A-1 Characteristics of Resources

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4. Characteristics of Geothermal Resources

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Geothermal resources derive from the distribution of temperatures and thermal energy beneath the Earth's surface. Present-day technology emphasizes the production of electricity from geothermal steam; requirements include reservoirs of high temperature (at least 180°C, preferably above 200°C), depths of less than 3 km, natural fluids for transferring the heat to the surface and power plant, adequate reservoir volume (>5 km^3), sufficient reservoir permeability to ensure sustained delivery of fluids to wells at adequate rates, and no major unsolved problems. Because this configuration of characteristics occurs only rarely in the Earth's crust, other means of exploiting geothermal heat must be developed if the resource is to become much more than a curiosity.

The major known geothermal systems of the world are shown in Fig. 1. Table 1 summarizes the existing capacity of the major producing fields. These hydrothermal-convection systems, whether dominated by vapor or by hot water, are generally associated with tectonic-plate boundaries and volcanic activity. Fields along the belt of mountains extending from the Mediterranean eastward through Turkey to the Caucasus are related to the complex zone of collision between the Eurasian and African continents. Detailed plate-boundary relations of most geothermal areas of the western United States and central Mexico are not yet clearly established, but these areas seem to be associated with recent volcanism, high regional conductive heat flow, and relatively shallow depths to the mantle.

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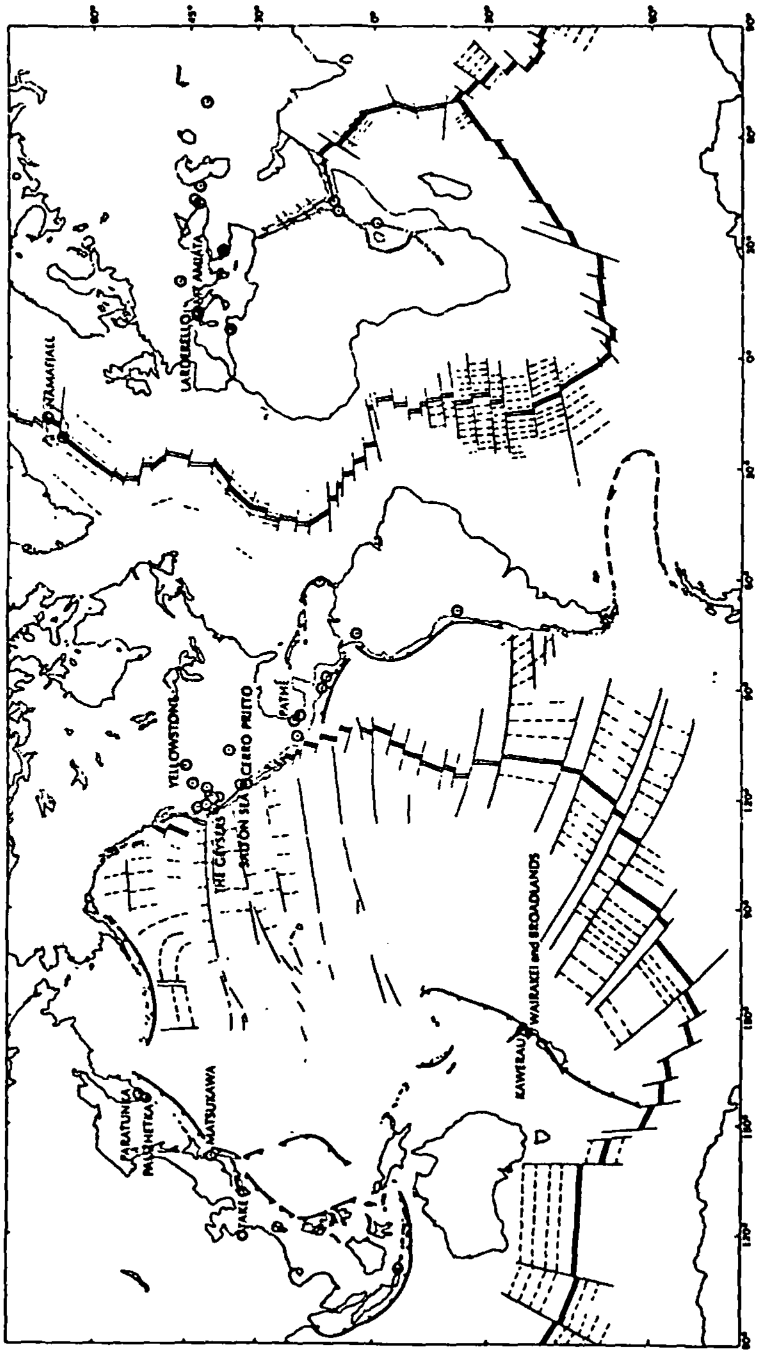

Fig. 1. The major known geothermal systems of the world. Named systems (other than Yellowstone) are those that are presently generating electricity or have power plants under construction (Muffler and White, 1972). Most fields are on spreading ridges (double lines) or rift valleys (heavy dotted lines) or are above subduction zones at plate boundaries (heavy barbed lines).

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TABLE 1
World Geothermal Power-Generating Capacity, 1972

Country	Field	Electrical capacity, Mw: Operating	Under construction	Vapor-dominated systems	Hot-water systems
Italy	Larderello	358.6		358.6	
	Monte Amiata	25.5		25.5	
United States	The Geysers[a]	302	110	412	
New Zealand	Wairakei	160			160
	Kawerau	10			10
Japan	Matsukawa	20		20	
	Otake	13			13
Mexico	Pathé	3.5			3.5
	Cerro Prieto		75		75
Soviet Union	Pauzhetsk	5			5
	Paratunka	0.7			0.7
Iceland	Namafjall	2.5			2.5
TOTAL		900.8	185	816.1	269.7

SOURCE: Muffler and White, 1972. Modified for present purposes.
[a] As of November 1972, additional capacity of 110 Mw is scheduled to be completed in 1973 and another 110 Mw (not listed) in 1974.

Near many hot-spring areas of the world, geothermal waters are used for space heating, horticulture, industrial processes, and spas (Muffler, in press). Such uses generally are individually small in scale, with the important exception of sites in Iceland (Pálmason and Zoëga, 1970), Hungary (Boldizsár, 1970), and Japan, but they demonstrate the potential of such sites for more extensive utilization along these varied lines.

Future technological developments may greatly change the requirements and uses summarized above. The most useful of the possible "breakthroughs" are itemized toward the conclusion of this paper and elsewhere in this volume.

Conductive Thermal Gradients

Temperatures below the Earth's surface are controlled principally by conductive flow of heat through solid rocks, by convective flow in circulating fluids, or by mass transfer in magma. Other modes of heat transfer are minor and are hereafter ignored. Moreover, transfer in magma is considered only through its effects on conduction and hydrothermal convection.

