HEAT AND MASS CIRCULATION IN GEOTHERMAL SYSTEMS

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INTRODUCTION

It is likely that hot water flowing out of the ground has been an attraction to man ever since he first stumbled across it. In Europe and Japan such water has been used for bathing for many centuries. In other parts of the world, it has been used for cooking and as a stimulus for tourism.

With the advent of mechanized drilling, we are no longer restrained to use only the fluid that escapes at the surface. We now have an indication of the extent of the resource and ready access to it. Information from the many wells drilled has also led to our greater understanding of the subsurface heat and mass movement. As the exploitation of these resources has intensified, demands on this knowledge have increased. Current research is directed to many aspects of the overall problem.

To engineers developing a geothermal reservoir, potential users of the energy, and financiers or entrepreneurs carrying the risks of development, the primary requirement is a good estimate of the energy reserve of that reservoir, i.e. the identified geothermal energy that can be extracted legally today at a cost competitive with other energy sources (Nathenson & Muffler 1975, Muffler & Cataldi 1978). These people also want to know the rate at which it would be best to extract this energy, the time span over which they might expect to extract it at that rate, and the changes that may occur in the reservoir due to the exploitation.

Given even relatively limited information about the reservoir, such as temperature and area, it is usually possible to estimate the in situ stored heat above some appropriate depth. The problem is estimating the fraction of this heat that is recoverable. For their recent assessments of the geothermal resources of the United States, White & Williams (1975) and Muffler (1978) used a model of heat extraction based on intergranular flow, or the sweep process, put forward...
by Bodvarsson (1974) and Nathenson (1975) to estimate the fraction of the stored heat that might be mined. Donaldson & Grant (1978), in contrast, used the similar nature of New Zealand fields and their knowledge of the Wairakei, Broadlands, and Kawerau reservoirs in their estimate of the electric power production potential of the New Zealand system.

The fraction of the energy that can be extracted must depend, however, on many factors: the nature of the reservoir, i.e. its size, structure, degree of fracturing, and temperature variation; the nature of the fluid, i.e. water, water and steam, large or small quantities of noncondensable gas; and the way in which the field is operated, i.e. the well distribution, reinjection strategy, and rate of withdrawal. This review looks at a selection of the work that has expanded our knowledge of the nature of geothermal systems and reservoirs, the effects of the different fluid states, and the basic consequences of exploitation.

Discussion is restricted to one type of geothermal system, the hydrothermal system, in which heat is transported primarily by the circulation of water and/or steam. There will be no discussion of naturally conductive "hot-dry-rock" systems, or of magma systems. Nor will there be discussion of geopressed systems, even though these are water charged.

GEOTHERMAL SYSTEMS

Before we can discuss the effects of fluid or heat withdrawal from a geothermal reservoir, we must have a conceptual picture of heat and mass circulation in the system that sustains that reservoir. Unfortunately, in general, only indirect information about the system at depth is available. It is thus necessary to speculate to some extent on both the flow paths for the fluid and the source of the heat.

Perhaps the most significant information available is that, in general, some 90–95% of the water in geothermal systems is of meteoric origin. This was first suggested early this century by V. Knebel (1906) and Thorkelsson (1910) (references from Einarsson 1942) but was not confirmed until oxygen (O\textsuperscript{18}/O\textsuperscript{16}) and hydrogen (D/H) isotope ratios of geothermal waters were measured (White 1957a,b, 1961). This meteoric water is believed to circulate to depths of 2–6 km.

Without evidence indicating contact with magma, it is difficult to prove magmatic material as the primary source of heat in these systems, even in active volcanic areas. Certainly there are many geothermal systems in the world where the water is hot and there is no obvious volcanism. One well-known example of such a system is the Carpathian basin in Europe (Boldizsar 1975, Boldizsar & Korim 1975). In this and similar systems, normal heat flow from the Earth's core is sufficient to maintain temperature. These systems would thus presumably have stabilized to a steady state.
The more intensive geothermal areas are commonly associated with volcanism, often recent and active. Magmatic intrusions are suggested as one potential source of heat (Elder 1965, 1966). The cooling of such intrusions would supply additional local heat. Geothermal systems associated with such intrusions will, however, be transient.

