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TOPICAL REPORT

**Aspects of Forced Convective
Heat Transfer in Geothermal Systems**

by

Kevin Kilty, David S. Chapman, Charles Mase

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ASPECTS OF FORCED CONVECTIVE HEAT
TRANSFER IN GEOTHERMAL SYSTEMS

by

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ABSTRACT

A knowledge of convective heat transfer is essential to understanding geothermal systems and other systems of moving groundwater. A simple, kinematic approach toward convective heat transfer is taken here. Concern is not with the cause of the groundwater motion but only with the fact that the water is moving and transferring heat.

The mathematical basis of convective heat transfer is the energy equation which is a statement of the first law of thermodynamics. The general solution of this equation for a specific model of groundwater flow has to be done numerically. The numerical algorithm used here employs a finite difference approximation to the energy equation that uses central differences for the heat conduction terms and one-sided differences for the heat convection terms. Gauss-Seidel iteration is then used to solve the finite difference equation at each node of a non-uniform mesh.

The Monroe and Red Hill hot springs, a small hydrothermal system in central Utah, provide an example to illustrate the application of convective heat transfer theory to a geophysical problem. Two important conclusions regarding small geothermal systems follow immediately from the results of this application. First, the most rapid temperature rise in the convecting part of a geothermal system is near the surface. Below this initially rapid temperature increase the temperature increases very slowly, and thus temperatures

extrapolated from shallow boreholes can be seriously in error.

Second, the temperatures and heat flows observed at Monroe and Red Hill, and probably at many other small geothermal areas, can easily result from moderate vertical groundwater velocities in faults and fracture zones in an area of normal heat flow.

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List of Symbols

- α - angle between the finite difference net and the streamlines of the fluid flow field.
- α_c - thermal diffusivity of the solid-fluid composite material.
- C_c - specific heat of the solid-fluid composite material.
- C_f - specific heat of the fluid.
- C_s - specific heat of the solid.
- ∇ - del operator.
- h - finite difference net spacing.
- H - heat transfer coefficient.
- K - Thermal conductivity.
- K_c - Thermal conductivity of the solid-fluid composite material.
- L_o - Characteristic length of the flow field.
- η - porosity of the solid.
- ρ_f - density of the fluid.
- ρ_s - density of the solid.
- T - temperature
- T_s - mean surface temperature.
- T_a - mean temperature of the air.
- T_o - characteristic temperature (Temperature at a depth of L_o in pure conduction).
- t - time
- τ_o - characteristic time increment
- $\vec{v}$ - fluid velocity field
- v_o - characteristic velocity of the field
- x - horizontal coordinate

y - horizontal coordinate

z - vertical coordinate

List of Dimensionless Symbols and Definitions

$\nabla^* = L_0 \nabla$ = dimensionless del operator

$N_{Pe} = \frac{c_f \rho_f V_0 L_0}{K_c}$ = Peclet number = ratio of heat convected to heat conducted in the Bulk fluid flow.

$N_{Nu} = L_0 \frac{\partial}{\partial z} \left(\frac{T - T_\alpha}{T_0 - T_\alpha} \right)_{z=0}$ = Ratio of heat transfer to heat transfer in the absence of convection.

$N_{sf} = \frac{V_0 \tau_0}{L_0}$ = Strouhal number

$\theta = \frac{T - T_a}{T_0 - T_a}$ = dimensionless temperature

$V^* = V/V_0$ = dimensionless velocity

$x^* = x/L_0$ = dimensionless horizontal coordinates

$y^* = y/L_0$ = dimensionless horizontal coordinates

$z^* = z/L_0$ = dimensionless vertical coordinates

$R = \text{ratio of heat capacities} = \frac{c_f \rho_f}{c_s \rho_s}$

THEORY OF CONVECTIVE HEAT TRANSFER IN GEOTHERMAL PROBLEMS

Introduction

The theory of heat conduction has achieved great success in such diverse geophysical studies as the growth of sea ice (Stefan, 1891), the cooling of intrusives and metamorphism (Jaeger, 1957, 1964; Verhoogen et al., 1970), and the bathymetry of the oceans (Parsons and Sclater, 1977). It has had somewhat less success, however, when applied to geothermal systems. These systems typically show rapid variations of geothermal gradient with depth and rapid spatial variations in heat flow (Elder, 1965). The rapid variations are caused by convective heat transfer; a phenomenon that the theory of pure conduction cannot describe, but an understanding of which is central to the study of geothermal systems and other systems of moving groundwater.

Studies of convective heat transfer in geophysical problems have taken a variety of forms. Stallman (1963) showed that temperature measurements could be used to estimate groundwater velocities in aquifers. However, he went only so far as a qualitative assessment of the problem. Using a solution to the one dimensional form of the energy equation, Bredehoeft and Papadopoulos (1965) and Sorey (1971) calculated vertical groundwater velocities from temperature logs. Donaldson (1970) formulated a simultaneous inversion of pressure and temperature data to determine both the permeability and flow rates

within an aquifer and then applied this scheme to the Wairakei geothermal system. Lewis and Beck (1977) modeled convective heat transfer numerically to explain the distribution of heat flow values they observed in a glacial complex. Bodvarsson (1961), in an early study, applied convective heat transfer by flow in conduits to the geothermal areas of Iceland and Elder (1965) made a similar study of the geothermal area at Wairakei, New Zealand. Domenico and Palciauskas (1973) analyzed heat transfer by forced convection in homogeneous, isotropic alluvium. Their results are only approximately correct, however, because they applied separation of variables to an equation that was not separable. Sorey (1975) and Witherspoon (1975) have studied convection numerically in more general porous solids.