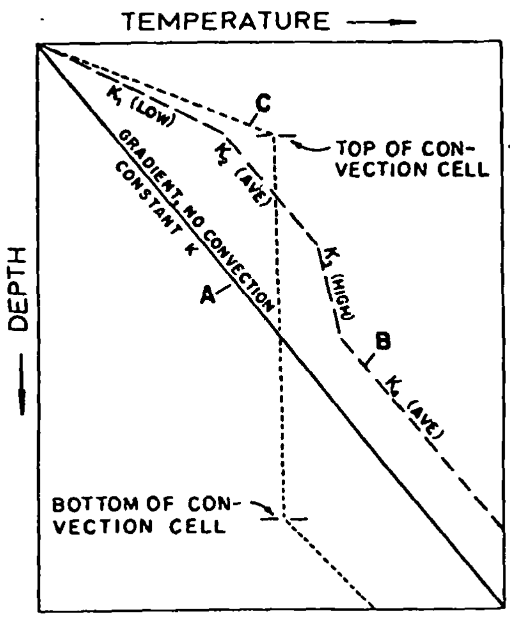

Fig. 2. Temperature/depth relations, where heat flow is controlled by thermal conduction in rocks of constant conductivity (A), or rocks of variable conductivity (B), or by major convective disturbance (C).

Conduction is the dominant mode of heat flow in most of the outer crust of the Earth. Where conduction is dominant, temperatures increase continuously with depth, but not at constant gradient. The important interrelations are those between thermal gradient, heat flow, and thermal conductivity of rocks, and the appropriate expression for the relationship is Fourier's law, $r = q/K$, where thermal gradient (r) is expressed in °C/km, heat flow (q) is in µcal/cm^2 sec, and thermal conductivity (K) is in mcal/cm sec °C. Thus, a measured thermal gradient is directly proportional to heat flow but inversely proportional to conductivity. Heat flow is the most fundamental parameter but ordinarily must be calculated from gradient and conductivity because, at low levels, heat flow cannot be accurately measured by any direct method.

In an area dominated by conduction and free from significant convective disturbances, heat flow is relatively constant in time and space, but conductivity of rocks differs greatly with depth as functions of mineralogy, porosity, and fluid content of pores. Therefore, temperature gradients may change greatly with depth, as in curve B of Fig. 2, which shows the effects of different thermal conductivities on thermal gradients, as compared with rocks of constant conductivity (curve A). A near-surface gradient cannot be reliably projected downward below explored depths because of likely changes in porosity and thermal conductivity with rock type and, especially, until possible convective influences can be evaluated.

Areas of near-"normal" conductive thermal gradient. The worldwide average heat flow is about 1.5 μcal/cm² sec (Lee and Uyeda, 1965; Sass, 1971), or 1.5 hfu (geothermal heat-flow units). This is about 1/2,000 of average solar energy at the Earth's surface, a very small quantity but an important one. For present purposes, we shall consider "normal" heat flow as ranging from 0.8 to 2.0 hfu. The thermal conductivities of most rocks range from 4 to 10 mcal/cm sec °C. Within these limits, temperatures can increase from 8° to 50°C/km (lines A and B, Fig. 3), averaging about 25°C/km (line C) or a bit more. At 3-km

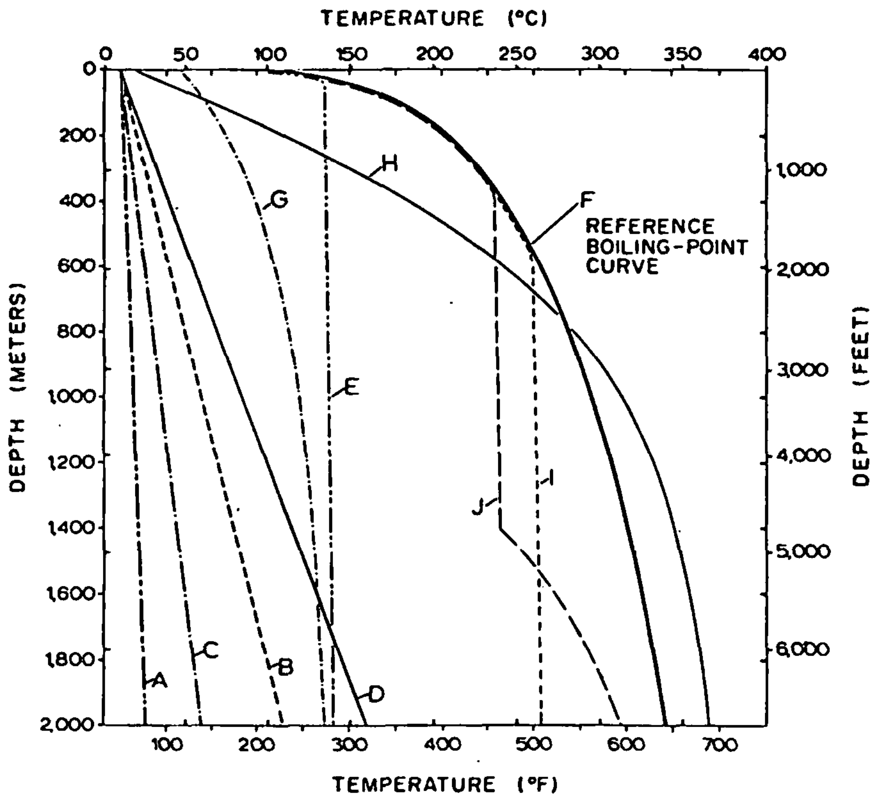

Fig. 3. Temperature profiles controlled by conductive gradients within the "normal" range (A, B, C) and above "normal" (D) and by convective transfer of heat in hot-water systems of different subtypes, compared with reference boiling-point curve (see text).

depth, with such gradients, temperatures range from 24° to 150°C above surface temperatures and average about 75°C. Most "normal" areas are not attractive for commercial geothermal exploitation, either now or in the immediate future, and their stored heat should not be considered as usable resources, much less as reserves recoverable under present conditions. Most, if not all, of this "low-grade" heat is as far removed from likely recovery as the trace contents of gold, copper, or uranium in average crustal rocks. Some areas may have a combination of heat flow in the higher part of the "normal" range and conductivities in the lower part of "normal"; such an area is the Gulf Coast of the U.S. (Jones, 1970), where gradients range up to 45°C/km or a bit higher (close to line B of Fig. 3). Such areas may prove to be exceptions warranting further study and evaluation, especially where existing oil and gas wells are already available for research utilization.

