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GEOTHERMAL RESERVOIRS IN HYDROTHERMAL CONVECTION SYSTEMS

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Abstract Geothermal reservoirs commonly exist in hydrothermal convection systems involving fluid circulation downward in areas of recharge and upwards in areas of discharge. Because such reservoirs are not isolated from their surroundings, the nature of thermal and hydrologic connections with the rest of the system may have significant effects on the natural state of the reservoir and on its response to development. Conditions observed at numerous developed and undeveloped geothermal fields are discussed with respect to a basic model of the discharge portion of an active hydrothermal convection system. Effects of reservoir development on surficial discharge of thermal fluid are also delineated.

Introduction Geothermal reservoirs containing concentrations of recoverable thermal energy exist under a variety of geologic and hydrologic conditions. This paper focuses on reservoirs in hydrothermal convection systems (HCS), in which heat transfer by fluid movement is the dominant process for concentrating thermal energy in reservoirs at exploitable depths. Geothermal reservoirs in conduction-dominated systems which do not involve significant vertical fluid movement are discussed in Muffler (1979), Smith (1978), and Sorey and others (1983), and Batchelor (this volume).

Development of an adequate understanding of basic relationships between fluid, rock, and heat in HCS requires synthesis of information generated in various types of investigations. These include studies of geologic structure and volcanic history, geophysical properties of the subsurface, geochemistry of surface and subsurface fluids, and hydrology. Coupling of results from such studies with data from drilling and well testing provides the best means for locating geothermal reservoirs and evaluating their potential for energy development. The importance of this synthesis is often overlooked, as is the comparison of similarities and differences between HCS in different areas.

The principal components of a HCS are the heat source, the working fluid(s), and the means for moving fluid through the system (permeable rock and pressure gradients). Although differences in the nature of these components give rise to significant variations in detail among HCS, it is possible to identify a simple basic model that describes the essential features of heat and fluid flow in and through reservoirs in most HCS. Various types of HCS discussed in the literature, for example, hot-water, liquid-dominated, vapor-dominated, high-temperature, low-temperature, can be compared and elucidated with reference to this basic model. The model is also useful for discussing changes in HCS accompanying fluid production for energy development.

The existence of HCS is commonly indicated by surficial discharge of hot water or steam. The thermodynamic and chemical states of these surface fluids are related to those of the underlying reservoir fluid. Care must be taken, however, in interpreting subsurface conditions based on the locations and the nature of surficial discharge features. In addition, the possibility of alteration or cessation of surface discharge in response to geothermal development may be an important limiting factor in such development, as discussed below.

Thermal Settings The thermal setting within which a HCS occurs refers to the nature of the heat source at depth and the interactions between circulating fluid and the heat source. Studies of regional conductive heat flow provide the background from which to identify anamolous heat flow and temperature gradients related to shallow crustal heat sources and/or convective heat transfer (Sass and others, 1981). On a regional scale, it is unlikely that conductive heat flow derived from mantle and lower crustal sources and radioactivity of the upper crust can exceed 3 HFU (130 mW/m^2), or would fall below about 1 HFU (40 mW/m^2). Local areas with heat flow outside this range could be considered anamolous and indicative of the factors noted above.

Shallow crustal heat sources are formed by silicic magma bodies whose residence time can be substantial (10^5 -10^6 years), especially where resupply of magma from deeper levels occurs. Long residence time may also be a requirement for the establishment of extensive HCS and associated geothermal reservoirs. Magma chambers underlying HCS have been delineated at many locations in the western United States, including the Rio Grande Rift and Valles Caldera in New Mexico (Sanford and Einarsson, 1982), Yellowstone Caldera in Wyoming (Iyer, 1979), Long Valley Caldera in California (Steeples and Iyer, 1976), and The Geysers in California (Iyer and others 1979). Depths to such chambers range from about 5 to 10 km; areal extents are on the order of 100 km^2. Magma bodies at shallower levels support HCS in ocean rift environments, as in Iceland (Einarsson, 1978) and Hawaii (Furumoto, 1978), and may function similarly in the Cascade Range in Oregon and California (Muffler and others, 1982, and MacLeod and Sammel, 1982).

Conductive heat flow in regions underlain by shallow crustal heat sources for periods exceeding 10^5 years should be greater than 3 HFU. For 800 °C magma at 8 km, for example, conductive heat flow would approach 5 HFU for average crustal rock thermal conductivities (5 mcal/scm°C). Where a HCS exists above such a heat source, the heat flux calculated by integrating convective and conductive heat output over the total area of the system will exceed 5 HFU. Thus, hydrothermal circulation decreases the "effective depth" to magma, or the depth that would be implied by heat flow observations if convection were absent.

