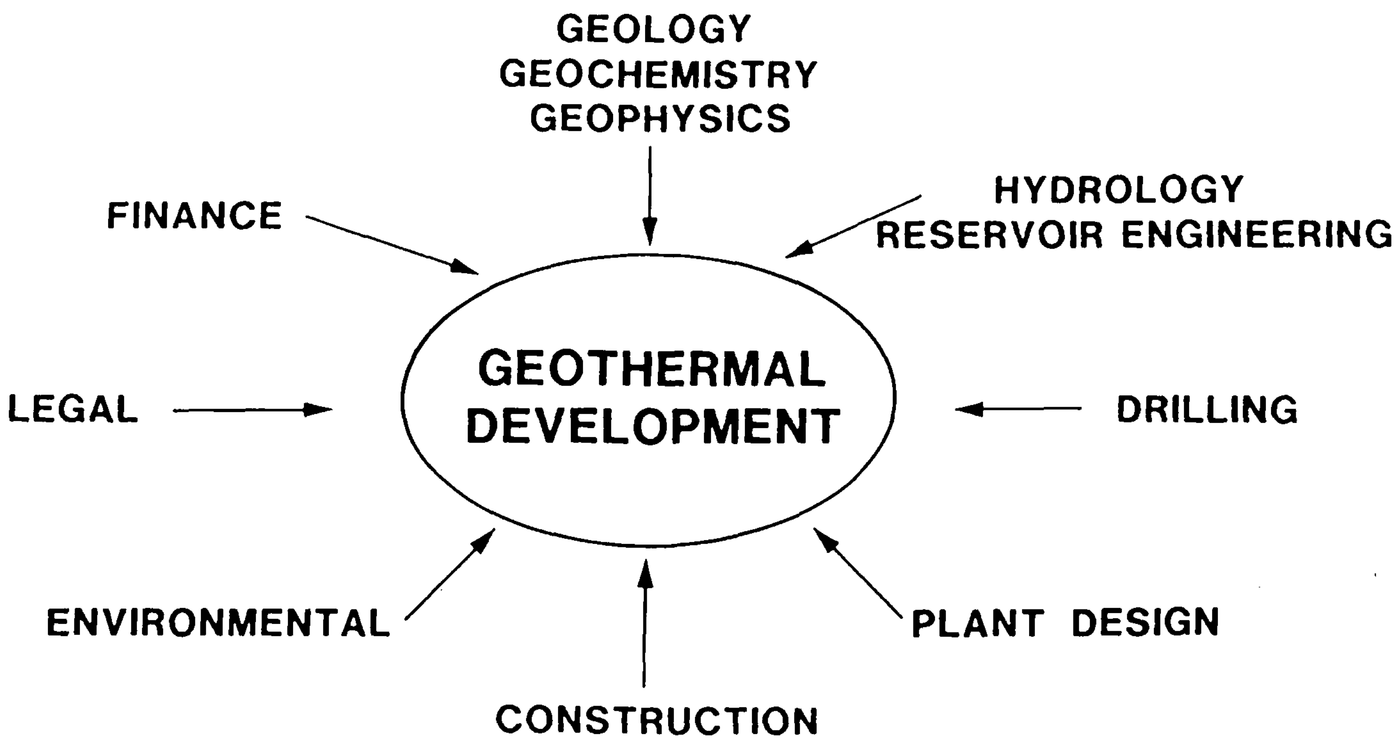
Geothermal Exploration

As previously mentioned, the geosciences have two primary applications in geothermal development: 1) exploration *for* geothermal systems, and 2) exploration *within* geothermal systems.

Figure 28 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.