Areas of abnormally high conductive thermal gradient. Abnormally high thermal gradients result from unusually high heat flow, unusually low thermal conductivity, or favorable combinations of the two factors. In some large favorable areas, such as the Hungarian Basin (Boldizsár, 1970), conductive gradients range from 40° to 75°C/km (line D, Fig. 3) and perhaps locally even higher. Several rather large areas of high heat flow that seem unrelated to convection systems are now known in the United States. The "Battle Mountain High" of Sass et al. (1971) has an indicated average heat flow of about 3 hfu. But its thermal gradients, which range from 30° to 60°C/km, are not as high as would be expected, because of its relatively high thermal conductivities (~9 mcal/cm sec °C). An area near Marysville, Montana (Blackwell, 1969) has even higher heat flow (~7 hfu); rock conductivities are also high, and measured temperature gradients average about 75°C/km (line D, Fig. 3). Both areas may be related to large igneous intrusions. Another large area that is likely to have high conductive heat flow combined with somewhat lower, favorable conductivities is the region surrounding Clear Lake, California, where conductive gradients as high as 100°C/km would not be surprising.

Hydrothermal-Convection Systems

In hydrothermal-convection systems, most heat is transferred in circulating fluids rather than by conduction. Convection occurs because of the heating and consequent thermal expansion of fluids in a gravity field; heat, which is supplied at the base of the circulation system, is the energy that drives the system. Heated fluid of low density tends to rise and to be replaced by cooler fluid of higher density, which is supplied from the margins of the system. Convection, by its nature, tends to increase temperatures in the upper part of a system as temperatures in the lower part decrease. Moreover, convection (curve C, Fig. 2) obviously disturbs the pure conductive gradients (such as curves A and B, Fig. 2) that would otherwise obtain. Thus no single temperature gradient or heat flow can characterize a convection system. Gradients are commonly very high near the surface, and locally exceed 3°C/meter of depth; such a gradient, projected, exceeds 3,000°C at 1 km and is impossibly high, greatly exceeding the melting temperatures of all normal rocks (700° to 1,200°C). Where tested by drilling, temperature gradients in convection systems have been shown to decrease greatly with depth until the characteristic base temperature of the circulation system is attained. Locally, temperature reversals may occur.

Two major types of hydrothermal convection systems are recognized, differing in the physical state of the dominant pressure-controlling phase.

Hot-Water Systems

Hot-water systems are characterized by liquid water as the continuous, pressure-controlling fluid phase (White, Muffler, and Truesdell, 1971; White, 1970). Some vapor may be present, generally as discrete bubbles in the shallow, low-pressure zones. Continuity of liquid can be inferred with confidence from the distribution of pressures and from the abundance of constituents that are soluble in liquid water but have low vapor pressures and lack significant solubility in low-pressure steam. These include most of the constituents of ordinary water analyses, such as SiO_2, Na,* K, Ca, Mg, Cl, SO_4, HCO_3, and CO_3 (though B, CO_2, H_2S, and NH_3 are both volatile and soluble in water, and thus are not diagnostic).

Water in a major water-convection system (Fig. 4) serves as the medium by which heat is transferred from deep sources to a geothermal reservoir at shallower depths—shallow enough, perhaps, to be tapped by drill holes. Cool rainwater percolates underground from sur-

* The electrical charge of ionized constituents is not specified unless important to the discussion.

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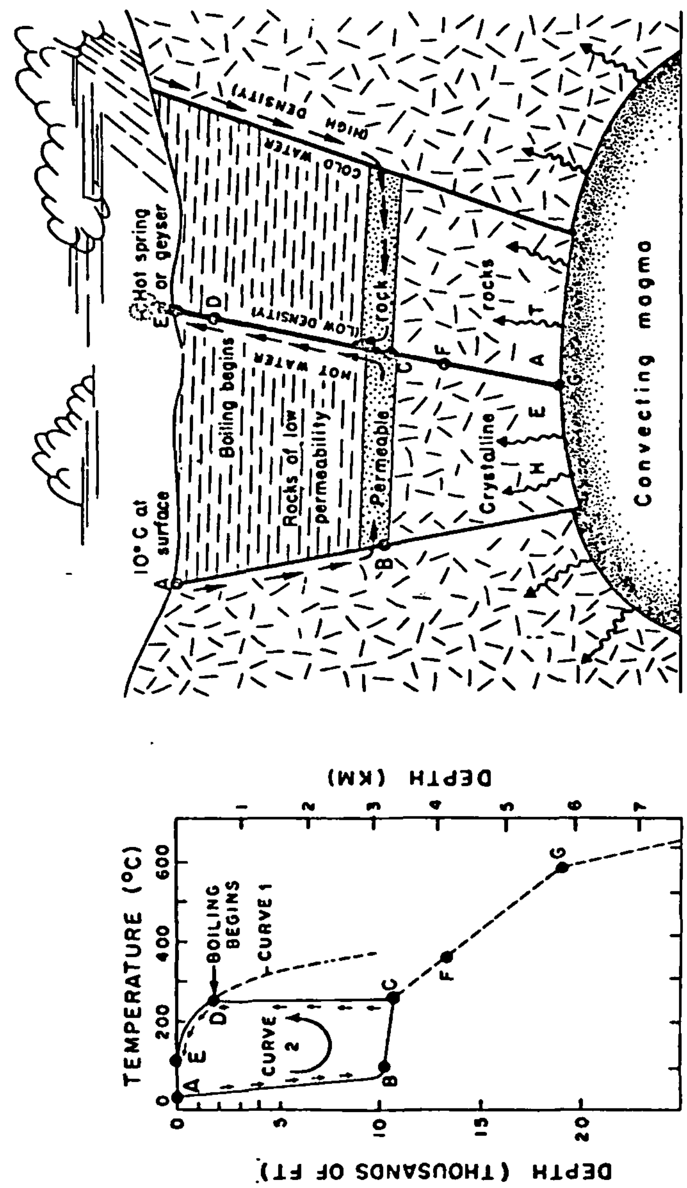

Fig. 4. Model of a high-temperature hot-water geothermal system. Curve 1 is the reference curve for the boiling point of pure water. Curve 2 shows the temperature profile along a typical circulation route from recharge at point A to discharge at point E.

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face areas ranging across tens to possibly thousands of square kilometers, and then circulates downward. At depths of 2 to 6 km, the water is heated by conduction from hot rocks that, in turn, are probably heated by molten rock. The water expands upon heating and then moves buoyantly upward in a column of relatively restricted cross-sectional area (1 to 50 km^2). If the rocks have many interconnected pores or fractures of high permeability, the heated water rises rapidly to the surface and is dissipated rather than stored. However, if the upward movement of heated water is impeded by rocks with few interconnected pores or fractures, geothermal energy may be stored in reservoir rocks below the impeding layers. Heat, of course, accounts for the density difference between cold, downward-moving recharge water and hot, upwelling geothermal water.