For many fields, such as Wairakei, New Zealand, for which there is evidence suggesting a continuous existence over a period of 100,000 years or more (Grindley 1965), such intrusions would need to be extremely large or intermittently recharged. This time scale has inspired many researchers to consider such systems as steady, rather than transient.

Models of geothermal systems are of two types, transient and steady. To those interested in exploitation there is probably little difference. The time scale of any exploitation is so short in comparison with any natural development that the natural state of the system may always be taken as pseudo-steady.

### Transient Models

Magmatic intrusion as a source of heat is by no means a new idea. Ingersoll & Zobel (1913), for example, considered the conductive problem. Convective flows stimulated by intrusion have, however, received detailed study only relatively recently. All of these studies, in effect, assume that the hot intrusion is suddenly injected into a deep, cold, water-saturated zone.

Norton (1977, 1978) and Norton & Knight (1977) recognize that both regional tectonic activity and the entry of the intrusion would lead to fracturing in the area, and thereby provide significant paths for convective fluid flow above and around the intrusion. The detailed analyses of Norton & Knight (1977) show that convective heat flux above the pluton could maximize at eight times the heat flux attained by conduction alone. Norton (1978) shows that fluid flow through the pluton itself accelerates cooling. He associates this flow with thermally induced stress cracking that would occur as the pluton cools.

In an independent, complementary study, Cathles (1977) computed the heat and mass flow developed in a similar system. Due to differences in scale and parameters, the peak (conductive + convective) surface heat flux above the pluton is only 3.6 times the peak flux that would be attained without convection. Cathles (1977) allowed the fluid to change phase as it circulated. He did not permit water and steam to coexist.

These transient models of Norton (1978) and Cathles (1977), reviewed in fuller detail by Garg & Kassoy (1981), illustrate some of the problems that arise in considering the source of heat as an integral part of our system. Nonetheless, these analyses have been used by Smith & Shaw (1978) in their assessment of the geothermal energy available in the United States from igneous-related geothermal sources.
Fluid behavior in the immediate vicinity of an intrusive is also receiving attention. Cheng (1976) discussed single-phase boundary layer flows along a dike and the cooling that these flows might induce. This work has recently been extended (Cheng & Verma 1980) to allow boiling in the boundary layer. The analysis depends critically on the assumptions that two-phase fluid does not exist in the boundary layer, and that the vapor-liquid interface is smooth.

The cooling of a lava sheet subject to direct rainfall has been studied recently by Shaw et al (1977) and Peck et al (1977). Here there is almost instantaneous contact of hot rock and water, both at the surface and internally, via a complex of contraction joints and cracks. While such sheets may be fully cooled in many cases before they are buried, the flow channels may be important in future geothermal systems. For example, flow channels at the surface of rhyolite domes are important features in the Whakarewarewa-Rotorua system in New Zealand (Donaldson & Grant 1981b).

**Steady Models**

Steady-state models of geothermal systems have developed along two lines: pipe models, in which it is assumed that the fluid flow is restrained to a network of channels, and porous medium models, in which the overall structural system is approximated as a homogeneous (although not necessarily isotropic) permeable zone.

**PIPE MODELS**  
Pipe systems have been proposed and described for several geothermal systems. The first detailed was that by Einarsson (1942) for the system feeding the hot springs in west Iceland. Einarsson (1942) suggested a deep circulation of meteoric water through cracks and fractures in otherwise virtually impermeable tertiary plateau basalts. Einarsson’s work was extended and elaborated on by Bodvarsson (1961), in particular for the system associated with the more intensive activity found in central Iceland.

Pipe systems have also been proposed and discussed for US systems (White 1957a, 1961). White’s conceptualization of a system is illustrated in Figure 1. In this model the water circulates to depths of 2 to 6 km where it receives heat from the hot rock, which is, in turn, heated by magma at greater depth. The heated fluid, being of lower density than the cold fluid, is then driven upwards through the available cracks and fractures by the difference in cold water/hot water head.