With the exception of the study by Domenico and Palciauskas (1973), the studies of convection in porous solids and in geothermal systems have been preoccupied with free convection. This preoccupation has two negative aspects. First, it implies that hydrothermal systems result only from free convection and therefore require exotic sources of heat. Second, there is an emphasis on the empirical relationships of dimensionless groups, such as the Rayleigh number, to the nature of the convection. However, dimensionless groups are local quantities that sometimes have only a superficial relationship to convection which affects every part of a fluid system and is therefore a global phenomenon (Rosenberger, 1978).

This analysis follows a simple, kinematic approach. Concern is not with the dynamics of the fluid motion but only with the fact that the fluid is moving and transferring heat. The goal is to determine

the range of fluid flows that are consistent with the observed temperature, discharge, and heat flow in a hydrothermal system. This knowledge is subsequently used to determine the driving force of the convection, the nature of the heat source, and to estimate what temperatures lie below the present depth of measurement.

Conceptual and Mathematical Models

Figure 1 shows a simple, conceptual model of a fault controlled geothermal system. The thermal water originates as surface water that percolates down through faults, fractures, and aquifers into a storage area beneath the hot springs. During the downward percolation and during its residence time in the storage area the water absorbs heat from its surroundings. It is eventually forced from the storage area through a fault zone to a discharge area at the surface. The pressure that forces the water to the surface is supplied principally by the difference in elevation between the recharge area and discharge area although bouyancy over heat sources could supply some additional pressure. At the surface part of the water discharges through the hot spring while the remainder is lost through leakage to the alluvium. The location of the hot spring is determined by the piezometric surface, topography, and the location of a permeable conduit in the fault zone.

The mathematical analysis of this model involves solving the conductive and convective heat transfer equation, also known as the energy equation, for the appropriate boundary conditions and velocity

Figure 1 Conceptual model of a fault controlled geothermal system

field of the groundwater. The time dependent form of this equation is (Stallman, 1963)

$$1.) \nabla \cdot K_c \nabla T - C_f \rho_f (\vec{\nabla} \cdot \nabla T) = C_c \rho_c \frac{\partial T}{\partial t}.$$

In this equation K_c , C_c , and ρ_c are the thermal conductivity, specific heat, and density of the solid-fluid composite, C_f and ρ_f are the specific heat and density of the fluid, $\vec{\nabla}$ is the volume average velocity field of the fluid, T is the temperature, and t is time.

The validity of this equation depends on two conditions. First, it is valid only if the divergence of the velocity field, $\vec{\nabla} \cdot (\vec{v})$, and the viscous dissipation are negligible. This is approximately so in steady, groundwater flow. Second, it is strictly valid only if the solid and fluid phases can be regarded as coexisting continua. This restriction is well satisfied if the pore spaces and fractures through which the flow takes place are much smaller than the distance over which there is a resolvable change in temperature (Singh and Dybbs, 1974). If these restrictions are not met the energy equation contains additional complicating terms. Singh and Dybbs (1974) and Slattery (1972) discuss the form of these additional terms.

The energy equation cannot always be solved independently of the flow equations which determine the velocity field, V . In forced convection problems the velocity field is solely the result of external pressures and can be determined independently of the temperature field. In free convection, however, the velocity field is solely the result of thermal buoyancy. In this case the energy equation and the flow equation, Darcy's law, have to be solved

simultaneously. Forced convection is the most common type of fluid flow in porous materials, free convection being important only in the vicinity of a powerful heat source such as a cooling intrusive. Thus, for the remainder of this paper free convection is neglected and the term convection refers only to forced convection.

Qualitative Aspects of Convective Heat Transfer

A dimensionless form of the energy equation is useful for qualitatively discussing the behavior of convective heat transfer. If the quantities L_0 , V_0 , τ_0 , and T_0 are, respectively, a characteristic length, velocity, time, and temperature in a convective flow, the transformations $\nabla^* = L_0 \nabla$, $v^* = \vec{v}/V_0$, $t^* = t/\tau_0$, and $(\theta = T - T_s)/(T_0 - T_s)$, where T_s is the mean surface temperature, result in a dimensionless energy equation

$$2.) \frac{1}{N_{Pe}} \nabla^{*2} \theta - (v^* \cdot \nabla^* \theta) = \frac{R}{N_{St}} \frac{\partial}{\partial t^*} \theta.$$

With the exception of the factor R , this is the equation derived by Slattery (1972). R is the ratio of the heat capacity of the solid-fluid composite, $C_c \rho_c$, to the heat capacity of the fluid, $C_f \rho_f$, N_{St} is the Strouhal number defined as

$$N_{St} = \frac{V_0 \tau_0}{L_0},$$

and N_{Pe} is the Peclet number defined as

$$N_{Pe} = \frac{C_f \rho_f V_0 L_0}{K_c}.$$

The Peclet number is the ratio of heat transferred by convection to the heat transferred by conduction (Rosenberer, 1978). If convection is dominating the heat transfer the Peclet number is large and the first term in equation 2) is negligible. In this case the time dependence of the temperature field is controlled by the factor R/N_{St} and the time required to reach thermal equilibrium is proportional to L_0/V_0 . On the other hand, if conduction dominates the heat transfer the Peclet number is small and the second term of equation 2) is negligible. The time dependence of the temperature field is now controlled by the factor RN_{Pe}/N_{St} . In this case the time required to reach thermal equilibrium is proportional to L_0^2/α_c , where $\alpha_c = K_c/P_c C_c$ is the thermal diffusivity of the composite material.