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

GG-105

Figure 27

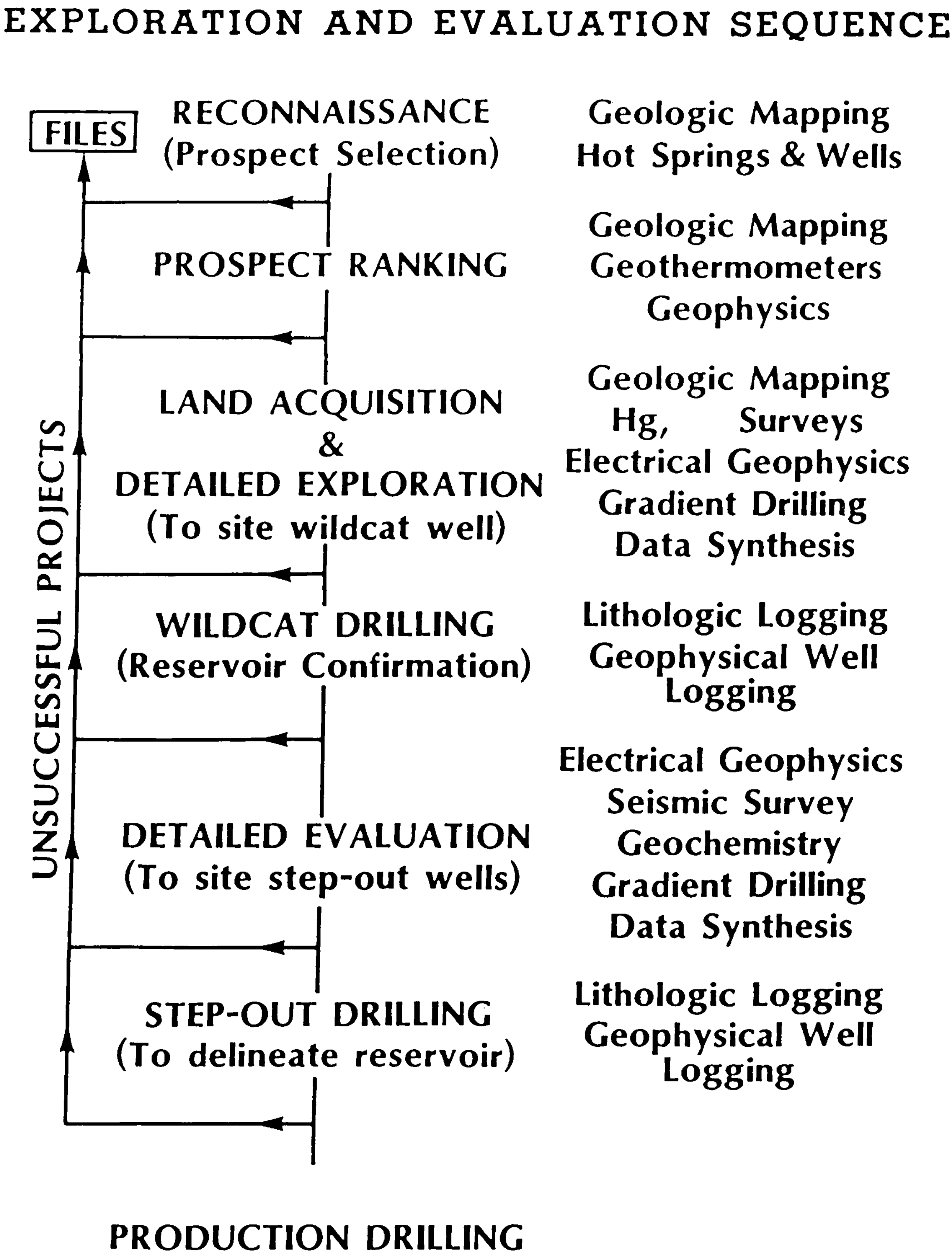

EXPLORATION AND EVALUATION SEQUENCE
FILES
RECONNAISSANCE
(Prospect Selection)
Geologic Mapping
Hot Springs & Wells
PROSPECT RANKING
Geologic Mapping
Geothermometers
Geophysics
LAND ACQUISITION
&
DETAILED EXPLORATION
(To site wildcat well)
Geologic Mapping
Hg, Surveys
Electrical Geophysics
Gradient Drilling
Data Synthesis
WILDCAT DRILLING
(Reservoir Confirmation)
Lithologic Logging
Geophysical Well
Logging
DETAILED EVALUATION
(To site step-out wells)
Electrical Geophysics
Seismic Survey
Geochemistry
Gradient Drilling
Data Synthesis
STEP-OUT DRILLING
(To delineate reservoir)
Lithologic Logging
Geophysical Well
Logging
PRODUCTION DRILLING
UNSUCCESSFUL PROJECTS

Figure 28

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 area must be designed specifically for application to that area by the geoscientists who are performing the work and interpreting the data.**