A useful relation was developed by Lachenbruch and Sass (1977) for the "effective depth", H_e, as a function of mean thermal conductivity, K, magmatic temperature, T, and measured heat flux, q, assuming steady-state conditions:

$$H_e = K\ T\ /\ q \tag{1}$$

In the case of the Long Valley caldera, q = 13 HFU and from (1), H_e = 3 km for K = 5 mcal/scm°C and T = 800°C. This implies that if hydrothermal circulation has been continuous at near present-day rates for 10^5 years or more, circulation must extend downward to within about 3 km of the magma chamber or to depths of 3-5 km. More detailed numerical simulations indicate that if the present HCS were more recently developed, depths of fluid circulation could be shallower (Sorey, Lewis, and Olmsted, 1978).

Hydrothermal convection systems also develop in regions where shallow crustal heat sources are absent. Most of the identified HCS within the Basin and Range Province occur in regions of slightly elevated crustal heat flow (2-3 HFU) and extentional tectonics. In such systems fluid circulation cannot significantly increase the average crustal heat flow, and under steady-state conditions, average surficial heat flow from conduction and convection cannot exceed the regional value. This places significant constraints on the depth of circulation and the areal extent of HCS.

The area, A, over which circulating fluid must adsorb a portion of the regional heat flow, Δq, can be calculated for steady-state conditions as

$$A = Q\ \Delta T\ c\ /\ \Delta q \tag{2}$$

where Q = total mass throughflow, ΔT = increase in fluid temperature between recharge and discharge areas, and c = average fluid specific heat. For parameters typical of high-temperature HCS in the Basin and Range (Q = 10 kg/s, ΔT = 200°C, Δq = 2 HFU), an area of 100 km^2 is obtained. Such results tend to preclude circulation systems restricted to a single fault conduit, except for convective activity of short duration or great depth (Welch, Sorey, and Olmsted, 1981).

Estimates of depths of fluid circulation in HCS supported by regional conductive heat flow are dependent on knowledge of the areal extent and the duration of convective activity. Minimum depths of circulation, D_{min}, are given by

$$D_{min} = K\ \Delta T\ /\ q \tag{3}$$

For K = 5 mcal/scm°C, ΔT = 200°C, and q = 3 HFU, D_{min} = 3.3 km. Downward circulation depresses the temperature distribution relative to the conduction-only isotherms; upflow has the opposite effect. With reference to (2), values of A decrease as the depth of circulation increases because Δq increases.

Hydrologic Settings Although HCS occur within a variety of geologic and thermal settings, several aspects appear common to those within continental regimes. The dominant source of fluid is of meteoric origin. Studies of stable oxygen and hydrogen isotopes in thermal and non-thermal fluids in many systems demonstrate the meteoric source and in many cases delineate potential recharge areas (Craig, Boato, and White, 1956; Arnason, 1976; Sorey, Lewis, and Olmsted, 1978). Enrichment of He^3 relative to He^4 in thermal fluids (water and/or steam) indicates a component of magmatic origin in HCS in Iceland (Mamyrin and others, 1972), Yellowstone National Park and Lassen Volcanic National Park (Craig and others, 1978), and at Cerro Prieto (Welham and others, 1979). Enrichment of He from magmatic sources is commonly observed in submarine HCS associated with spreading ridges (Lupton, 1979).

An active HCS is one in which fluid circulation is presently occuring, downward in recharge areas and upward toward discharge areas. Because of the potential for chemical self-sealing in discharge areas, HCS can become inactive, in some cases leaving behind reservoirs of stored thermal energy in so-called blind anamolies (Flynn and Trexler, 1982; Phillips Petroleum Co., unpublished data on Desert Peak, NV). In active HCS, secondary permeability created by faulting plays an important role in the establishment of fluid circulation to substantial depths and maintenance of permeability at shallow depths. As noted in the previous section, depths of circulation for high-temperature HCS (>150 °C) can be several kilometers or more. Permeable connections for lateral flow between recharge and discharge areas may also be provided by faulting, although in some cases other mechanisms may be effective (e.g. carbonate and clastic formations and brecciated contact zones). In the extreme, hydrothermal circulation through and above cooling plutons appears to have occurred at depths of up to 10 km through fractures created by differential thermal expansion and thermal fluid pressurizing (Norton and Taylor, 1979; Stefansson and Bjornsson, 1982).

Pressure gradients for moving fluid through HCS are provided by a combination of thermally induced density differences and elevation differences between recharge and discharge areas. Density differences producing pressure differences of tens to hundreds of bars can drive HCS associated with shallow magmatic heat sources. In regional conductive environments, elevation differences appear to be required to initiate convective flow that in turn sets up density differences and additional pressure gradients.