In a geothermal system, the local value of the Peclet number varies drastically from place to place. The system therefore equilibrates unevenly. Heat is transferred more rapidly in regions affected by the flow than in regions depending exclusively on conduction for heat transfer.

A similar analysis is also possible for the steady state behavior of the energy equation. The steady state energy equation is simply equation 2) with the right hand side set equal to zero. If the Peclet number is small the second term of the equation (convection) is negligible. In this case the solution is very similar to that of pure conduction. Alternatively, if the Peclet number is large the first

term of the equation (conduction) is negligible and the energy equation reduces to

$$V^* \cdot (\nabla^* \theta) = 0.$$

The only realistic solution of this equation is θ equal to a constant throughout the most rapid part of the flow (Rosenberger, 1978), and so at large Peclet numbers a constant temperature fault plane is a fair approximation to the model of figure 1 (Blackwell and Chapman, 1977).

Although this qualitative discussion of the energy equation demonstrates the character of convective heat transfer, the analysis of a real geothermal system requires a solution to the energy equation for a specific flow field. Unfortunately analytic solutions cannot be obtained except for the simplest situations. In the general case a numerical solution must be sought.

A finite difference approximation to the equation approximates the second order partial derivatives as second order central differences and approximates the first order partial derivatives as first order, forward or backward differences oriented upstream in the fluid flow (Courant, 1952). This upstream difference equation can be used in a matrix iterative algorithm, such as Gauss-Siedel iteration, even when the groundwater velocities are very large.

The finite difference equation is solved at each node of a finite difference mesh. The size of the mesh need not be uniform but can be made very fine where detailed temperature information is sought and coarse where detailed information is unnecessary. The non-uniform grid makes it possible to move the boundaries far from the interesting

part of the flow. A complete description of the numerical algorithm is included in the appendix.

The accuracy and rate of convergence of the algorithm is strongly dependent on the type of boundary conditions that are implemented. Constant temperature boundaries provide the most rapid convergence but are the least realistic. A constant flux boundary condition is a better approximation to the lower and vertical boundaries and the upper boundary, which is the air-ground interface, is best represented by a mixed boundary condition expressed mathematically as

$$H(T_s - T_a) = K_c \left. \frac{\partial T}{\partial z} \right|_{z=0}$$

$(T_s - T_a)$ is the difference between the surface temperature and the ambient air temperature, and H is a semi-empirical film coefficient between the ground surface and the air. This boundary condition frees the surface from an isothermal state and allows phenomena such as warm or steaming ground to evolve naturally in the model.

An approximate value of the film coefficient can be determined from boundary layer theory by knowing the mean wind speed in the area (Chapman, 1974). Sorey (1975) includes a table of values for the film coefficient at various wind speeds in his Ph.D. thesis. However, estimating the film coefficient from wind speeds is not completely satisfactory because it assumes that the heat transfer between the earth and air occurs by conduction through a stagnant film of air. The actual heat transfer mechanism is probably transfer of water vapor to the air in evapo-transpiration (Sophocleous, personal communication).

A value of $0.6 \text{ Watts m}^{-2}\text{K}^{-1}$ is probably a reasonable value of the film coefficient. Using this value, steaming ground appears in the numerical models at vertical groundwater velocities that are reasonable in geothermal areas.

A phenomenon that seriously degrades the accuracy of this numerical method, and all other methods that employ one-sided finite differences, is the phenomenon of fictitious diffusion (Gosman et al., 1969). This is a smearing of the solution that occurs at high fluid velocities which is caused by the truncation error of the one-sided differences. The magnitude of the fictitious diffusion is proportional to $|V|h \sin 2\alpha$, where $|V|$ is the absolute value of the groundwater velocity, h is the spacing of the finite difference nodes, and α is the angle between the streamlines and the finite difference net (Gosman et al., 1969). In geophysical problems h is generally quite large ($\sim 50\text{m}$) so that the fictitious diffusion may be substantial even for small groundwater velocities.

Two ways of reducing the effect of fictitious diffusion are to use a fine finite difference net (small h) or a finite difference net that parallels the streamlines of the flow. Gosman et al. (1969) discuss this in more detail.