Exploration Strategy. Figure 29 is a diagram of a basic generic exploration strategy. Before such a strategy can become truly useful, much more detail must be added to each of the steps. Several aspects of Figure 29 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 assessing odds for success at each decision point, and then comparing the project to others or other uses of the money and manpower, optimum exploration will result and the risks and costs of exploration will be 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

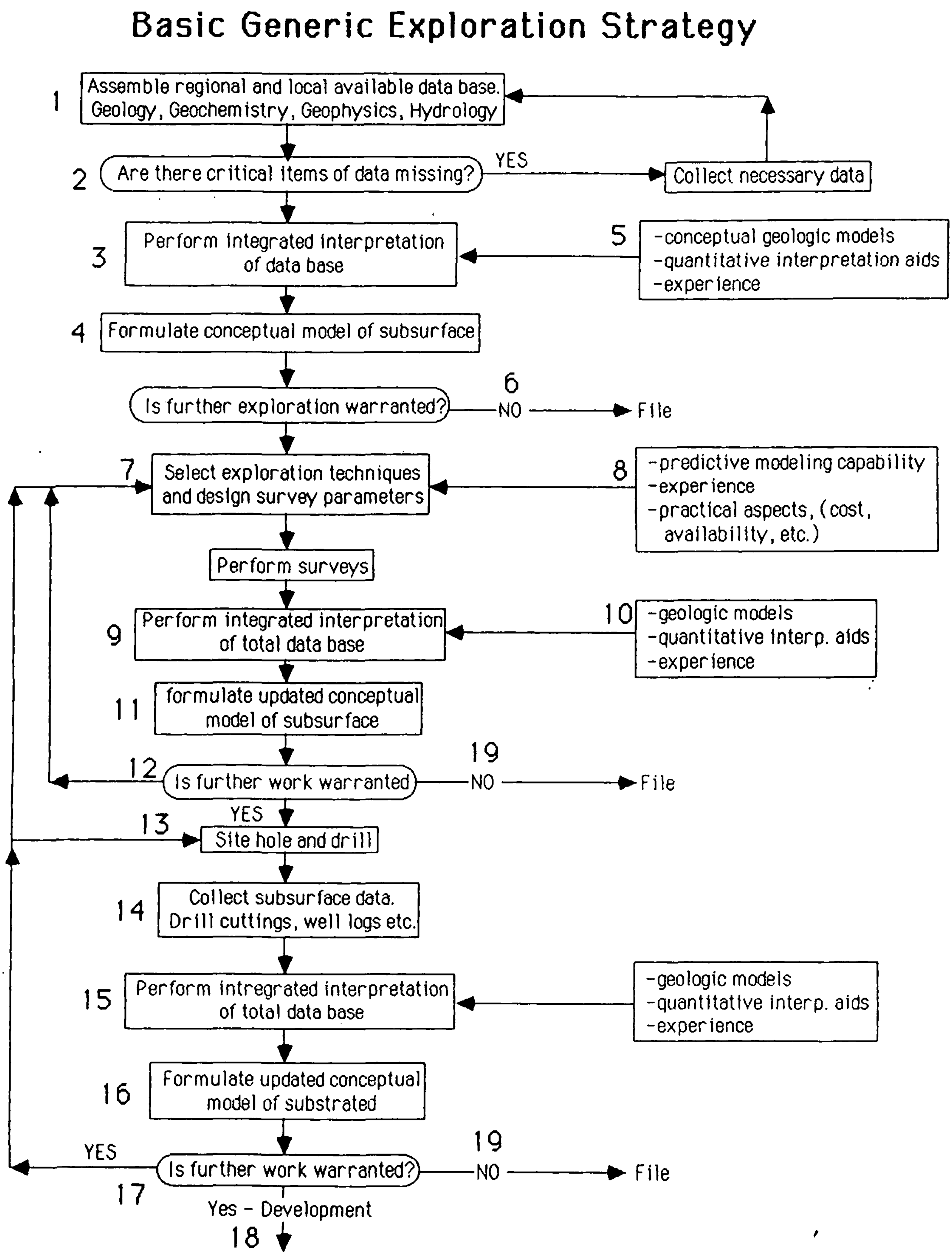

Basic Generic Exploration Strategy
1
Assemble regional and local available data base. Geology, Geochemistry, Geophysics, Hydrology
2
Are there critical items of data missing?
YES
Collect necessary data
3
Perform integrated interpretation of data base
5
-conceptual geologic models
-quantitative interpretation aids
-experience
4
Formulate conceptual model of subsurface
6
Is further exploration warranted?
NO
File
7
Select exploration techniques and design survey parameters
8
-predictive modeling capability
-experience
-practical aspects, (cost, availability, etc.)
Perform surveys
9
Perform integrated interpretation of total data base
10
-geologic models
-quantitative interp. aids
-experience
11
formulate updated conceptual model of subsurface
12
Is further work warranted
19
NO
File
YES
13
Site hole and drill
14
Collect subsurface data. Drill cuttings, well logs etc.
15
Perform intregrated interpretation of total data base
-geologic models
-quantitative interp. aids
-experience
16
Formulate updated conceptual model of substrated
YES
17
Is further work warranted?
19
NO
File
Yes - Development
18

Figure 29

that money and human resources are deployed in the optimum way.

Available Data Base (1). All available regional and local geological, geochemical, geophysical and hydrological data should be assembled for the prospective exploration area and its surroundings. Once assembled, specialists in each of the earth science disciplines should assess the data in a preliminary fashion to determine its quality and to identify any obvious gaps (2). Often basic geologic data will be missing, 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 upon it and must be in agreement with it.