Some subtypes. Hot-water systems actually include many subtypes that are not yet universally accepted or precisely defined. Different classifications can be based on total salinity, dominant chemical characteristics, temperature range, structural and stratigraphic environments, presence or absence of permeable reservoirs, and insulating cap-rocks. I shall not attempt systematic classification by any single system, but some subtypes of particular interest to geothermal exploration to date include:

1. Systems characterized by low to moderate temperatures, generally ranging from about 50° to 125°C in most cases but reaching as high as 150°C in Iceland (Bodvarsson, 1964); and by chemical similarity to surface and shallow ground waters of the region. Some systems in the higher part of this temperature range may be characterized by impressive boiling springs of high discharge. A drill-hole into such a system is likely to show a temperature profile similar to that of curve E of Fig. 3. This curve assumes rapid upflow of liquid water under pressures that are close to hydrostatic. Pressures are too high for boiling to occur at existing temperatures except in the upper 20 or 30 m, where curve E impinges upon curve F. The latter curve, a significant and useful one for many geothermal considerations, describes the calculated temperatures for boiling of pure water at pressures controlled by liquid densities up to the ground surface; the water densities are corrected for the assumed temperatures. Actual temperature profiles in such systems depend to a large extent on their deep temperatures and their rates of upflow. With a high upflow rate (as assumed for Curve E)

little heat is lost by conduction from the margins of a system, but with a low upflow rate (as for curve G), a higher proportion of the heat contained in the water is lost by conduction; with decreasing rates of upflow, such systems grade into purely conductive environments.

2. Systems in deep sedimentary basins, commonly bearing saline waters of moderate temperature similar to oilfield waters. These waters are, at least in part, nonmeteoric in origin (White, 1967b; White, Barnes, and O'Neil, 1973). Some brine systems, such as Wilbur Springs in California, are characterized by springs that discharge thermal water, but convective circulation in most deep basins is so inhibited (by low permeability, salinity gradients, or low temperature contrasts) that conductive thermal gradients are only slightly disturbed, and thermal discharge is lacking. Hydraulic gradients, however, commonly provide some evidence for subsurface circulation.

3. Hot-water systems known to contain brines of very high salinity. The Salton Sea geothermal system and the Red Sea brine pools (Craig, 1966b; White, 1968a; Helgeson, 1968; Muffler and White, 1969; Ross, 1972) both contain brines of about 26 percent salinity. But the two systems differ markedly: temperature relations and the bulk chemistry of the associated rocks and sediments differ greatly between these two systems, probably accounting for major differences in composition of the brines. Because of the effect of salinity on boiling (Haas, 1971), the temperatures deep in the Salton Sea system (curve H of Fig. 3, from Helgeson, 1968) are considerably above the reference curve for pure water.

4. Systems with natural cap-rocks that tend to inhibit discharge and also to insulate their reservoirs; thus conserving heat. The Salton Sea and Cerro Prieto systems of California and Mexico have cap-rocks of low-permeability, fine-grained sediments. This configuration largely explains the conductive-gradient dominance of the low near-surface temperatures of curve H (Fig. 3) from the Salton Sea system.

5. High-temperature hot-water convection systems that tend to create their own insulating cap-rocks by "self-sealing"; hydrothermal minerals are deposited in pore spaces, especially in near-surface parts where temperatures decrease abruptly upward because of the influence of the boiling-point curve. Figure 4 and curve I of Fig. 3 are idealized from Wairakei, New Zealand, and the geyser basins of Yellowstone Park. Because of thermo-artesian pressure and channels restricted by

self-sealing, actual near-surface pressure gradients may exceed hydrostatic, as in Yellowstone Park (White et al., 1968), and temperatures may plot somewhat above reference curve F.

General characteristics. The principal characteristics of hot-water systems have been discussed by White, Muffler, and Truesdell (1971) and White (1970) and may be summarized as follows:

1. Hot springs are a common but not universal indication of hot-water convection systems. Where the water table is at or very near the ground surface, all or most discharge from a system is visible as hot springs, but where near-surface rocks are permeable and the water table is low, as much as 100 percent of discharge may be subsurface, and therefore dispersed into surrounding ground water and not directly evident at the surface.

2. Springs that are highest in temperature and discharge are generally also highest in SiO_2, Cl, B, Na, K, Li, Rb, Cs, and As, relative to surrounding ground waters. An analysis of the SiO_2 content and the Na-K-Ca ratios of such springs (Fournier and Truesdell, 1970 [1972]; 1973) is generally the best means of evaluating subsurface temperatures.

3. A "base" temperature (Bodvarsson, 1964, 1970) characterizes the deeper parts of many hot-water convection systems. The zone of little temperature change between points C and D of Fig. 4 represents the base temperature of the Wairakei system, which was about 260°C prior to exploitation (Banwell et al., 1957). The base temperatures of some other systems with moderate to low salinity (<5,000 ppm total dissolved solids) are as high as 300°C, but no such systems with substantially higher temperatures are yet known.

4. The insulated Salton Sea brine system is as hot as 360°C (curve H of Fig. 3, from Helgeson, 1968), and the Cerro Prieto system of Baja California (with about two-thirds the salinity of sea water) may be as hot as 388°C (Mercado, 1970); such temperatures, which are near or above the critical temperature of pure water (373°C), can occur in brines because their physical properties differ from pure water (Sourirajan and Kennedy, 1962; Haas, 1971).

5. SiO_2 is the most important self-sealant of hot-water systems. Quartz and chalcedony are generally dominant at temperatures above about 140°C, but opal and β-cristobalite are characteristic of low-temperature margins and self-sealed cap-rocks (White, Brannock, and

Murata, 1956; Honda and Muffler, 1970); zeolites, clay minerals, and calcite may also be important.

6. Natural geysers and amorphous or recrystallized SiO_2 deposited at the surface by flowing hot water imply upflow of subsurface waters with base temperatures of 180°C or higher (White, 1970; Fournier, 1972). Travertine ($CaCO_3$ from hot-spring waters), by contrast, implies low subsurface temperatures (or, more rarely, solution of limestone after temperatures have decreased nearly to surface temperatures).