Rates of throughflow under natural conditions are usually limited by overall differences in pressure or head and by rock permeability rather than by recharge potential. Typical throughflow rates for HCS in various areas are listed below.

1. Dixie Valley, NV - 5 kg/s
2. Beowawe, NV - 6 kg/s
3. Grass Valley, NV - 10 kg/s
4. Valles Caldera, NM - 20 kg/s
5. Steamboat Springs, NV - 60 kg/s
6. Long Valley, CA - 250 kg/s
7. Wairakei, NZ - 400 kg/s
8. Yellowstone Caldera, WY - 3000 kg/s

In general, flow rates for HCS with magmatic heat sources (4 - 8) exceed those for HCS in regional conductive environments (1 - 3). Although the magnitude and perhaps the duration of this natural flow provide some indication of thermal and hydrologic conditions at depth, they are by no means definitive. For example, low flow rates may result from low permeability at shallow depths in discharge areas below which high permeability reservoirs may exist.

For high-temperature HCS, depths of fluid curculation where maximum temperatures are attained are generally too great to be drilled directly. Consequently, exploitable reservoirs in such systems (and in many lower-temperature systems) must be fed or charged by upward flow and subsequent lateral leakage of thermal water. In some cases upward flow may be restricted to narrow fault zones and no lateral leakage may have occurred. In HCS where laterally extensive reservoirs do exist, fluid production during development can be expected to induce changes in the natural rate of inflow and outflow that may singificantly affect reservoir response.

Reservoirs in the Natural State The general features of the initial fluid states within the discharge portions of active HCS can be discussed with reference to the basic model suggested by Donaldson and Grant (1981) and shown in figure 1. The reservoir is pictured as a vertical cylinder containing hot fluid and surrounded by relatively cold water. Under undisturbed conditions, there is an upflow of fluid from below, M_b, and a discharge of fluid at the surface, M_f. Hence, the natural state is dynamic, not static; this upflow is largely responsible for the observed predevelopment pressures, temperatures, and enthalpies. For high-enthalpy inflow, the upflowing fluid boils when its pressure falls to the saturation pressure for its temperature and a zone of two-phase flow extends from that depth to a near-surface zone involving mixing with cold groundwater.

The boundaries of each zone are idealized here. In the field they are diffuse and irregular. Inflow at the base may be locallized along major fault structures, some of which may continue to transmit fluid to the surface. At the margins of the field there usually are intrusions of cold water and outflows of hot water, so that the temperature pattern may be quite disturbed. The single-phase/two-phase boundary can also show considerable uneveness that tends to follow distributions of temperature and gas content. The upper boundary where cold water overlies hotter fluid and where cold water may enter the reservoir under production is often the least regular. Within this zone, large variations in temperature near conduits carrying hot water and steam to the surface would be expected.

Most of these boundary irregularities are caused by variability in the permeability of the rock. In general, the spatial scale of these variations is usually sufficiently small that field-scale computations of initial conditions and effects of production can be accomplished by treating the rock as more uniformly permeable and the boundaries as smooth and regular. Permeability variations cannot, however, be neglected in more detailed exploration and development activities.

The general distributions of pressure and temperature with depth in the basic model are shown in figure 2. The pressure distribution OMA is nearly linear and super-hydrostatic, as would be required to drive the upflow to the surface. Gradients approximately 110 percent of hydrostatic have been measured in several New Zealand fields and in the Redondo Creek field in the Valles caldera (Grant, 1979a; Grant and Garg, 1981). Alternatively, in the case of the Redondo Creek reservoir, the measured pressure gradient of 79 bars/km (0.348 psi/ft) could be considered to correspond to a hydrostatic gradient for water at 260 °C so that fluid upflow is implied in regions where temperature exceeds 260 °C. The temperature distribution in figure 2 is isothermal along CD, and above the depth of boiling, B, temperatures following a boiling point-for-depth curve along OC. Localized effects of cool, near-surface groundwater on these profiles are indicated by the linear portion along OC (commonly much more irregular).

The pressure distribution in the two-phase zone OM remains superhydrostatic as long as both phases flow upward. This is the essential criterion for a liquid -dominated HCS. Pre-exploitation conditions at Wairakei matched those in figure 2, with the boiling level approximately 400 m below land surface and nearly uniform reservoir temperatures near 250 °C. Surficial discharge included both hot springs and fumaroles, with a total outflow of about 400 kg/s (Donaldson and Grant, 1981).