Integrated Interpretation (3). When the data base is judged to be sufficient, it should be interpreted by specialists. By "integrated interpretation" we mean to convey the necessity for the various specialists to work closely together in the data interpretation process. The objective of this integrated interpretation is to formulate a conceptual geologic model of the subsurface (4) in the exploration area that agrees with all of the available data. Of course, the model should concentrate on those features that are pertinent to the potential for occurrence of a geothermal resource in the area.

In order to perform this interpretation step, a number of ingredients must be available (5). These include (a) knowledge of geologic models of geothermal resources in other areas as a basis 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 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.

Exploration Techniques and Survey Design (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. 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. These will be used to help define lithology, petrography and hydrothermal alteration and for measurement of physical properties. Conventional geophysical well logs should be run in the hole, with a minimum logging suite probably being temperature, caliper, resistivity, gamma ray and acoustic logs. If the well is flowed or if there is a drill-stem formation test, samples of the fluids from the well should be carefully collected and preserved for analysis. Often a hydrothermal component of such fluid samples can be detected through chemical analyses, lending encouragement for further exploration. Chemical geothermometer calculations can be made from the analyses 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).

Basin and Range Exploration Strategy

Ward et al. (1981) reviewed a great deal of exploration data from the Basin and Range province and suggested the exploration strategy shown in Figure 30 for this area. I use it here as an example of the thought process that should be used to design strategies for other areas.