Application of Theory to Geophysical Problems

The most fruitful approach toward applying the theory of convective heat transfer to geophysical problems should be the approach taken in fluid mechanics. In fluid mechanics the solutions to many problems are empirical relations, determined experimentally or

numerically, between a determining criterion such as the velocity of a specific flow field and a non-determining criterion such as the momentum that is transferred. These relationships are most useful if the comparison is between dimensionless criteria (Luikov, 1966). In heat transfer problems the appropriate determining criteria are the dimensionless coordinates and the Peclet and Strouhal numbers. The appropriate non-determining criteria are the dimensionless temperature and the Nusselt number, N_{Nu} , which is the ratio of total heat transfer to the heat that would be transferred by conduction alone (Luikov, 1966). An example of two relationships between determining and nondetermining criteria in a geophysical context is illustrated in figures 2 and 3.

Figure 2 is a numerically determined relationship between the Nusselt number at the earth's surface directly over a fracture and the Peclet number of the flow up the fracture. As expected, the Nusselt number is a monotonically increasing function of Peclet number and asymptotically approaches a value of 1 as the Peclet number approaches zero.

Figure 3 is the relationship between the Nusselt number on the earth's surface and the dimensionless distance away from the fracture for a Peclet number of about 10^4 . The Nusselt number in the figure is normalized by dividing by the Nusselt number directly over the fracture and the distance is made dimensionless by dividing by the depth at which the flow originates. As expected, the Nusselt number assumes a maximum value over the fracture and decreases in value in

Figure 2 Relationship between local Nusselt number and Peclet number on the earth's surface directly over flow up a fracture

LOCAL NUSSELT NUMBER AT HYPOTHETICAL HOT SPRING

Figure 3 Relationship between Nusselt number and distance away from the surface trace of the fracture

both directions away from the fracture with an asymmetry that depends on the dip of the fracture.

These two figures completely specify the heat flow that would be observed near a fracture that has water flowing up it. Also, these figures can be used for the inverse problem which is that of determining fracture geometries and groundwater velocities that are consistent with heat flow measurements at the surface.

Many other relationships between determining and nondetermining criteria are possible. For instance Jaffrennou et al. (1974) experimentally determined a relationship between Nusselt number and filtration Rayleigh number in free convection and Sorey (1975) numerically determined a relationship between the dimensionless drop in water temperature between the reservoir and hot spring and the dimensionless discharge of the hot spring. The relationships that are most useful, of course, depend on the nature of the specific problem.

APPLICATION TO THE MONROE, UTAH HYDROTHERMAL SYSTEM

Introduction

The Monroe and Red Hill hot springs are located on the east side of the central Sevier Valley near Monroe, Utah. They issue from the Sevier fault which is a major, west-dipping normal fault that separates the Tertiary volcanic rocks of the Sevier Plateau from the alluvium of the Sevier Valley.

The first detailed studies of these hot springs (Parry, et al., 1976; Miller, 1976) were concerned with the geochemistry and alteration mineralogy of the systems and provide valuable constraints on the origin and history of the thermal waters. During the autumn of 1976 and throughout 1977, geophysical surveys including precision gravity, ground-magnetics, and dipole-dipole electric soundings were conducted and a borehole drilling program was undertaken to provide direct subsurface temperature information (Halliday, 1978; Mase, 1978).

Hydrogeology

The hydrogeologic situation in the central Sevier Valley is shown in figure 4. The geologic and hydrologic information in the figure is generalized from Young and Carpenter (1965), Hahl and Mundorff (1968), and Stokes and Hintze (1963).

Figure 4 Generalized hydrogeologic map of the area adjacent Monroe and Red Hill hot springs

	UNDIFFERENTIATED IGNEOUS ROCKS.
IMPERMEABLE EXCEPT IN FRACTURES
AND FAULTS.		HOT SPRING
	UNDIFFERENTIATED PERMEABLE
SEDIMENTARY ROCKS.		COLD SPRING
	UNDIFFERENTIATED SEDIMENTARY ROCKS.
IMPERMEABLE OR OF UNKNOWN PERMEABILITY		MAPPED FAULT
(DASHED WHERE
INFERRED)
	QUATERNARY ALLUVIUM		

5 4 3 2 1 0 5 10 15 KILOMETERS

The subsurface hydrology of the central Sevier Valley is very complex. This is partially because of complicated geological structure and partially because of the number of aquifers that could contribute to the subsurface flow. The Sevier Plateau and the Pavant Range southwest of the town of Richfield are covered with up to 4 km of Oligocene and Pliocene volcanic rocks (Young and Carpenter, 1965). The volcanic rocks consist of pyroclastic rocks in the lowest part overlain by latite flows which are, in turn, overlain by basaltic andesite flows, rhyolite, and tuffs. Young and Carpenter (1965) indicate that these volcanic rocks generally have very low permeabilities. The pyroclastic rocks, tuffs, and latite flows surely have low permeability but, the basaltic andesite flows are locally very permeable as shown by springs that discharge from them. Moreover, all of the volcanic rocks have higher permeabilities in the fractured and faulted zones.

The volcanic rocks lie upon sedimentary rocks that range in age from Oligocene to Jurassic, depending on location. The silty and shaly sequences, such as the Jurassic Arapien shale, have low permeability but many of the sandstones and limestones, such as the Jurassic Navajo sandstone and the tertiary Flagstaff limestone, are very permeable and make good aquifers (Young and Carpenter, 1965). These aquifers outcrop in the Tushar Mountains southwest of Marysvale and they also outcrop extensively in the Pavant Range west of Richfield where they dip to the south and southeast under the volcanic rocks (Callaghan and Parker, 1961). Their structure under the volcanic rocks is unknown. The Mesozoic and older sedimentary rocks are folded

and thrust faulted (Hardy, 1952) and all of the rocks, including the volcanic rocks, were faulted and fractured during Basin and Range normal faulting which began during Miocene times and has continued sporadically to the present.