This model, being inherently simple, has not stimulated a great deal of general heat and mass transfer research. Both Elder (1966) and Donaldson (1970) used it to represent deep circulation, but were primarily interested in porous medium flows in and around the hot upflow column. Donaldson (1968) also used it for the deep circulation in his two-phase model of Wairakei.

Probably the most interesting recent use of a pipe model is by Goyal &
Kassoy (1977, 1980) and Goyal (1978) in their conceptualization of the East Mesa system in the Imperial Valley, California. Although they are discussing the reservoir rather than the system, they suggest that the hot fluid flows up a single fault from an extensive basement fracture system in which the heating takes place. In the reservoir, this fluid is forced to move out horizontally by an overlying capping layer, vertical mixing in the reservoir being restrained by impermeable strata intermixed with the permeable beds.

POROUS MEDIUM MODELS It is now acknowledged that most geothermal systems and reservoirs are well fractured. To simulate these systems many researchers have used models featuring homogeneous permeable structures.
For steady and background system studies, such simulations appear to have been adequate. In reservoir studies, transient effects and two-phase flows may be different in the fractured and porous materials.

The form of a porous medium model depends on the basic assumptions and requirements of the modeler. In the early period of research, the models were idealized and the studies aimed at the determination of the conditions of the onset of convection (following Horton & Rogers 1945, Lapwood 1948). Later the modelers commenced to look at the form of the convective circulation. Wooding (1957, 1958) used perturbation methods to include the nonlinear terms; Donaldson (1962), on the other hand, used relaxation (finite difference) methods. Wooding (1963) and McNabb (1965) worked with the hot column alone, treating it as a jet flow.

Since that time an extensive literature on steady convection in porous media has developed, both for idealized and simulated real systems. This literature has been fully reviewed by Combarnous & Bories (1975), Cheng (1978), and Garg & Kassoy (1981).

If a pipe system is assumed it is possible to visualize the surface manifestation of a geothermal system at almost any location. In contrast, the convective circulation in a porous medium model defines a horizontal spacing for any upwelling (hot) flows. Thus, it might be assumed that the pipe model is the more likely in many situations. Wooding (1978), allowing for the difference in vertical and horizontal structuring by using anisotropic permeability, was able to show, however, that using a permeable bed depth of 3 km, he could produce upflows in his porous medium model at a spacing of about 11 km. This spacing is in good agreement with the spacing of the high heat anomalies in the Taupo Volcanic Zone in New Zealand. Whether systems are pipe or porous medium controlled is thus still an open question.

**UNDISTURBED GEOTHERMAL RESERVOIRS**

For this discussion we call the hot section of the geothermal system that is tapped and affected by the wells once exploitation is underway the *geothermal reservoir*. In general this is the hot zone at the top of the upflow column, although in capped systems it may be at some depth. It is usually assumed to extend horizontally to the boundaries of the anomaly at approximately the level of exploitation. The depth of the reservoir may be defined by the structures or by some analysis of the extent of the effects of exploitation.

Although each reservoir is unique, there are enough similarities between them to be able to group them. Donaldson & Grant (1981a) put forward definitions that depend on the state of the reservoir fluid tapped by the wells once exploitation is underway and the vertical pressure gradient in the reservoir. If the pressure gradient is close to hydrostatic, Donaldson & Grant
(1981a) use the term *liquid-dominated*, as the liquid water must be the connective phase even though water and steam may coexist in some sections of the reservoir. In contrast, if the pressure gradient is close to vapo-static, they use the term *vapor-dominated*, following White et al (1971). To date there appear to be no reservoirs with intermediate gradients.

Within the liquid-dominated class, Donaldson & Grant (1981a) define 3 subclasses: *warm water*, reservoirs in which no boiling will occur; *hot water*, in which boiling does or may occur, but within which the majority of wells are likely to tap liquid water, rather than water and steam; and *two-phase*, in which boiling also occurs, but at sufficient depth that the majority of wells will be tapping this boiling region. Typical examples of the three types might be Heber, Imperial Valley, California (warm water), Wairakei, New Zealand (hot water), and Broadlands, New Zealand (two-phase).