SUGGESTED HIGH TEMPERATURE HYDROTHERMAL EXPLORATION STRATEGY

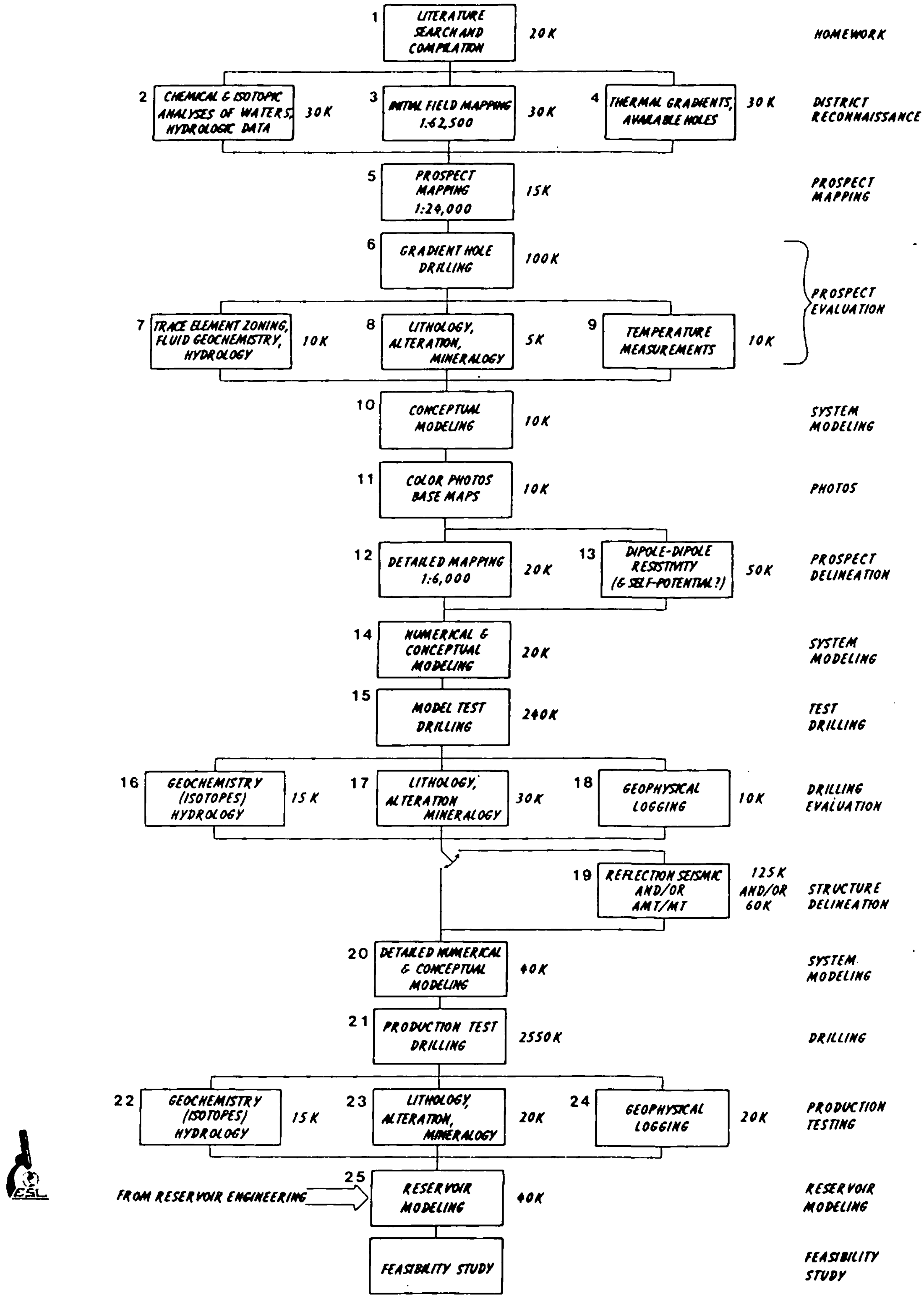


Figure 30

GG-031a

EXPLORATION EXAMPLE -- COSO, CALIFORNIA

It seems appropriate in this overview to illustrate a few selected exploration data sets in the geothermal environment. Because of space limitations relative to the very large amount of data available, we have chosen just one area for which the geology is well known and where drilling has established the presence of a significant high-temperature convection system. The Coso geothermal system, Inyo County, southeastern California (Figure 16) provides an instructive example where both regional and detailed geophysical data contribute to an understanding of the geothermal resource.

Geologic Setting

The Coso geothermal area is located in the Coso Range of the western Basin and Range province, immediately east of the southern Sierra Nevada. Regional geologic mapping of the area was completed by Duffield et al. (1980), who expanded the results of several earlier workers. Northerly-trending fault-block mountains are formed of diverse lithologies which vary in age from Precambrian through Holocene. The oldest rocks are complexly folded Precambrian through Early Mesozoic marine sedimentary and volcanic rocks, many of which are regionally metamorphosed (Hulen, 1978). This older sequence is intruded by Jurassic-Late Cretaceous granitic stocks and plugs which appear to be portions of the southern Sierra Nevada batholith. Late Cenozoic volcanic rocks were erupted in two periods, 4.0-2.5 my and $\leq$ 1.1 my (Duffield et al., 1980), and formed domes, flows and pyroclastic deposits which covered much of the crystalline rocks in the Coso geothermal area. Hulen (1978) completed detailed geologic mapping and alteration studies of approximately 40 km^2 of the immediate Coso geothermal area in support of the U. S. Department of Energy drilling program at well CGEH-1. A generalization of his map, Figure 31. provides a useful reference base for our evaluation. Hulen (1978) and Duffield et al. (1980) describe hydrothermal alteration and active thermal phenomena (fumaroles, steaming boreholes, and "warm ground") which occur throughout an irregular 20 km^2 area along the eastern margin of the Coso rhyolite dome field. Drill hole CGEH-1 was drilled to a depth of 1 470 m in 1977 primarily in a mafic metamorphic sequence and a leucogranite which intruded the metamorphic rocks. This hole indicated temperatures in excess of 177°C and convective heat flow which appeared to be limited to an open