The recharge of the sedimentary aquifers occurs in the mountainous areas which receive 75cm of water or more per year as precipitation. The aquifers can recharge directly from surface precipitation wherever they outcrop; otherwise their recharge depends on available fractures and faults in the overlying volcanic rocks. Once the groundwater has infiltrated these aquifers it may move laterally into reservoirs deep beneath the valley floor.

The valley floor is formed of narrow, down dropped fault blocks covered with Pliocene and Pleistocene sediment which vary in thickness from a thin veneer to as much as 500 meters (Young and Carpenter, 1965). The fill lies upon an unknown thickness of volcanic rocks which isolate it from the deeper, sedimentary aquifers. The fill is composed of conglomerates, clays, landslide deposits, and terrace and streambed gravels. This heterogeneous composition causes the permeability of the fill to vary from good to poor.

The Hot Springs

Hot springs occur in well defined seeps on the Sevier fault at Monroe, Red Hill, and Johnson warm spring and on the Dry Wash fault at Joseph hot spring. The springs occur at changes in the trend of the surface trace of the Sevier fault. Perhaps these are the only locations where the fault zone is permeable enough for the water to

penetrate the surface. Conceivably, hot water rises from depth in many other places along the fault only to mix with cool water and leak into the alluvium before it can discharge at the surface.

Mase (1978) has drawn inferences about the near-surface, hot spring flow from the first separation dipole-dipole apparent resistivity shown in figure 5. The contours generally follow the Sevier fault as located by the ground magnetic survey (Mase, 1978). The ionic strength of the thermal fluids is 6.5×10^{-2} which is equivalent to 3800 ppm NaCl. At the temperature of the hot springs (near 70°C), 30% porous sediments saturated with these fluids would have an apparent resistivity of 5 Ω -m (Keller, Frischnecht, 1966). Once the fluids have cooled to 25-30°C the sediments containing them would have an apparent resistivity of 15-10 Ω m. On this basis the elongate low resistivity zones ($\rho_a < 10 \Omega$ m) probably indicate the warmest and most altered parts of the system. The arm of low resistivity projecting northwest from Red Hill is interpreted as leakage of cooling thermal fluids through the alluvium. A similar leakage zone is indicated extending westward from Johnson Warm Springs.

Subsurface Temperature Data

Mase (1978) obtained direct subsurface temperature data in 11 boreholes, the positions of which are indicated on the insets in figures 6 and 8 with the temperature-depth curves. The appearance of the temperature-depth curves is similar to those that Lachenbruch et al. (1976) describe at Long Valley, California and those that Smith (1970) describes at Broadlands, New Zealand.

Figure 5 First separation, dipole-dipole resistivity map of the hot springs

A peculiarity of temperature-depth curves near strongly convecting zones is that the thermal gradient is quite variable with depth, so that making temperature extrapolations on the basis of these curves is risky. The nearly isothermal appearance of two of the boreholes at Red Hill below 30 m depth indicates a strongly convecting zone. The other boreholes at Red Hill and all those at Monroe show a strong, consistent, downward curvature. This curvature may indicate an increase of thermal conductivity with depth, a weak, diffused upward convection, or lateral heat conduction away from a strongly convecting zone. Mase (1978) measured the thermal conductivity in all of the boreholes at 5 m intervals, and found no increase in thermal conductivity with depth. The cause of the curvature is then either upward convection or lateral heat conduction.

An unfortunate aspect of the measurements at Red Hill and Monroe is that none were made away from the immediate area of the hot springs. Such data contain information about the subsurface flow and the location of the recharge area (Lachenbruch et al., 1976).

The temperature data at Monroe and Red Hill are displayed on the east-west cross sections of figures 7 and 9. In each the isotherms are upwarped under the discharge vent and taper off to the east and west. The asymmetry in the isotherms is attributed to several effects, such as the dip of the fault zone through which the hot water is rising, lower thermal conductivity of the alluvium west of the fault, and possible warm water leakage to the west.

Figure 6 Temperature-depth curves at Monroe Hot spring

Figure 7 Temperature cross-section at Monroe Hot spring

Figure 8 Temperature-depth curves at Red Hill hot spring

Figure 9 Temperature cross-section at Red Hill hot spring

Geochemical and Geophysical Constraints

From geochemistry, Parry et al. (1976) determined that the spring waters are most likely a mixture of a hot, saline component and a cool component that is chemically similar to the surface water in the area. A mixing model, based on the measured silica content of the warm water and the warm water enthalpy, indicated a mixture of 62% hot water at 118°C and 38% cold water at 10°C. The Na-K-Ca geochemistry, indicated a hot component temperature of 180°C. In a normal Basin and Range heat flow area these temperatures are reached between 2.5 and 4.0 km depth. The Na-K-Ca temperature estimate is less reliable than the silica temperature because the hot spring water may have acquired its salinity from evaporites in the Arapien shale. Consequently the hot reservoir is probably nearer a temperature of 118°C than 180°C. What aquifer constitutes the hot fluid reservoir is a matter of pure speculation but, the most likely possibility is the Jurassic Navajo sandstone because it is generally permeable and probably underlies the Sevier Valley at a depth between 2.5 and 4.0 km (Mase, 1978).