**Liquid-Dominated Reservoirs**

Many aspects of the behavior of a liquid-dominated geothermal reservoir can be studied with a relatively simple model (Donaldson 1978, Donaldson & Grant 1981a). This model, illustrated in Figure 2, basically consists of a column of hot fluid flowing up through the formations, surrounded by, and in
good hydrological contact with, cooler fluid. By adjustment of the temperature of the hot column at depth, the vertical permeability of the formations, the rate of upward flow, and (if necessary) the noncondensable gas content, the fluid may be made to flow up as water all the way to the surface or boil to various depths on the way up. It may thus be used to simulate warm water, hot water, and two-phase reservoirs. Structural stratification may be incorporated, although complete capping is only possible if there is no net upward mass flow.

Although this model has proved useful for some general studies—it has been used, for example, for studies of two-phase flow (Donaldson 1968) and to illustrate the effects of extraction of heat and mass from geothermal reservoirs (Donaldson & Grant 1981a)—it is of limited value for the study of particular reservoirs. These commonly have specific features that play an important role in the circulation of mass and heat within them.

We have already discussed the effect of the feed fracture, capping structure, and aquifer structural stratification on the fluid flow in the East Mesa reservoir, Imperial Valley, California (Goyal & Kassoy 1977, 1980, Goyal 1978). In that model the structure also affects the heat flow. As the heat can only escape by conduction through the capping layer, the cooling only extends to the deeper parts of the aquifer as we move out from the feeder fault. The temperature profiles thus are dome shaped.

In their study of the Heber reservoir in Imperial Valley, Tansev & Wasserman (1977) use a model similar to that used for East Mesa, in that it has a capping structure and some aquifer structural stratification. In their model, however, there is no central feed fracture and the overall vertical permeability of the aquifer is sufficient to allow circulation of the fluid. The water in the aquifer thus rises over the heated area in the center of the aquifer base, moves out beneath the capping structure, and sinks as it cools to return in along the base of the aquifer. The isotherms in this case are mushroom shaped. Thus, relatively small changes in structure can make significant differences in the natural flows in reservoirs.

A hydrological model of Cerro Prieto, Mexico, has been developed by Mercado (1975). This was established before production began and was based on temperature, pressure, enthalpy and flow measurements in wells, and the geothermochemistry of the discharged fluids. Mercado's (1975) model shows a horizontal and ascending movement of hot fluid from the eastern and central zones toward the west beneath a thick clay cap rock. It also suggests some recharge to the eastern section of the field, where the permeable reservoir structure extends to greater depth.

East Mesa and Cerro Prieto studies suggest both horizontal flow in the reservoir and some fracture control. Both of these are features of many reservoirs. Healy & Hochstein (1973) suggested that the El Tatio geothermal reservoir in Chile is a horizontal flow aquifer. Horizontal flow is also postu-
lated in the Rotorua section of the Whakarewarewa-Rotorua geothermal reservoir in New Zealand. A recent detailed analysis of all the available data suggests that some of the fluid coming up from depth in the Whakarewarewa area is being driven out to the north through channels and fractures in the outer layers of a shallow rhyolite dome. Other fractures and faults may carry more fluid out to the northeast (Donaldson & Grant 1981b).

Other factors that can play a role in reservoir form are the extent of the two-phase zone and the existence of noncondensable gases in the reservoir fluid. The coexistence of water and steam in a reservoir in its natural state is difficult to confirm, as any fluid withdrawal can affect the state of the fluid in the reservoir around a well. The early Wairakei data have, however, recently been analyzed by Grant & Horne (1980). This confirms a preexploitation boiling profile and indicates the variation of the upper and lower boundaries of this two-phase zone along a profile across the field.