Terms like liquid-dominated and vapor-dominated HCS are actually overly restrictive. Many HCS contain both liquid- and vapor-dominated zones and transitions from one type to another may occur during the lifetime of the system. For example, Grant (1979a) discusses a model that accounts for different physical states within many HCS in mountainous terrain, such as occur in the Valles caldera, Lassen Volcanic National Park, and the Tongonan thermal area in the Philippines. As shown in figure 3, within the upflow region where boiling occurs there is a separation of phases. Liquid flows laterally to discharge as chloride hot springs at lower elevations, and steam rises through a parasitic vapor-dominated zone to discharge as fumaroles or steam-heated springs at higher elevations. Corresponding pressure distributions show liquid-dominated conditions in the region of choride spring discharge and beneath the vapor zone. Vapor-static conditions in the steam zone are overlain by a near-surface condensate layer where pressures are hydrostatic. Note that pressures in and below the two-phase zone are controlled by the elevation of the chloride springs and appear subhydrostatic with respect to elevations of the steam-heated springs.

Unpublished numerical simulations indicate that the flow system pictured in figure 3a could develop from an initial state involving high-enthalpy, liquid-dominated upflow to the surface. Necessary conditions appear to be (1) opening of an outlet for liquid at lower elevations and (2) low permeability near-surface rocks. Condition (2) is analogous to a low-permeability caprock above the two-phase reservoir and is required to prevent liquid from recharging and quenching the reservoir after the low-elevation outlet is created. Based on data for the Valles system (Union Oil Co., 1978) and the Tongonan system (Whittome and Smith, 1979), these conditions appear reasonable.

Additional variations on the basic model in figure 1 make it more applicable to other geothermal fields. The enthalpy and flow rate of the base inflow are critical parameters in this regard. For low enthalpy inflow, or higher enthalpy inflow at low flow rates, boiling may only occur near the surface or be non-existent. Such systems, as occuring at Raft River, Idaho, and Klamath Falls, Oregon, have been termed low- or intermediate-temperature systems. With high enthalpy inflow, the reservoir can be two-phase down to exploitable drilling depths. In addition, the presence of significant concentrations of non-condensible gas lowers the range of reservoir temperatures for which boiling occurs and so tends to increase the depth of boiling (Pritchett and others, 1981). Such conditions exist in New Zealand fields at Broadlands (boiling to 1.5 km) and Ngawha (boiling to 4 km).

Hydrothermal convection systems in which extensive vapor-dominated conditions prevail throughout the drillable depths and minimal surficial fluid discharge is observed fit the usual description of a vapor-dominated reservoir. As observed at The Geysers in California, Larderello in Italy, and Kawah Kamojang in Indonesia, circulation of steam and slightly mobile liquid occurs within the reservoir in the natural state; giving rise to relatively large heat flow at the surface.

Conditions leading to the development of vapor-dominated reservoirs are conjectural. White, Muffler, and Truesdell (1971) suggest evolution from an early hot-water stage as net discharge exceeds recharge and steam boils from a declining water table. Pruess and Truesdell (1980) attempted to numerically model this type of development by applying a potent heat source (190 HFU) to the base of a cylinder whose upper boundary was held at 1 bar and 10°C. During a transient period lasting for 93,000 years, convection under all-liquid conditions gave way to two-phase conditions beneath a low-permeability caprock layer. The establishment of vapor-dominated conditions, however, could not be determined from the results presented. Alternatively, the model presented in figure 3 suggests that vapor-dominated reservoirs could evolve in HCS with lateral outflow of liquid and upward migration of steam. Once a vapor zone is established, chemical deposition could act to restrict upflow, while continued lateral outflow induces boiling and expansion of the vapor zone.

<u>Reservoir Exploitation</u> The response of a geothermal reservoir in an active HCS to exploitation involves changes in mass and energy storage within the reservoir and motion of boundaries connecting the reservoir with the rest of the system. The basic model (figure 1) provides a useful reference for discussion of these effects.

The initial response to production is a decline in reservoir pressure in the production zone. With time pressure declines spread horizontally and vertically. Pressure declines can be expected to extend downward roughly twice the depth of the production zone (Grant, 1977a). In figure 1, induced inflow of cooler water from the sides of the reservoir is indicated by the inward bow of the dashed lines along the sides of the reservoir. An increase in base inflow, M_b, is not indicated in figure 1 because in most high-temerature HCS pressure drops accompanying production remain small compared with total pressure differences driving the throughflow. In practice, it is difficult to distinguish between recharge induced from cooler regions surrounding the reservoir and any enhancement of throughflow since both reach the production zone as hot water (until cold-water breakthrough occurs).