The location of mixing between the hot and cold waters is also speculative. The temperature of the cold springs in the area ranges from about 12 to 20°C. Assuming that the water in the cold springs loses no more than 10°C between depth and the surface (Sorey and Lewis, 1976), the maximum temperature that the cool spring waters attain is about 30°C which implies that this water circulates no deeper than 500m. This temperature estimate is in apparent conflict with Miller's (1976) geochemical estimates of the maximum subsurface temperature of the cool springs. These estimates, which were based on

silica content, ranged from 80°C to over 110°C. There is no satisfactory explanation for this conflict but part of the problem might be that silica thermometry is not always accurate when applied to cool surface water.

The discharge at the Red Hill spring is about 20 lit.sec⁻¹ and the combined discharge of all the seeps at Monroe hot springs is about 6 lit.sec⁻¹ (Mase, personal communication). Supposing that the upward flow takes place over an area of 10⁴m², which is about the area of the 5 m contour in figure 5, the vertical groundwater velocities in the permeable zones range from 5 x 10⁻⁷m sec⁻¹ to 2 x 10⁻⁶m sec⁻¹. These velocities are minimum estimates because the affect of leakage to the alluvium has not been considered.

The permeability of the fault zone is not known. Sorey and Lewis (1976) consider permeabilities of 1 to 10 darcies appropriate values for the fault zones in Long Valley. If an intermediate value of 5 darcies is chosen for the permeable part of the fault zone at Red Hill, a vertical piezometric gradient of 2% is required to account for the spring discharge. This piezometric gradient is equivalent to a 50 m decrease in piezometric head in a 2.5 km rise from the reservoir to the surface. Since the piezometric gradient is the same at Monroe as it is at Red Hill, the lower vertical velocities at Monroe imply a lower permeability in this system.

Heat Output of the Hot Springs

The total power loss at Monroe and Red Hill, estimated by integrating the heat flow over the area enclosed by the 20 m contour

in figure 5 and adding the enthalpy of the discharge, is approximately 8 MW. This is equal to the normal Basin and Range heat flow over a 110 km² area. Chapman and Wilson (personal communication) observe that recharging groundwater absorbs 30 to 50% of the heat flow in a recharge area. Thus, heat absorbed over a recharge area of 220 to 370 km² could maintain the observed power loss at Monroe and Red Hill. This is only about 1/4 of the maximum possible recharge area east, west, and south of the hot springs. In addition to the heat absorbed in a recharge area, groundwater also absorbs a substantial amount of heat in lateral flow to the storage reservoir.

Parry (personal communication) has suggested that exothermal alteration of clays in the hydrothermal system may provide a significant portion of the observed power loss. The power provided by chemical reactions depends on the length of time the system has been active and the extent of the alteration which are not known.

Discussion of Model Results

The numerical algorithm was used to compute the temperature distributions of several models of the subsurface flow. Figure 10 shows the steady state temperature field resulting from a flow of $1 \times 10^{-6} \text{ m sec}^{-1}$ up a vertical fault. The flow originated at a depth of 2.0 km and it was assumed that no water leaked from the fault zone but that mixing with cool water occurred between 400 and 500 meters depth. The steady state spring temperature in the model was 81°C compared to 77°C measured at Red Hill. An important aspect of this model is that

Figure 10 Temperature cross-section and heat flow profile for groundwater flow up a hypothetical vertical fracture

in the strongly convecting part the temperature increases rapidly with depth near the surface but increases very slowly at greater depths.

Figure 11 is a comparison of the observed heat flow at Monroe and Red Hill with two other models of the subsurface flow. The figure is a graph of normalized Nusselt number against dimensionless distance from the hot spring perpendicular to the strike of the fault. The distance is made dimensionless by dividing the ground distance by the depth at which the flow originated. One model is that of water flowing up a 75° dipping fault from a depth of 2.4 km without leakage. The other model is the same except that complete leakage occurs in the upper 400 m of alluvium.

The observations at Red Hill agree closely with the models east of the hot spring, and so the vertical flow at Red Hill is evidently up a narrow zone (20m wide) that has a dip near 75° . West of the hot spring, where leakage is presumably taking place, the observations at Red Hill plot between the two models. The leakage at Red Hill is different from that in the model for several possible reasons. First, only a fraction of the flow at Red Hill leaks into the alluvium whereas the model assumes complete leakage. Second, the leakage is probably confined to a narrow zone extending northwest of the hot spring. The narrowness of the leakage zone makes three dimensional fluid flow and heat conduction effects important.

At Monroe Hot Springs the observations fall below the model predictions west of the hot springs. Thus the upward flow at Monroe is evidently accomplished without leakage in a nearly vertical zone. East of the hot springs the observations at Monroe are consistently

Figure 11 Comparison of normalized Nusselt numbers in the
boreholes at Monroe and Red Hill hot spring with those
of two models of the hot spring's flow

above the predictions which indicates a diffused upward flow over a broad zone.