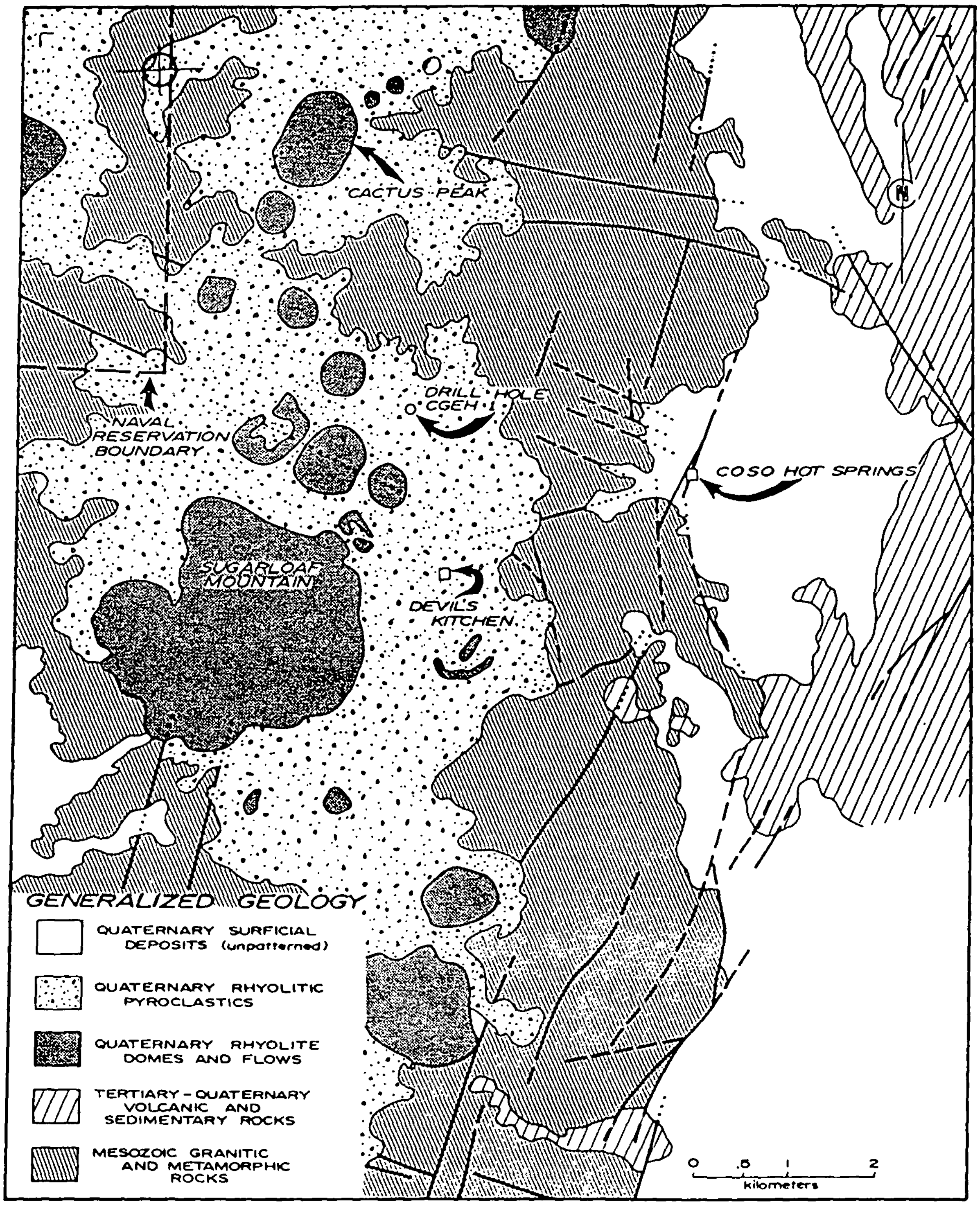


Figure 31. Geology of the Coso Geothermal Area, California.

fracture system between depths of 564 m and 846 m (Galbraith, 1978). Subsequently, several successful drill holes completed by California Energy Corporation have established the presence of a hydrothermal system.