Reductions in reservoir pressure tend to cause reductions in the surface discharge, M_f, resulting in an increase in the net inflow of thermal water to the reservoir ($M_b - M_f$). The effects of this increase in inflow on reservoir productivity (discharge/

pressure decline) depend on the relative magnitude of M_b compared with well discharge, M_w. Data listed previously for throughflow rates in HCS indicate that in most cases $M_w >> M_b$. However, if a large fraction of the well discharge is injected into the reservoir, capturing the natural throughflow by reducing M_f to zero could significantly increase reservoir productivity.

Where a near-surface zone of cold groundwater overlies the reservoir, the potential exists for cold water to be drawn downward into production zones. Differences in the extent and timing of this cold-water fingering at various developed fields suggest that it may be more important for reservoirs in sedimentary rocks (like at Cerro Prieto, Mexico) than for reservoirs in fractured rocks (like Wairakei) (Grant and others, 1981), although other factors may be involved.

Motion of the cold side boundaries occurs as water from surrounding regions pushes hot water toward the production area and mines heat from the reservoir rocks. The rate of inward movement is considerably slower than the actual fluid velocity; in uniformly permeable rocks, tens to hundreds of years may be required for cold water to reach production zones (Robinson, 1977). In some fractured reservoirs, more rapid quenching of wells along high-permeability channels can occur (Horne, 1981).

The magnitude of induced recharge depends primarily on reservoir pressure declines and permeabilities in the surrounding region. In general, such permeabilities are much lower than reservoir values - in vapor-dominated reservoirs, the former may be effectively zero. At Wairakei, reservoir permeabilities on the order of 100 md (10^{-9} cm^2) exist within a region of permeability on the order of 10 md (Fradkin, Sorey, and McNabb, 1981; Mercer and Faust, 1979). Models of the response of the Wairakei reservoir to production and repeat gravity measurements (Hunt, 1977) indicate that with approximately 30 bars of reservoir pressure decline the rate of induced recharge is nearly equal to well discharge.

Within the exploited reservoir, pressure declines induce changes in fluid properties. Production from single-phase zones (liquid or steam) causes a decrease in fluid density, ρ_f, and a decrease in rock porosity, ϕ

The associated change in fluid mass per unit reservoir volume is given by

$$\Delta M = \rho_f \, \phi (c_r + c_f) \, \Delta P$$

where c_r = rock compressibility $(\phi^{-1} \partial\phi / \partial P)$ and c_f = fluid compressibility $(\rho_f^{-1} \partial \rho_f / \partial P)$. Values of c_r for reservoirs in porous rocks are roughly the same magnitude as values of c_f for liquid but less than c_f for steam. For reservoirs in many fractured rocks, $c_r << c_f$. The product of ΔM times reservoir volume gives an estimate of reservoir productivity in the absence of induced recharge.

The speed of transmission of pressure changes induced by production is also an important aspect of reservoir response. It is proportional to the hydraulic diffusivity, D, given under single-phase conditions by

$$D = \frac{k}{\phi \, \mu_f (c_r + c_f)}$$

where μ_f = fluid dynamic viscosity. From observations of the time, t, required for measurable pressure changes to occur at a distance L from the production zone, the following relations can be derived for D:

$$D \cong \frac{L^2}{t} \quad \text{(radial flow)}$$

$$D \cong \frac{L^2}{4t} \quad \text{(linear flow)}$$

Values of D for liquid reservoirs are relatively high because $(C_r + C_f)$ is relatively low. Under such conditions, pressure changes can propogate over distances of several kilometers in periods of months.

Production from two-phase reservoirs induces boiling and accompanying temperature declines as heat is removed from the rock to vaporize liquid. The change in fluid mass per unit of reservoir volume is then given approximately by

$$\Delta M = \rho_s \, \phi \, c_2 \Delta P$$

where ρ_s = steam density and c_2 = effective compressibility due to phase change. The expression derived by Grant and Sorey (1979) for c_2 as a function of latent heat of water and heat capacities of rock and water indicates that c_2 is 100 to 1,000 times larger than values of compressibility for steam and liquid, respectively. Thus, for the same P, changes in mass per unit reservoir volume are roughly 100 times larger under two-phase conditions than under single-phase conditions. This difference tends to result in larger reservoir productivity with boiling than under single-phase conditions. However, under two-phase conditions with both phases mobile, low relative permeabilities tend to force large pressure and temperature drops to maintain production. Under vapor-dominated conditions, production is maintained by boiling without the negative effects of low relative permeabilities.

Values of diffusivity under two-phase conditions are given by

$$D = \frac{k}{\phi \mu_2 c_2}$$

where μ_2 = total dynamic viscosity of two-phase flow (Grant and Sorey, 1979). Pressure transmission tends to be much slower than under single-phase conditions because $c_2 >> c_f$. This tends to minimize interference between wells but also restricts the inducement of recharge from surrounding regions.