Although most of the interpretation has thus far assumed flow in fracture zones, other interpretations are possible. Figure 12 illustrates the ambiguity of interpreting the temperature-depth curve of borehole M₄. Over the depth it was measured, the temperature-depth curve in M₄ is indistinguishable from the theoretical curve for diffused upward flow. This theoretical curve was calculated from a one-dimensional solution of the energy equation. From the parameters of this theoretical curve ($N_{pe} = 4.1$) the interpretation of the temperature-depth curve would imply a vertical flow of $1.2 \times 10^{-8} \text{ m sec}^{-1}$ from a constant temperature reservoir of 60°C at 160m depth. However, the observed data is also indistinguishable from the temperature-depth curve that would result from flow up a 60° dipping fracture 100 m away. This curve was calculated by similarity (Luikov, 1966) from a numerical solution. From the parameters of this solution, the interpretation of the data from M₄ would imply an upward flow of $1 \times 10^{-7} \text{ m sec}^{-1}$ in the fracture, a local undisturbed geothermal gradient of 400 K km^{-1} , and a temperature of 65°C at a depth of 120 m.

While the two theoretical curves are very similar above 70 m depth; below this depth they diverge sharply. Another 20 m of data in this borehole would favor one model over the other.

Figure 12 Comparison of a temperature-depth curve at Monroe hot spring with two theoretical curves for hypothetical flow geometries

Implications for Monroe-Red Hill

Temperatures extrapolated to great depths are not generally accurate. However, on the basis of many models, the nearly isothermal behavior of two of the boreholes at Red Hill will probably continue until the zone of mixing between the hot and cold components is reached. The temperature could increase 15-20°C through the mixing zone and then continue isothermally to the hot reservoir. The strong downward curvature noted in many of the boreholes likely continues until the estimated reservoir temperature of 120°C is reached or until a strongly convecting zone is intercepted.

A cooling magma chamber is not necessary as a heat source for the Monroe-Red Hill hydrothermal system. Furthermore, there is no reason to believe that the area has an average heat flow higher than normal for the Basin and Range Province. The following evidence supports these conclusions:

- (1) Heat transfer by moving groundwater in a normal Basin and Range heat flow environment can easily provide the power loss that is observed at Monroe-Red Hill. Moreover, there may be an additional contribution to the heat flow from exothermic chemical reactions.
- (2) There is no indication of Quaternary volcanism in the area.
- (3) A piezometric gradient caused by the difference in elevation between recharge areas and the hot spring can maintain the spring's discharge. There is no need for bouyance over a heat source.

- (4) Upward groundwater flow from a reasonably shallow depth of approximately 2.0 km can maintain the temperatures that are observed at Monroe-Red Hill.

The minimum upward velocity that is consistent with the discharge and temperatures at the Red Hill spring is $1 \times 10^{-6} \text{ m sec}^{-1}$. The maximum possible piezometric gradient at the spring, which is the maximum topographic relief divided by the depth to the storage reservoir, is 60%. Therefore, the observed velocity in the hydrothermal system is attainable only if the permeability of the permeable parts of the fault zone is greater than .1 darcy.

CONCLUDING REMARKS

A simple, numerical approach to convective heat transfer provides some very basic information on the nature of hydrothermal systems. When applied to a specific geothermal system, such as Monroe-Red Hill, this approach provides some knowledge of the system's flow, its heat source, its permeability, and what temperatures can be expected. A natural extension of this approach would be to consider the complications introduced by free convection and a more sophisticated energy equation.

Studies concerned with the natural evolution of hydrothermal systems and with the evolution of these systems during exploitation are two obvious applications of this approach. Further applications may be in studies of mineral alteration near intrusives, in studies of heat transfer in the crust, and in the estimation of groundwater velocities.

APPENDIX: THE SOLUTION OF THE HEAT TRANSFER EQUATION

The differential equation of heat transfer by combined conduction and convection is simply a statement of the first law of thermodynamics. In simple terms, it states that the rate at which the heat content of an arbitrary volume of material changes is equal to the time rate of heat transfer across the boundary of this volume added to the rate at which heat is generated internal to the volume.

The rate at which the heat content of a fixed volume of porous material changes can be expressed as $\int_V C_c \rho_c \frac{\partial T}{\partial t} dv$, where C_c and ρ_c are the specific heat and density of the solid-fluid composite, and T is the temperature. Neglecting dissipation and the kinetic energy of the fluid motion, the time rate of energy transfer across the boundary of added to the rate of internal heat generation can be written in integral form as $\int_V [K_c \nabla T - \nabla \cdot C_f \rho_f \vec{v} T + Q] dv$. The first term of the integrand is the heat transferred by conduction; K_c is the thermal conductivity of the solid-fluid composite. The second term accounts for the heat transferred by convection; C_f and ρ_f are the specific heat and density of the fluid and $\vec{v}$ is the volume average velocity field of the fluid. Q is a source term representing the heat generation per unit volume per unit time.