7. Low-temperature convection systems have little potential for self-sealing because their waters are not sufficiently high in SiO_2. In fact, systems with maximum temperatures below about 150°C may in general become *more* permeable with time because as much as 140 ppm of SiO_2 is dissolved during the heating of cold meteoric water of low SiO_2 content. A temperature of 150°C is high enough to increase the porosity of quartz-bearing reservoir rocks but may not be high enough to be offset by hydrated alteration minerals, which tend to decrease porosity. This may explain why some systems of moderate temperature tend to discharge from large, impressive single springs, with no evidence of self-sealing with time (as for curve E, Fig. 3).

8. In contrast, high-temperature systems with maximum temperatures above about 180°C tend to decrease in permeability in their upper parts with time because of self-sealing. Perfect sealing may seldom occur; leakage takes place through any available permeable channels. New channels may form and old channels may be reopened by tectonic forces or by thermo-artesian buildup of vertical pressure gradients. Local pressure gradients may greatly exceed hydrostatic. Water-pressure gradients in the upper few hundred meters of research drillings in Yellowstone Park averaged about 20 percent above hydrostatic, and the excess in one drill hole was nearly 40 percent (White et al., 1968). Such high gradients are probably rare, and may tend to be localized in the shallow parts of convection systems.

9. Wells in permeable reservoirs generally produce 70 to 90 percent of total mass flow as water; the proportion of steam that forms when pressure is reduced is related to initial fluid temperature and to final separating pressure (Muffler, in press, Fig. 1). For example, water flashed to separator pressure of 50 psig from an initial temperature of

300°C yields 33 percent steam; 200°C yields 11 percent steam, but 150°C yields none!

10. Wells in ground of low permeability, however, may first erupt water and steam, which may then change to wet steam and finally to dry steam, in the manner of the eruption stages of some geysers (White, 1967a). The increasing steam content may seem to be a favorable characteristic, but it is not; the stored heat of the reservoir rocks is not only evaporating all local water, thereby temporarily producing steam, but also precipitating dissolved matter, thereby decreasing permeability.

11. In most hot-water fields, the only fluid that enters a producing well is liquid water. This remains entirely liquid as it flows up the well, until pressure decreases enough for steam bubbles to start forming. With continued flow upward, more water "flash boils" to steam (as pressure and mixture temperature decrease). The buoyancy of the expanding steam displaces the remaining water upward, thereby increasing the velocity of the mixture and ejecting the residual water above the ground surface (unless diverted horizontally in pipes or separator), just as in natural geysers. The depth of first boiling in the well depends mainly on the initial temperature of the water, but also on formation-fluid pressure and separator pressure. Crude depths of first boiling can be calculated, but the principles are rather complex. A better method, where adequate equipment is available, is to measure in-hole pressure gradients under producing conditions; a break in slope of the pressure gradient indicates the depth of first boiling (where water flashes to steam).

12. The chloride content of water that has been above about 150°C is nearly always higher than 150 ppm. However, a very few hot-water systems with Cl content as low as 40 ppm have temperatures above 200°C.

13. Chloride is the most critical single constituent in distinguishing hot-water systems from vapor-dominated systems. Most metal chlorides are highly soluble in liquid water, and the chloride of most rocks is easily leached by water at high temperature (Ellis and Mahon, 1967). The common metal chlorides, however, have negligible volatility at temperatures as high as 400°C (Krauskopf, 1964) and are not appreciably soluble in low-pressure steam. Thus a chloride-bearing water

body definitely indicates a hot-water system (Cl > 50 ppm). Some hot-water systems, however, do produce surficial acid-sulfate springs that are low in chloride; such springs are sustained by steam boiling from an underlying chloride-bearing water body, but are otherwise chemically similar to springs associated with vapor-dominated systems.

Vapor-Dominated Systems

A few geothermal systems, including the important Larderello fields of Italy and The Geysers of California, produce dry or superheated steam with no associated liquid. For this reason they are commonly known as "dry-steam" systems, but White, Muffler, and Truesdell (1971) conclude that liquid water and vapor normally coexist in the reservoirs, with vapor as the continuous, pressure-controlling phase. Thus "vapor-dominated systems" seems to be a more appropriate term.

Opinions on the physical nature of the initial fluid(s) in the Larderello fields include initial saturation with liquid (Facca and Tonani, 1964; Marinelli, 1969) and superheated steam in a reservoir that, with production, is replenished by boiling from a deep water body (Elder, 1965; James, 1968) that may be a brine (Craig, 1966a). Sestini (1970) favors a reservoir filled largely with vapor but locally containing "disturbance water" (p. 637) that evidently has no significant influence on subsurface pressures. According to White, Muffler, and Truesdell (1971), vapor-dominated systems of the Larderello type develop initially from hot-water systems characterized by very high heat supply and very low rates of recharge; if and when the heat supply of a developing system becomes great enough to boil off more water than is being replaced by recharge, a vapor-dominated reservoir begins to form (in known systems, probably thousands to tens of thousands of years ago). The fraction of discharged fluids that exceeds recharge is supplied by water previously stored in the larger fractures and pores. Some liquid water is held in the smaller pores and on fracture surfaces by surface tension, and some additional water is no doubt locally retained in closed pores and various open spaces that terminate downward. Conductive heat losses from the margins of a reservoir, moreover, result in condensation of steam. And, finally, liquid not retained in all these ways drains downward under gravity to deeper water-saturated rocks. Vapor, however, is the continuous, pressure-controlling phase in large pores and open channels.