The noncondensable gases, such as CO₂, found in significant quantities in many geothermal reservoirs, are now also being incorporated in the reservoir models (see, for example, Grant 1977). Sutton and McNabb (1977) showed that pressure-temperature data for the Broadlands geothermal reservoir, New Zealand, can be fitted very closely by the boiling curve of a CO₂-water mixture (4% CO₂ by weight). Boiling commences at a depth of about 1500 m at a temperature of 304°C and a pressure of 166 bars. Straus & Schubert (1979) found that the buoyancy of geothermal fluids depends critically on this CO₂, because of the large fluid volume changes that occur as this gas enters or leaves solution. They also showed that the thermal expansivity, compressibility, specific heat, and adiabatic temperature gradient of steam-water-CO₂ mixtures are significantly greater than those of liquid water—for some properties by as much as 5 orders of magnitude (Schubert & Straus 1981).

**Vapor-Dominated Reservoirs**

Although there were several earlier models of vapor-dominated geothermal reservoirs, there appears to have been no new general model proposed since that of White et al (1971). White et al (1971) suggest a reservoir containing coexisting steam and water, the water saturation being sufficiently low that the liquid phase is almost immobile and the steam maintains the vertical pressure gradient. This steam is assumed to move up through the reservoir to an overlying condensation layer. The small amount of mobile water flows back down. The existence of this system requires very low permeability in all surrounding structures. White et al (1971) therefore suggested that the reservoir developed from a more free-flowing liquid-dominated form through a process of self-sealing of the bounding flow channels, i.e. by chemical deposition in these channels. This process of development has recently been followed through in a numerical study by Pruess & Truesdell (1980).
Although the upflow of steam and water through porous beds has received only limited attention (Donaldson 1968, Sheu et al 1979), counter-current flow, as postulated for vapor-dominated reservoirs, has received considerable attention. Soldergeld & Turcotte (1977) studied this upflowing steam/downflowing water condition in a laboratory sandbox, while Herkelrath (1977), Schubert & Straus (1977, 1979, 1980), Straus & Schubert (1981), and Schubert et al (1980) have addressed various aspects of the problem theoretically. Schubert & Straus (1977) considered the convective nature of the flow and showed that convection occurs more readily in a porous medium containing saturated liquid by the phase-change instability than it would in a porous layer filled with liquid water by the Rayleigh-Bénard buoyancy-driven instability. They also indicate that phase-driven convection is concentrated toward the bottom of the porous layer, and that the cells are narrow in comparison to their depth.

Concern over the ability of a thick layer of liquid condensate to rest stably on top of the vapor-dominated reservoir stimulated the studies by Schubert & Straus (1980) and Schubert et al (1980). These studies show that the water layer will remain stable provided the permeability of the rock at the steam-water/liquid water interface does not exceed about 40 nm² (0.04 millidarcy). In their most recent paper, Straus & Schubert (1981) model vapor-dominated reservoirs similar to the Geysers in California and Kawah Kamojang, West Java, as one-dimensional flow systems in a porous medium saturated with water and steam. Temperature and pressure data for the Kamojang reservoir are best fitted by models with a net mass flow rate/thermal conductivity ratio near \( 2.5 \times 10^{-7} \) K s² m⁻³ and a vertical permeability/thermal conductivity ratio between \( 10^{-15} \) and \( 10^{-14} \) m s⁻³ K kg⁻¹. By analyzing the gas contents of discharging wells, Grant (1979) has deduced a water saturation of 35% for this reservoir.

EXPLOITED GEOTHERMAL RESERVOIRS

When water is extracted from wells tapping a warm water geothermal reservoir, the pressure transient propagates out from the well (or wells) in the same way as in any other liquid-saturated system. The theory is well documented in both the groundwater and petroleum literature for a wide range of system types and constraints (see, for example, Bear 1972, Earlougher 1977).

The heat extracted from such reservoirs is that contained in the water unless cooler water is brought into contact with the hot rock. Such a cool flow may be induced in two ways: by the pressure drop in the hot reservoir propagating out into the colder surrounding region, or by the direct injection of cool water into the reservoir.
In the former case, the reservoir boundaries will move in with time as the cool water extracts the heat from the boundary rock (Donaldson & Grant 1981a). Under ideal conditions, i.e. the reservoir being uniform in every respect, Donaldson & Grant (1981a) estimate that about 50% of the heat would be extracted from a reservoir 3 km deep and 4 km in diameter through wells tapping a zone about 1 km deep and 1 km in diameter before the cold front reached the outer wells. The rate of movement of a front has been computed and discussed by Bodvarsson (1974).