Geophysical Studies

The Coso geothermal area is well expressed in quantitative thermal data. Combs (1980) completed a comprehensive study of the heat flow as determined in 24 shallow (35-110 m) and 2 deeper boreholes in an area of approximately 240 km^2 centered about the rhyolite dome field. He measured thermal gradients ranging from 25.3°C/km to 906°C/km, which he attributed to convecting hot water and former convective transport of heat by dikes that fed the domes and flows. Terrain-corrected heat-flow values ranged from 67 to 960 mW/M^2. The heat-flow anomaly is principally confined to the east-central portion of the rhyolite dome field and trends northeast to include Coso Hot Springs. LeSchack and Lewis (1983) describe shallow temperature surveys completed at Coso. The shallow (2 m) temperature measurements were made with a thermistor probe backfilled in a 2-m deep augered hole, after the thermistor equilibrated with surrounding earth temperatures. Temperatures of approximately 27.4 to 31.7°C form an anomaly pattern quite similar to the 400 mW/m^2 HFU contour of Combs (1980).

The University of Utah Research Institute completed a detailed dipole-dipole resistivity survey in September, 1977 as part of the U. S. Department of Energy resource assessment program which included the drilling of CGEH-1 (Fox, 1978a). A grid of three north-south lines and six east-west lines was surveyed to map the resistivity structure of a 41 km^2 area. An electrode spacing of 300 m was used for 41 line-km of survey, and a 150 m spacing for an additional 13 line-km. A $\leq$ 15 Ω·m low-resistivity zone was observed in the survey in a background resistivity of ± 200 Ω·m, and the resistivity low includes Coso Hot Springs, Devil's Kitchen and much of the surrounding area.

A detailed low-altitude aeromagnetic survey of 927 line-km was completed over the Coso area by the University of Utah Research Institute for the U. S. Department of Energy in September 1977 (Fox, 1978b). The data were recorded on north-south flight lines with a 400 m line spacing at a mean terrain clearance of approximately 230 m. Basement lithologic and structural information are apparent in the magnetic data in the form of magnetic discontinuities which correspond in part to mapped faults and structural trends. Most significant is a broad magnetic low which covers about 26 km^2 in the southeast intersection of the two major trends. Rock magnetization measurements, geologic mapping and alteration studies indicate that the magnetic low is due in part to magnetite destruction resulting from hydrothermal alteration by the geothermal system, as well as to primary lithologic changes at depth.

The region that includes the Coso Range and the southern Sierra Nevada is one of the more active seismic areas in southern California as summarized by Walter and Weaver (1980), who established a 16-station seismographic network over an area approximately 40 km north-south by 30 km east-west in the Coso range as part of the U. S. Geological Survey studies to evaluate the geothermal resource potential. They recorded 4 216 local earthquakes ($0.5 < m < 3.9$) during the first 2 years of operation. Many of these events occured in a 520 km^2 area which included Coso Hot Springs (CHS), Devil's Kitchen and the rhyolite domes. In addition, Young and Ward (1980) presented a three-dimensional attenuation model for the Coso Hot Springs area as determined from teleseismic data. They determined that a shallow zone of high attenuation exists with the upper 5 km in the Coso Hot Springs-Devil's Kitchen-Sugarloaf Mountain area which they believed corresponds to a shallow vapor-liquid mixture or 'lossy' near surface lithology. No zone of significantly high attenuation was interpreted for the 5 to 12 km depth interval but high attenuation was noted below 12 km. Reasenberg et al. (1980) analyzed teleseismic P-wave residuals and mapped an area of approximately 0.2 s excess traveltime which they attributed to a low-velocity body between 5- and 20-km depth in the area of high heat flow and hydrothermal activity. They hypothesized that the low-velocity body could be caused by the presence of a partial melt in the middle crust.