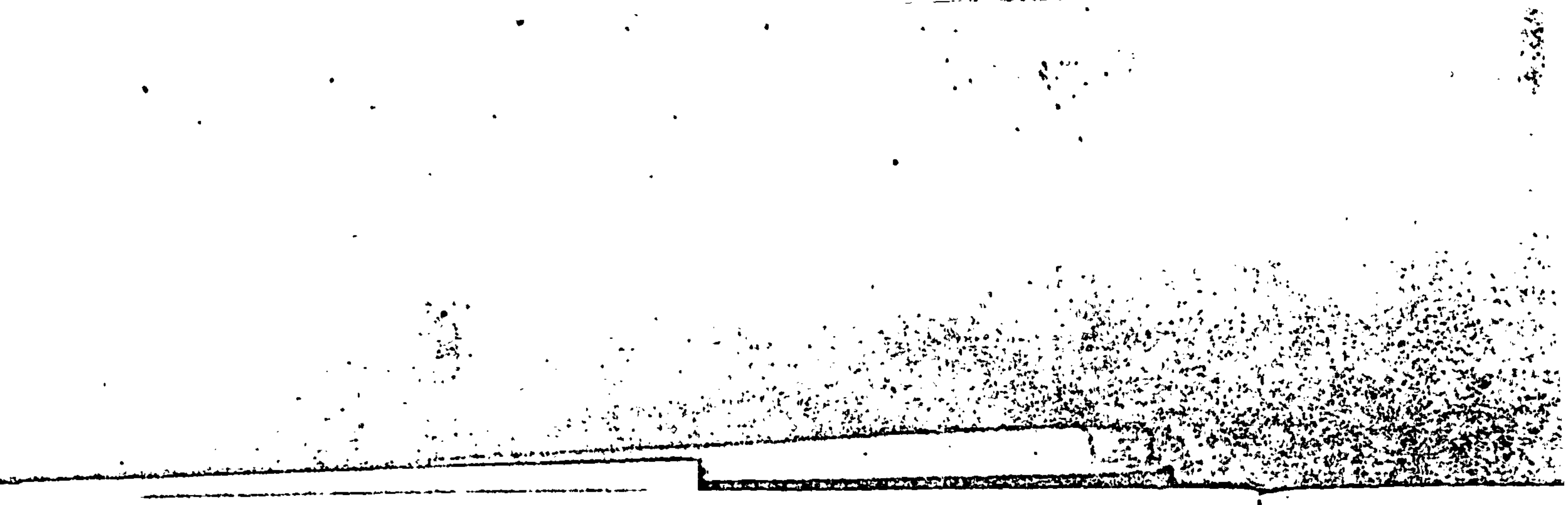
According to this model, a deep water table, perhaps crudely horizontal, separates an underlying zone, which is dominated by liquid, from the overlying zone, where pressures are controlled by the vapor phase. This deep water table continues to decline as long as the fluid discharged (mainly as steam) exceeds the recharge; and at the same time the reservoir of steam continues to develop. Only a part of the steam from the deep water table is discharged at the surface as vapor; probably most of the total upflow recondenses to liquid on the margins of the reservoir, where heat can be lost by conduction, as mentioned above. A conductive heat flow of 20 hfu from the margin of the reservoir to the surface, for example, requires the heat from 29 kg condensing steam per km² of surface area per second (White, 1970). Conductive heat flow through the reservoir cannot supply this much heat because of the reservoir's nearly constant temperature; the heat of vaporization of steam provides the only reasonable explanation for the phenomenon. An important implication of this conclusion is the downward draining of condensate through the reservoir, ensuring the coexistence of liquid and vapor in the natural systems prior to exploitation.

Two subtypes of the vapor-dominated system, the Larderello and the Monte Amiata, appear to be distinguishable.

Larderello subtype. The physical, chemical, and geologic characteristics of The Geysers, Larderello, and Matsukawa vapor-dominated systems, as known to date (White, Muffler, and Truesdell, 1971; White, 1970; Nakamura et al., 1970), are consistent with the model just described and may be summarized as follows:

1. Reservoirs occurring at or below about 350 m in depth tend to have *initial* temperatures near 240°C (curve J, Fig. 3) and pressures near 35 kg/cm². The published data for Larderello and Matsukawa are convincing (Penta, 1959; Burgassi, 1964; Sestini, 1970; Nakamura et al., 1970), but the fragmentary data from the deep reservoir of The Geysers system are less conclusive (Otte and Dondanville, 1968). Little initial physical difference is yet recorded for holes drilled at greater depths (changes in Larderello with exploitation are considered in item 9).

2. The relatively uniform initial temperatures and pressures are evidently strongly influenced by the maximum enthalpy of saturated steam (669.7 cal/g at 236°C and 31.8 kg/cm²: James, 1968; White, Muffler, and Truesdell, 1971; Sestini, 1970, p. 625). As the gas content



of the vapor increases above a few percent, these physical characteristics change greatly (White, Muffler, and Truesdell, 1971). For example, at a constant temperature of 236°C for coexisting liquid and vapor, 1 percent of other gases in the vapor increases the total pressure only to 32.1 kg/cm^2; but corresponding pressure for 5 percent of other gases is 33.5 kg/cm^2, and that for 10 percent is 35.3 kg/cm^2.

3. Pressures in these vapor-dominated reservoirs are well below hydrostatic and, with few exceptions, the difference increases with depth (Truesdell, White, and Muffler, in preparation). In-hole pressures, of which few details are yet published, increase only slightly with depth because of the low density of the pressure-controlling vapor. Expected changes related to depth alone are given by White, Muffler, and Truesdell (1971, Table 3). Obviously, such a system could not form or persist if the water-saturated rocks that surround the reservoir could supply a high rate of recharge. The water thus supplied would flow into the reservoir under hydrostatic drive at a rate exceeding discharge, and the underpressured reservoir would "collapse."

4. Fumaroles, mud pots, mud volcanoes, turbid pools, and acid-leached ground characterize the discharge areas where surface activity is most intense. Springs in such areas are generally acid because of the H_2SO_4 produced by oxidation of H_2S in the escaping gas; pH's are as low as 2 to 3 except where NH_3 is abundant enough to neutralize the acid. Sulfate contents tend to be high, but Cl contents are uniformly low (<15 ppm). Likewise, the springs, streams, and ground water of the immediately surrounding area are low in chloride. Areas lacking intense surface activity are characterized by slightly acid to slightly alkaline bicarbonate-sulfate spring waters that may be high in total CO_2, B, or NH_4, but are low in Cl; some spring waters of such areas are also anomalously high in SiO_2.

5. Where the natural total discharge of fluids from vapor-dominated systems has been observed closely prior to exploitation, the discharge is consistently low, ranging from a few tens to several hundred liters per minute. Detailed descriptions of natural discharge at Larderello prior to initial subsurface exploitation in 1904 apparently do not exist. Sestini (1970), in attempting to evaluate the early records, concludes that "from documentation and other evidence, there is reason to believe that the total flow of water, steam, and gas was in the order of some hundreds of tons per hour." This flow is equivalent to a few thou-

sand liters per minute, expressing all H_2O as liquid, and is probably much too high for discharge solely from the reservoirs, judging from my observations of The Geysers and Matsukawa prior to exploitation, and of virgin systems in Yellowstone Park (Wyoming), Lassen Park (California), Steamboat Springs (Nevada), and Valles Caldera (New Mexico, perhaps a vapor-dominated system). This Larderello flow is *not*, however, unreasonably high if much of the credited discharge consisted of near-surface runoff heated by rising steam. And, in fact, most of the natural thermal discharge of liquid from vapor-dominated systems, especially at The Geysers and Lassen Park, consists of steam-heated surface waters.