In HCS for which the basic model (figure 1) applies, optimum reservoir response may be realized if well production from the deeper single-phase region induces expansion of the overlying two-phase zone with accompanying downward drainage of liquid. The Wairakei reservoir appears to have responded in this manner. With an initial boiling level at a depth of about 400 m, most production bores were drillled to depths of 0.5 - 1.5 km. During over 20 years of production at an average rate near 2000 kg/s, most of the discharge has entered the wells from liquid-filled portions of the reservoir. This is evidenced by relatively uniform discharge enthalpy and rapid tranmission of pressure changes across the reservoir (Fradkin, Sorey, and McNabb, 1981; Bolton, 1970). Measurements in shallow bores indicate that the overlying two-phase zone has expanded laterally and vertically and become drier. The boiling level has fallen about 200 m, implying drainage of liquid in amounts sufficient to supply approximately 10 years of production (McNabb, 1975). The resultant rate of reservoir pressure decline with discharge for the Wairakei reservoir is lower by a factor of about 50 than would be expected in the absence of two-phase conditions (Pritchett, Rice, and Garg, 1979). In effect, the reservoir response appears much more compressible than under single-phase conditions.

Concurrent with changes in storage within the two-phase zone at Wairakei, pressure declines in the production zone have been transmitted out to the sides of the field where recharge of colder water has been induced. The relative contributions of induced recharge and storage changes (drainage) have been delineated from repeat gravity surveys (Hunt, 1977) and modeling simulations of the reservoir pressure history (Mercer and Faust, 1979; Prittchet, Rice, and Garg, 1980; Fradkin, Sorey, and McNabb, 1981). These studies show that drainage effects were most significant in the early stages of field development, and that recharge has increased as reservoir pressures have declined to the point where over 90 percent of current production is supplied by recharge. As noted previously, lateral boundaries are not likely to be uniformly permeable and induced recharge would not be uniformly distributed around the reservoir. Such is the case at Wairakei, as evidenced by the variability in pressure response in cold peripheral wells and pressure drawdowns in wells in the adjacent Tauhara field (Grant, 1978).

To a large extent, the near-optimal response of the Wairakei reservoir results from its relatively high permeability-thickness (kh=10^5 md-m). In fields where kh is significantly lower than at Wairakei, or in fields with similar kh produced at significantly greater rates, two-phase conditions may invade production zones that were initially single-phase. For example, at the Redondo Creek field in the Valles caldera, where kh values appear to be near 2 x 10^3 md-m, well tests and modeling simulations predict extensive development of two-phase conditions in the reservior accompanying production at rates near 400 kg/s for 50 MW_e (Bodvarsson and others, 1980). Corresponding decreases in total dynamic viscosity with two-phase flow at Redondo Creek would necessitate large pressure drops to maintain a constant production rate, although this effect can be partially offset by reinjection and by the need for less total production to supply 50 MW_e as the field dries out. The two-phase Ohaki reservoir at Broadlands, NZ, on the other hand has kh = 2 x 10^4 md-m and may be able to sustain production at 100 MW_e if recharge from surrounding liquid-filled regions can be induced (Grant, 1977b; Donaldson and Grant, 1981).

Effects on Surface Features Surface dischages of thermal fluid in active HCS vary considerably in their physical and chemical characteristics, and studies of these characteristics have provided much of our current understanding of HCS (White, 1970; Truesdell, 1976; Ellis and Mahon, 1977). The potential for alteration of the natural rate of fluid discharge during reservoir exploitation was noted previously. Because this effect may significantly influence the development of geothermal resources (from both physical and regulatory standpoints), this issue is discussed in more detail here.

The nature and magnitude of changes in flow rate, temperature, and chemistry of hot springs and fumaroles accompanying reservoir exploitation depend in part on how such discharge features are connected with the reservoir and the rest of the HCS. In the simplest case (1), a hot spring contains fluid of chemical composition similar to that of the reservoir fluid and is fed simply by upflow along one or more conduits from the reservoir to the surface. Hot springs with high chloride, neutral pH waters typify this situation. A slight variation (2) involves the mixing of reservoir fluid with cooler, more dilute water within the upflow conduit. In some cases (3), warm springs represent shallow groundwater heated by conduction from an underlying convective anomaly, and therefore contain no water from a geothermal reservoir (Goff and others, 1982). In other cases (4) shallow groundwater may be heated by addition of steam from a deeper vapor-dominated zone, producing low-chloride, acid-sulphate hot springs. Steam discharge in fumaroles (5) commonly occurs along with steam-heated springs and both features may be indicative of vapor-dominated reservoirs or of parasitic vapor zones in active HCS (figure 3).