Reinjection of the cold fluid will establish a cooling front around each injection well (or group of injection wells). As this front expands it will “sweep” a proportion of the heat through to the extraction wells. For conceptual ideal reservoirs good results have been predicted (see, for example, Martin 1978), but as Horne (1981) points out in his analysis of geothermal reinjection in Japan, in many cases, reinjected water moves through fractures or fissures of extremely high permeability in the reservoir. In many reservoirs around the world, fluid has been found to move hundreds of meters in periods of only a few hours. Estimates of recoverability of energy from reservoirs in which short-circuiting possibly takes place thus may be markedly different from the ideal. Any boiling in a reservoir also introduces some differences in the response of that reservoir to exploitation.

**Fractured Reservoirs**

The first models of fractured reservoirs involved single fractures. In one approach (Bodvarsson 1974), a planar fracture was assumed to be continually recharged at one end with cool water. Classical heat-conduction theory then gives the heat theoretically extractable per unit fracture area under a minimum outflow temperature assumption. This single fracture system was extended to a multiple system by Nathenson (1975).

The Nathenson (1975) model was one of the models used by Muffler & Cataldi (1978) in their discussion of heat recovery factors for geothermal reservoirs. For an original rock temperature of 250°C, a recharge temperature of 40°C, a minimum outflow temperature of 162°C, a fracture spacing of 338 m, and a timescale of 25 yr, Muffler & Cataldi (1978) computed a theoretical heat recovery factor of just under 20%. They point out that in a real situation it could be much lower.

Pressure transient studies in systems with fractures are another method of learning more about local and total reservoir behavior under exploitation. Type curves for pressure transients in wells tapping single horizontal, vertical, and sloping planar fractures have been developed by Gringarten et al (1974), Gringarten & Ramey (1975), and Cinco-L. (1974), respectively.

On account of its initial performance and relationship to nearby wells in the
Travale-Radicondoli geothermal reservoir in Italy, Travale Well 22 has been extensively tested. These tests have been analyzed by Barelli et al (1975, 1978) and Atkinson et al (1978a). Although the short time data fit curves for a well intersecting a fully penetrating vertical fracture in a finite system, at a longer time (about 400 days) the system seems to reach steady flow. A new model, in which the well is intersected by a partially penetrating vertical fracture in a parallelepiped whose bottom side is a constant pressure boundary, was therefore proposed. The data appear to match the curve for a dimensionless formation thickness (reservoir thickness/half fracture length) of about 2.5 (Barelli et al 1978).

Multiple and randomly fractured beds of rock saturated with water may simulate real geothermal reservoirs more closely than the above idealized models. Such beds have recently been under study by Hunsbedt et al (1978). In a series of experiments a volume of such saturated rock has been heated to 260°C and pressurized to 55 bars and then swept of heat by passing through cold water. The results show that up to 175 kilo-Joule of heat per kilogram of rock mass could be extracted by this process (Hunsbedt et al 1978). The rock elements were not permeable.

Permeable, fractured systems have not been neglected. The pioneering isothermal studies for groundwater and petroleum reservoirs were carried out by Barenblatt & Zheltov (1960) and Warren & Root (1963). These analytic studies consider flow from primary porosity blocks to secondary porosity fissures. They do not, however, describe the flow within the blocks. The finite difference study by Kazemi (1969) and the analytic work of Boulton & Streltsova (1977) take this flow into account. In recent extensions of this work, Da Prat (1981 and Da Prat et al 1980) have obtained the longer time-pressure transients. It is thought that their results may be applicable to geothermal reservoirs. A fuller survey of the background literature is given by Evans (1981).

Nonisothermal flow in such fractured, permeable structures has been studied by Moench (1978), who developed a radial flow, finite difference model. This was later used to simulate pressure buildup data for a steam well in Larderello, Italy (Moench & Neri 1979). In that model the blocks were assumed to be impermeable, but capable of conducting heat to the fissures as they cooled because of vaporization of water to steam. In a recent study, Moench & Denlinger (1980) have revised the model to allow for steam transport and vaporization in the blocks.