6. Production wells normally produce dry to superheated steam (from a few degrees to more than 50°C of superheat); however, liquid water evidently occurs in some noncommercial wells on the borders of reservoirs and in the fluid initially produced from some wells that change from wet steam (i.e. steam containing a little water) to dry steam. Many shut-in wells (i.e. wells that have been capped off to recover earlier output pressure) contain vapor as the only fluid, but Sestini (1970, p. 636) notes that "fairly high-temperature water" flows into some Larderello wells that have been shut in for a while. Sestini calls this "disturbance water," without discussing its origin or characteristics in detail; it presumably is water locally perched on impermeable rocks or contained in downward-terminating fractures or caverns, and is surely *not* a part of a large liquid-dominated water body, since other, deeper drill holes are characterized by initial pressures of 31 to 34 kg/cm^2 (Sestini, 1970, p. 640) that are typical of vapor-dominated systems.

7. Most of the heat content of the reservoir is stored in solid phases (James, 1968; White, Muffler, and Truesdell, 1971; Truesdell, White, and Muffler, in preparation), which generally carry 80 to 90 percent of total heat.

8. Superheated steam forms from saturated steam by flow and decompression through hot rocks already dried by transfer of heat from solid phases to evaporating pore liquid. Critical aspects of these relations are the stored heat of solid phases and the decrease in boiling temperature with decrease in pressure (Truesdell, White, and Muffler, in preparation).

9. With long-term production, most Larderello wells show a rather

steady increase in well-head temperature (as high as 260°C in 1966, with most wells then starting to decline: Sestini, 1970). Enthalpy had increased to as much as 710 cal/g, along with increasing superheat. These gradual changes had been implied in earlier publications but are now clearly substantiated by Sestini, who proposes an increasing dependence on supercritical H_2O (>374°C) from a deep magmatic environment. Probably a more satisfactory explanation consists of increasing dependence on boiling from a deep, declining, saline-water body that becomes increasingly more saline as water is vaporized (Truesdell, White, and Muffler, in preparation). Curve J of Fig. 3 shows a possible distribution of temperatures within a deep brine-water body (assumed 25-percent salinity, starting at a depth of 1,400 m). The maximum enthalpy of steam coexisting with 25 weight percent NaCl brine, according to Haas (1971; written communication, 1972), is 681.5 cal/g at 275°C. Much higher steam temperatures and enthalpies are obtainable as residual salts become more highly concentrated and boiling occurs in environments of much greater initial temperature; Haas's data are computed only to 35 percent NaCl, where coexisting steam has its maximum enthalpy of 688.7 cal/g at about 295°C.

The previous discussion is concerned chiefly with systems such as Larderello, The Geysers, and Matsukawa (White, 1970) that are characterized by initial reservoir temperatures near 240°C, shut-in pressures near 35 kg/cm², and contents of gases (other than steam) of about 5 percent or less. My associates and I conclude that discharge areas seem essential for such systems, thus permitting the net loss of much initial pore water to establish domination by the vapor phase and the flushing out of gases other than steam. A large vigorous system is likely to have at least one prominent vent area that cannot be accommodated by discharge into ground water. Under some circumstances (see point 4 in the characteristics just concluded), less vigorous discharge of steam and gases can be accommodated. If CO_2, H_2S, and other gases are not permitted to escape, they are selectively concentrated on the cooler borders and tops of reservoirs as heat is lost by conduction; the decreasing of temperatures upward and outward requires some condensation of steam (see White, Muffler, and Truesdell, 1971, Table 5). The flow of fluid in systems of this subtype is limited by the low permeability of the *recharge* channels; these channels constitute the limiting impedance on fluid flow throughout the system.

Monte Amiata subtype. A second variety of vapor-dominated system,

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here called the Monte Amiata (Italy) subtype, is not yet well understood but is evidently similar in many respects to hot natural-gas fields (White, Muffler, and Truesdell, 1971). Temperatures in the Monte Amiata fields tend to be much lower (~150°C) and initial gas contents much higher (>90 percent: Burgassi et al., 1965; Cataldi, 1967) than for comparable initial pressures (20 to 40 kg/cm²) in the Larderello fields. Thus, steam is a relatively minor initial constituent, presumably because of condensation of water vapor near the relatively cool borders of the reservoirs, as previously mentioned. With production and decompression, the initial vapor of high gas content is flushed out of the reservoir and is replaced by the relatively low-pressure steam of lower gas content that results from the boiling of water at only modest temperature. Another characteristic of the fluids produced from the Bagnore field of the Monte Amiata district is a trend from dry vapor to vapor plus liquid H_2O (Cataldi, 1967, Table 2).

The latter relations are interpreted to indicate water-flooding, or a rise in a deep water table that responds to the surrounding hydrostatic environment as reservoir pressures decline with production. The most restrictive impedance to fluid flow for the Monte Amiata subtype evidently occurs in the *discharge* part of the system, where low-permeability cap-rocks limit the discharge of gases to rates that are equal to or less than rates of generation or supply of gases. In contrast to the Larderello subtype, discharge of gases and steam is not required to form or maintain vapor-dominated reservoirs of the Monte Amiata subtype, although some leakage of gases is no doubt characteristic of most reservoirs.

Problems of Utilization

The problems attendant upon large-scale utilization of the various geothermal-field types and subtypes are manifold. A basic listing might proceed as follows:

1. Because of its special geologic and physical requirements, the commercially attractive Larderello subtype of vapor-dominated system is rare, accounting perhaps for only 5 percent of all geothermal systems with temperatures above 200°C; the advantages of this subtype for utilization are demonstrated by its dominance of present geothermal power-generating capacity (an estimated 73 percent of the world total, operating or under construction through 1973).

2. A discharge area is probably essential for the Larderello subtype,

with characteristic, recognizable manifestations of activity, chemistry, and ground bleaching. If this is so, then completely concealed deep systems are not available for future discovery.

3. The Monte Amiata subtype of vapor-dominated system, characterized by relatively high content of noncondensable gases and moderately low temperatures, may be more common than the Larderello subtype but will be more difficult to discover because of the absence of conspicuous surface characteristics. In any case, because of its physical and production characteristics this subtype is not as attractive for exploitation (2.3 percent of Table 1).

4. The high-temperature hot-water fields (25 percent of Table 1) are attractive for near-future increases in power production, but present utilization technology is not efficient, converting only about 1 percent of stored reservoir energy into equivalent electrical energy (Bodvarsson, 1970; Muffler, in press).

5. The water of many hot-water systems, when flash-erupted and cooled, deposits SiO_2 or $CaCO_3$ scale in wells and surface pipes; if similar flashing and mineral deposition occur in the reservoir immediately adjacent to wells, permeability and production rates decrease dramatically.

6. Some hot waters are corrosive because of high salinity, high CO_2 or O_2 content, or high acidity from H_2SO_4 or, rarely, HCl.

7. Some hot-water systems do not have adequate volume, temperature, or permeability to maintain commercial production. White (1968b) has suggested, from general experience at Broadlands and Waiotapu, New Zealand, and Beowawe and Steamboat Springs, Nevada, that inadequate permeability and reservoir characteristics may be as common as inadequate temperature.