In the natural state, pressure gradients within and above the reservoir drive thermal fluid toward the surface. Fluid production during exploitation causes a reduction in reservoir pressure and in the vertical pressure gradient. As a consequence, the rate and temperature of liquid discharging at the surface (as in cases 1 and 2) will decrease, although the time scale over which such changes occur varies with the properties of rock and fluid in the upflow conduits. Within zones of two-phase flow or vapor flow, however, vertical pressure gradients may exceed vapor-static conditions during exploitation and increased steam mobility can result in increased upflow of steam (Grant, 1979b). Thus, in cases (4) and (5) reservoir production may result in additional fumarolic activity and temperature increases in steam-heated springs. Discharge rates and temperatures of warm springs (case 3) should be the least affected by reservoir exploitation.

Most of the effects described above have been observed at Wairakei during its 30-year production history. Significant fluid withdrawal began around 1953 and reductions in natural discharge were observed by 1954 (Grant, 1979b). By 1962 major reductions in the discharge of high-chloride hot springs in Geyser Valley, located 1 km northeast of the production field, were found (Glover, 1977). By 1968, the activity in Geyser Valley had changed from mainly flowing springs and active geysers to steam-heated non-flowing pools, mudpots, and fumaroles. During the same period (1955-1968), rates of heat and mass flow from thermal areas to the west, north, and south of the production field increased due to the expansion of vapor-dominated conditions above the reservoir in those areas (Allis, 1980). By 1979, total mass outflow from thermal areas had declined by about 60 percent, from 440 kg/s to 180 kg/s. A change from mainly liquid discharge to mainly steam discharge, however, resulted in an increase in natural heat discharge from 400 MW to 600 MW. The estimated net heat flow loss from the field of 700 MW_t in 1979 (including production and recharge) would produce an average thermal rundown of the two-phase zone of 1.5 °C/yr, and further reduction in the mass outflow from thermal areas (Allis, 1980).

Changes in surficial discharge of thermal fluid due to reservoir exploitation in other active HCS can be expected to be of a similar nature to those observed at Wairakei. However, the timing and magnitude of such changes depend on several factors. These include reservoir and caprock permeability and diffusivity, production rates and depths of producing intervals, distances from production zones to discharge areas, and locations of reinjection wells. Significant reductions or cessation of natural liquid discharge following reservoir testing or development has been observed in geothermal areas in Mexico (Cerro Prieto), Japan (Otake), Iceland (Hveragerdi), and the United States (Beowawe and Steamboat Springs, Nevada). However, quantitative analyses that account for the factors noted above are generally lacking except for preliminary results for the Rotorua-Whakarewarew field in New Zealand (Donaldson, 1980) and the Valles Caldera (unpublished reports by Union Oil Company and Geotrans, Inc.).

It should be noted that some amount of reservoir pressure drawdown will accompany fluid production, even at rates below that of the natural throughflow. This drawdown tends to reduce liquid outflow, as does any increase in boiling in the two-phase zone (which reduces liquid relative permeability). Reinjection of waste liquid offers possibilities for reservoir pressure maintainance and increased recovery of stored thermal energy. Short-term effects of reinjection may include increased reservoir production, increased natural discharge of liquid, and reduction in natural discharge of steam. However, long term effects may be detrimental to the borefield production enthalpy in fields like Wairakei where induced hot liquid inflow from depth supplies most of the heat extracted from the reservoir.

<u>Summary</u> Consideration of geothermal reservoirs as existing within larger hydrothermal convection systems goes a long way toward explaining the general thermal, chemical, and hydrologic observations made in explored geothermal fields. Information on the characteristics of HCS can quide exploration for geothermal reservoirs and the analysis of energy recovery from such reservoirs. In addition to indications provided by surficial discharges of thermal fluid, regions of upflow of thermal water delineated from temperature measurements in shallow or intermediate-depth wells may offer attractive targets for deeper drilling. The response of developed fields in active HCS has also been shown to depend on the nature of the boundaries between the reservoir and the rest of the system. The consequences of changes in flow across these boundaries during development may vary for different fields and additional experience and analysis is required before predictions can confidently be made for undeveloped fields.

Studies of HCS offer unique opportunities for the blending of information from a variety of disciplines and specialties. With such input, geothermal reservoir engineering can evolve from more narrowly defined limits imposed by previous experiences restricted to fields like petroleum engineering and groundwater hydrology. In addition, knowledge gained in the study of geothermal reservoirs should be applicable to other efforts such as volcanic hazards assessment and mineral recovery from ore deposits associated with fossil hydrothermal systems and submarine hydrothermal systems.