Integrated Summary

Figure 32 summarizes the spatial overlap of the magnetic and resistivity lows, the 400 mW/m^2 heat flow anomaly and the anomalous ($\geq 26°C$) ground temperatures at 2 m depth. The data are superposed on alteration and thermal features mapped by Hulen (1978). The prospect areas as indicated by the various data sets are generally in good agreement except perhaps for the

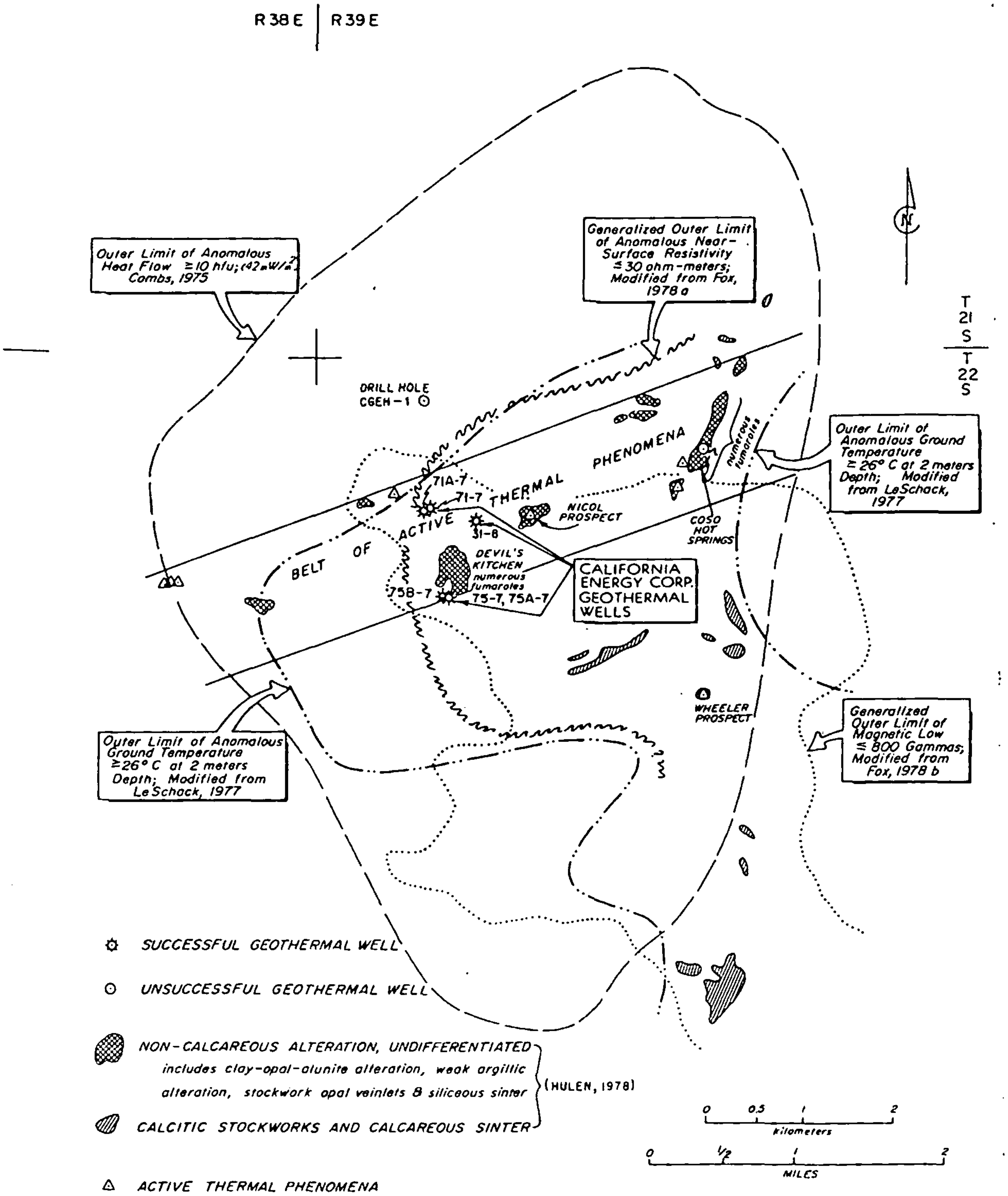


Figure 32. Geophysical Anomaly Summary for the Coso Geothermal Area.

extension of the heat flow high north of the belt of active thermal phenomena. The locations of several successful wells drilled by California Energy Corporation are also shown. This drilling has confirmed the presence of a high-temperature convective hydrothermal system in which the fluids are confined to major fracture zones within the crystalline rocks. Active exploration continues in the Coso area and future geoscientific studies and drilling will continue to improve our model of the hydrothermal system.

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