8. Most hot-water effluents involve some environmental hazard, since they are generally higher in dissolved salts, B, NH_3, As, and heavy metals than most surface and ground waters. Such effluents will require disposal by some satisfactory means, with reinjection generally favored.

9. Some hot-water effluents may not be compatible with reservoir or other formation fluids, even though the fluids were initially identical. Compatibility and reliability of reinjection must be tested, and better principles for early recognition of the attendant problems must be developed. Reinjection has been tested for one year in the Salton Sea

system and for short intervals in the Long Valley system of California and the Ahuachapán field of El Salvador, but the only long-sustained test (about 3 years through 1972) has been at The Geysers in California. In the latter field, cool condensate is being reinjected successfully into an underpressured vapor-dominated reservoir; presumably much of the liquid is vaporized by transfer of heat from the still-hot rocks. These results are commonly interpreted as proof that reinjection elsewhere will be equally successful; but because individual hot-water systems vary greatly in fluid chemistry and precipitation potential, such a conclusion is hazardous without adequate testing. Production engineers estimate that if reinjection is successful for 1 year, it can be continued for several more years, indeed for the mechanical life of the injection equipment (Otte, private communication).

10. Most chemical problems are not serious for the low-temperature hot-water systems, but self-eruption will be unreliable or lacking for water that is too low in temperature or that must be "steam-lifted" from depths too far below the ground surface (Bodvarsson, 1970). The percentage of water that flashes to steam in a producing geothermal well depends mainly on the initial temperature and the pressure of steam separation from residual water, with liquid constituting 70 to 90 percent of most commercial production. The steam-lifting of moderate-temperature waters (150° to 200°C) becomes increasingly less effective as reservoir pressures and temperatures decline with production. As water levels (fluid potentials) decline below the ground surface, the energy required to lift liquid water increases. Thus, if produced at all, such water probably must be pumped.

11. Desalination of low-temperature waters involves more chemical and effluent-disposal problems, since the dissolved solids are concentrated into a small proportion of residual water. Soluble constituents, such as NaCl, normally will not precipitate, but constituents of low to very low solubility, such as SiO_2, $CaCO_3$, and $CaSO_4$, are potentially troublesome.

12. Thermal, noise, and air pollution (principally H_2S) may constitute environmental hazards requiring some control, depending on their severity.

13. Seismic hazards from reinjection must also be evaluated, especially for hot-water systems. But reinjection into underpressured vapor-dominated systems should involve little or no seismic hazard.

14. Subsidence will occur over hot-water reservoirs consisting in part of clay, silt, or shale where produced water is not locally replaced by reinjection. Sand and sandstone are less subject to compaction, unless pore fluids are overpressured. But subsidence over vapor-dominated reservoirs (initially already underpressured relative to hydrostatic) is likely to be slight.

Recoverability and Reserves of Geothermal Energy

This paper focuses attention on the physical and chemical nature of potentially useful concentrations of geothermal heat near the surface of the Earth. For the most part, the hydrothermal (vapor-dominated or hot-water) convection systems have been emphasized, since these include the principal geothermal resources that are usable and recoverable under present economic and technological conditions.

One or more of several potentially important breakthroughs in utilization technology may greatly expand the development of geothermal systems, hopefully in the immediate future. The most significant of the possible breakthroughs are:

1. Heat-exchange technology that would permit utilizing the heat from fluids down to 100°C or less (Jonsson, Taylor, and Charmichael, 1969), since total heat contained in easily recoverable natural fluids at temperatures of 100° to 180°C is far greater, perhaps by a multiple of 100, than total easily available heat above 180°C (see Anderson's discussion of the vapor-turbine cycle, this volume).
2. Multipurpose developments, including desalination and/or chemical recovery, that would yield significant sharing of total costs.
3. Low-cost mechanical, chemical, or nuclear fracturing of hot, dry rocks to increase permeability, thus permitting introduction of fluids and recovery of stored energy (these are described in later papers in this volume).
4. New methods for drilling low-cost holes to great depth.
5. New technology or other developments that favor wide applications to space heating, horticulture, and product processing.
6. Solution or control of all geothermal-resource problems at no greater cost than for corresponding environmental and other problems of competing sources of power.

Some of these breakthroughs could have profound effects on the recovering of geothermal energy from very large gradient-dominated volumes of rock, such as the deep sedimentary basins and hot, dry crystalline rocks, which are unlikely to be utilized within present prices and technology.

Strikingly disparate estimates have been made in recent years for the power potential, desalination potential, relative costs, and environmental-pollution aspects of geothermal energy. Depending on the source, the expressed view ranges from conservative (locally important, but with a relatively small potential for supplying national needs) to highly optimistic (very promising, with implied reliable potential for supplying a major part of all future needs for both power and desalinated water). The discrepancy is related largely to (1) a lack of agreement on the various categories of resources, with respect to certainty of existence and feasibility and cost of recovery (McKelvey, 1972; geothermal factors treated in detail by Muffler, in press); (2) differing assumptions on future technology and on whether hoped-for breakthroughs are likely to be realized with reliably predicted costs; and (3) a lack of agreement on the characteristics and nature of different types and subtypes of geothermal deposits, with respect to individual problems of discovery and energy recovery.

In my opinion, world geothermal power production is unlikely to exceed 30,000 Mw with present prices and technology. My estimate of proved, probable, and possible reserves recoverable in the United States under present conditions is approximately 600 Mw-centuries. Paramarginal reserves (recoverable with present technology but with as much as one-third increase in price) may be from 2,000 to 4,000 Mw-centuries. I am reluctant to offer estimates of geothermal resources that are now submarginal but that may be utilized with appropriate technological breakthrough; adequate cost data are completely lacking. However, major geothermal contributions (>10 percent of our energy needs) could result from such breakthroughs.

In contrast, the geothermal resource base (total stored heat, without regard to cost of recovery) can be estimated with some reliability, depending only on the assumed depth (3 km, 10 km, etc.) and thermal gradient, by utilizing the concept of volumetric specific heat (White, 1965). In that study I assumed an average gradient of 20°C/km and calculated 3×10^{26} cal of stored heat (i.e. heat above surface temperatures) under the surface of the Earth to a depth of 10 km, with 6×10^{24} cal of that worldwide total under the United States. The assumed average gradient may be too low, and 25°C/km may be a more likely average; but even if an improbable 30°C/km is assumed, the resource-base calculations are raised only to 4.5×10^{26} cal and 9×10^{24} cal, respectively. These calculations define upper limits for recoverable resources and are probably too high by at least two orders of magnitude.

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