In an effort to better understand the process of heat transfer within and around a single block, Pinder et al (1979) carried out a simple numerical experiment. They assumed a steam-saturated block to be suddenly surrounded with cool reinjection water. The simulations indicated that the liquid water propagated into the block and was heated to the original reservoir temperature.
A fractured, permeable model of the Wairakei geothermal reservoir was proposed by A. McNabb (unpublished report) in 1975. A main role for the fractures in this model was for the drainage of mobile water in the two-phase zone down to the liquid water/two-phase interface. This model has recently been shown to be the physical counterpart of the lumped parameter model of Fradkin (1981, Fradkin et al 1981).

**Reservoirs in Which Boiling Occurs**

Once boiling occurs within a geothermal reservoir, some changes in the behavior of the reservoir under withdrawal and injection are to be expected. In the hot water reservoir these changes are limited, because the pressure transients will still propagate out to the side boundaries through the liquid water, as was the case in the warm water reservoir.

These pressure transients will, however, also propagate in the vertical direction and hence, in time, reach the liquid water/two-phase interface in the hot water reservoir. At this interface the pressure and temperature are interrelated (at saturation conditions). Any drop in pressure there thus stimulates a drop in temperature and, hence, additional boiling as heat transfers from the rock to the fluid. The interface, therefore, will only drop slowly and tend to act as a pressure-stabilizing boundary (Donaldson & Grant 1981a). Any drop in the interface level will have two effects: the pressure will drop in the two-phase layer, and the vertical pressure gradient through that layer will be reduced. The drop in pressure will mean more boiling and the transfer of heat from the rock to the fluid. The drop in pressure gradient will reduce the upward water flow. With a sufficient drop in gradient the water flow will in fact reverse, i.e. it will commence to drain down toward the interface (Donaldson & Grant 1981a). These pressure gradient changes have been found in Wairakei, New Zealand (Grant & Horne 1980).

When production takes place from within the two-phase zone of the reservoir, withdrawn fluid is replaced by steam supplied by vaporizing some of the liquid water. The pressure drops only when cooling takes place due to boiling and the associated transfer of heat from the rock matrix to the fluid. If the pressure changes are small, or if one phase is immobile, the pressure variation with time in the two-phase fluid may be solved by using the linear diffusion equation (Grant 1978, Garg 1978, Moench & Atkinson 1978, Grant & Sorey 1979). The compressibility in this equation is a two-phase one, i.e. it takes into account the boiling process, and may be 100–10,000 times that of liquid water, or 10–100 times that of superheated steam.

For larger pressure changes nonlinear effects are to be anticipated, and these have been studied by Sorey et al (1980) using numerical and quasi-analytical solutions of the two-phase flow equations for a well discharging at constant mass rate. The flowing enthalpy is shown to increase as the liquid saturation
decreases during drawdown. It does, however, reach a stable value near the well face at sufficiently short times that well-test analyses can still be carried out. Since this work was done, several two-phase transient analyses have, in fact, been carried out with credible results. One such study is reported by Grant (1980). The studies are now being extended to cases in which phase boundaries occur, i.e. where the boiling only takes place in the immediate vicinity of the well (Horne & Satman 1980).

For reservoir analysis, the major effect of the two-phase zone is the slowing down of the propagation of pressure drawdown across the field. The unified drawdown of a two-phase reservoir, thus, may not occur for some time. In the case of low permeability reservoirs, it may never occur. Specific fields, such as Broadlands and Ngawha in New Zealand, are discussed in this regard by Donaldson & Grant (1981a).

Although the water may be virtually immobile and the pressure gradient approaching vapo-static, the behavior of vapor-dominated geothermal reservoirs will be similar to that of two-phase liquid-dominated ones. Vapor-dominated reservoirs have, however, been the specific topic of several studies. Many more field data have become available since the effects of exploitation were discussed qualitatively by Truesdell & White (1973). Data analyses can thus be carried out and theories tested much more readily (see, for example, Atkinson et al. 1978a,c, Lipman et al 1978).