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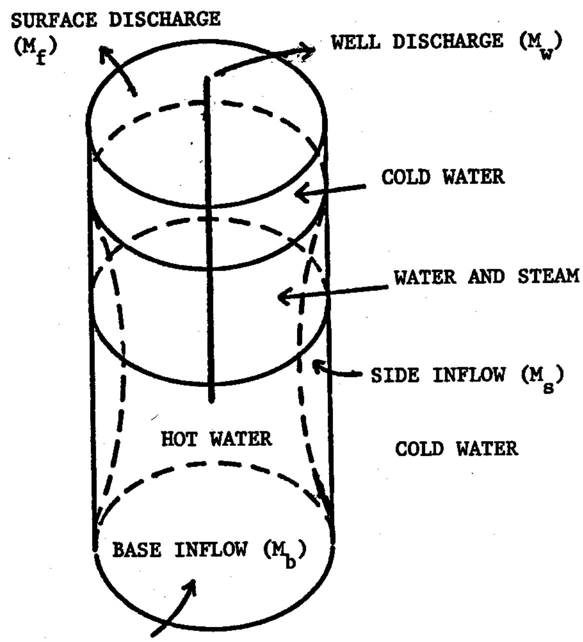

Figure 1. Basic model of a geothermal reser-reservoir in the discharge portion of an active hydrothermal convection system (adapted from Donaldson and Grant, 1981).

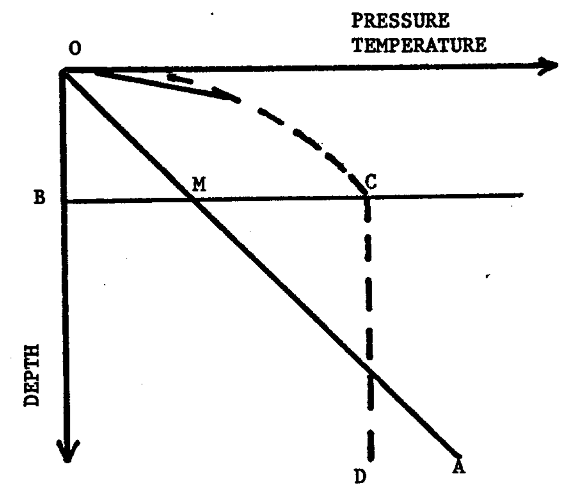

Figure 2. Generalized distributions of temperature and pressure with depth in the basic model of figure 1 (adapted from McNabb, 1975)

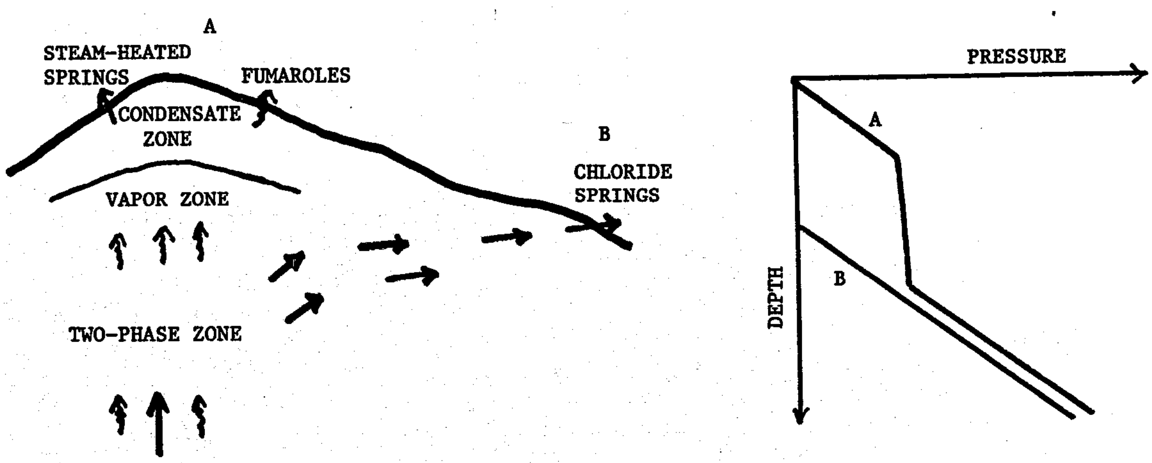

Figure 3. Conceptual model of hydrothermal convection systems in regions of high relief in which phase separation occurs in the upflow zone, producing a vapor-dominated zone above and a zone of lateral outflow of liquid. Generalized pressure distributions with depth in region A (upflow zone) and region B (outlfow zone) are shown at right (adapted from Grant, 1979a).