Theoretical and experimental studies relating to fluid withdrawal from vapor-dominated geothermal reservoirs have been carried out by Moench & Atkinson (1978) and Herkelrath & Moench (1978, 1980). Moench & Atkinson (1978) used a finite difference model for the radial horizontal flow of steam through a porous medium containing immobile but vaporizing water to evaluate pressure transient effects. The enhanced compressibility due to vaporization has already been discussed above. Herkelrath & Moench (1978, 1980) extended that study and cross-checked their numerical results with laboratory experiments.

Although reinjection is being carried out in several geothermal fields, it has not been as successful as hoped, nor does the heat and fluid movement stimulated by the reinjection match the idealized theoretical conception (Horne 1981). When cold fluid is reinjected into a liquid-saturated section of the reservoir, it is probably only the heterogeneous nature of the reservoir structures that creates problems. In a recent study, however, Grant (1981) showed that reinjection of cold fluid into a two-phase zone can cause a decrease in pressure over and above that due to the extraction of the hot fluid, rather than the increase in pressure that would be anticipated if we injected into liquid. The cold fluid must condense some of the steam in order to come into thermal balance with the fluid and rock in place. Only if the temperature of the reinjected fluid is close to that of the reservoir will the volume of fluid injected make up for the steam condensed and, hence, maintain or increase the pres-
sure. In most real situations the pressure will drop. Grant (1981) further points out that the removal of heat alone, as, for example, through the use of a downhole heat exchanger, gains nothing in this situation. The heat must be supplied from the fluid and rock around the well, and under saturation conditions that must mean a drop in pressure.

**Real Reservoirs**

In studies of real reservoirs, hard field data must be matched. It is thus surprising that in spite of the extreme complexity of the reservoirs, many of the simplest models have been moderately successful in matching the real field behavior. Zais (1979, 1980) has shown decline curve methods, using the exponential equation, to work well on geothermal production data; other workers are matching the behavior of various reservoirs with lumped-parameter or simple material balance models. The recent analysis of Wairakei data by Fradkin (1981) produced a lumped-parameter model with only three identified parameters. Vapor-dominated systems are being matched using the linear nature of the p/Z v cumulative production behavior suggested by Brigham & Morrow (1977). This approach has been applied recently to the Serrazano (Atkinson et al 1978c) and Gabbro (Brigham & Neri 1980) zones of Larderello, and to Travale-Radicondoli (Atkinson et al 1978a). (All of these fields are in Italy.) Atkinson et al (1978b) have also applied the procedure to the reservoir at Bagnore, Italy. This reservoir has a very high content of noncondensable gas.

To understand the movement of heat and fluid within a geothermal reservoir in detail it is obviously necessary to include reservoir structure in the model. In this event, numerical modeling may be the only approach. Its use, however, is dependent on obtaining extensive and detailed information concerning the reservoir. To date, various models of several geothermal reservoirs have been attempted, with varying degrees of success (see, for example, the Heber model of Tansev & Wasserman 1977; the East Mesa models of Riney et al 1979 and Morris & Campbell 1981; and the Wairakei models of Mercer & Faust 1979 and Pritchett et al 1980).

Considerable effort is now going into the improvement and application of such modeling. A recent intercomparison test of most models available that are capable of analyzing mixed single- and two-phase systems showed little spread in results (Stanford Geothermal Program 1980).

**CONCLUSION**

On paper, at least, an understanding in general terms of how heat and fluid move in geothermal systems and reservoirs has now been reached. We have assessments of flows in ideal, fractured, boiling, or gas-charged reservoirs; in
liquid- or vapor-dominated states; and under natural or exploited conditions. We appear to have the tools, in lumped-parameter or numerical models, to forecast the flows, the energy recovery, and other behavior changes of a reservoir if it is exploited or if there is a change in exploitation. However, there remains this paradox—until a variety of geothermal reservoirs have been developed, we cannot prove our models, but until be prove our models, we cannot be sure that geothermal reservoirs will be developed.

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CIRCULATION IN GEOTHERMAL SYSTEMS