According to the first law of thermodynamics these two volume integrals must be equal for an arbitrary volume. This results in the differential form of the heat transfer equation.

$$C_c \rho_c \frac{\partial T}{\partial t} = \nabla \cdot K_c \nabla T - \nabla \cdot C_f \rho_f \vec{v} T + Q$$

Equation 1 is difficult to solve analytically, solutions having been found only for very simple situations (Bredevhoeft and Papadopolus, 1965). The complicating factor in all cases is the groundwater velocity field, $\vec{v}$. In the case of forced convection, $\vec{v}$ can be determined independent of the temperature field and enters the equation as a variable coefficient ($\vec{v} = \vec{v}(x,y,z)$). Therefore, equation 1 is not generally separable.

In the case of free convection, a further complication arises from v not being independent of the temperature field. In this case equation 1 is nonlinear and coupled to the flow equations which determine $\vec{v}$. This means that the transfer equation and the flow equations must be solved simultaneously.

In view of the difficulty involved in analytic solutions, it makes sense to solve the problem numerically. A finite difference approximation to equation 1 is constructed by writing the second order partial derivatives as second order central differences and writing the first order partial derivatives as first order forward or backward differences oriented into the fluid flow. As an example, the two dimensional, steady state, finite difference form of equation 1

$$(K_c (\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2})) = C_f \rho_f (v_x \frac{\partial T}{\partial x} + v_y \frac{\partial T}{\partial y}) - Q \text{ is}$$

$$\begin{aligned} K_c \left\{ \frac{T_{i+1,j} - 2T_{i,j} + T_{i-1,j}}{(\Delta x)^2} + \frac{T_{i,j+1} - 2T_{i,j} + T_{i,j-1}}{(\Delta y)^2} \right\} \\ - \frac{C_f \rho_f}{2\Delta x} \{ (T_{i,j} - T_{i-1,j})(v_x + |v_x|) + (T_{i+1,j} - T_{i,j})(v_x - |v_x|) \} \\ - \frac{C_f \rho_f}{2\Delta y} \{ (T_{i,j} - T_{i,j-1})(v_y + |v_y|) + (T_{i,j+1} - T_{i,j})(v_y - |v_y|) \} + Q = 0. \end{aligned}$$

The subscripts i and j represent nodal locations in the x and y directions respectively. Δx and Δy represent finite displacements in the x and y directions and V_x and V_y represent the components of fluid velocity in the x and y directions.

Mixing central differences with forward and backward differences has the advantage that the resulting system of difference equations can be solved iteratively even when the groundwater velocities are very large and the equations are nearly hyperbolic (Courant, 1952). It has the disadvantage, however, of being less accurate than if all central differences had been used. The loss of accuracy results from the large truncation of the one sided differences which are accurate only to $O(\Delta x)$ (Gerald, 1970). Gosman et al. (1969) point out that the result of mixing central differences with forward and backward differences is a fictitious increase in thermal diffusion.

One effect of fictitious diffusion is shown in fig. 13. This is a graph of the local Nusselt number at a hypothetical hot spring against Peclet number of the groundwater flow toward the surface. The numerical algorithm was used to obtain each data point. In each solution the grid interval, Δy , was kept equal to 50m.

The broken curve is the normal relationship between the Nusselt number and the Peclet number unaffected by fictitious diffusion. This curve was determined by orienting the finite difference net parallel to the streamlines. The lower points are the relationship as determined by orienting the streamlines oblique to the finite

Figure 13 The affect of fictitious diffusion on numerical models
of convective heat transfer

difference between the curve and the points is the effect of fictitious diffusion. The heat transferred by fictitious diffusion is a substantial fraction of the total heat transfer at a Peclet number of only 100.

Two ways of reducing the effect of fictitious diffusion are to use a fine finite difference net (small h) or use a finite difference net that parallels the streamlines of the flow. Gosman et al. (1969) discuss this in more detail.

When the finite difference equation is evaluated at each node of the finite difference net, the result is a system of equations that forms a sparse, banded matrix. An iterative method, such as Gauss-Seidel iteration, is the simplest way to solve such a matrix. During iteration, the following quantities are evaluated at each node:

$$a_{ij} = \Delta x^{-2} - \frac{C_{Pf}}{2\Delta x K_c} (V_x - |V_x|),$$

$$b_{ij} = \Delta x^{-2} + \frac{C_{Pf}}{2\Delta x K_c} (V_x + |V_x|),$$

$$c_{ij} = \Delta y^{-2} - \frac{C_{Pf}}{2\Delta y K_c} (V_y - |V_y|),$$

$$d_{ij} = \Delta y^{-2} + \frac{C_{Pf}}{2\Delta y K_c} (V_y + |V_y|),$$

$$e_{ij} = a_{ij} + b_{ij} + c_{ij} + d_{ij}, \quad g_{ij} = Q_{ij}/K_c.$$

If a nonuniform grid is used these values are evaluated differently. Letting Δx_i and Δy_j denote the finite differences between the $i-1$ th and i th nodes in the horizontal direction and the $j-1$ th and j th nodes in the vertical direction, the above quantities are evaluated through the formulas:

$$a_{ij} = \frac{22}{(\Delta x_i + \Delta x_{i+1}) \Delta x_i} + \frac{C_f \rho_f (V_x + |V_x|)}{2 K_c \Delta x_i},$$

$$b_{ij} = \frac{2}{(\Delta x_i + \Delta x_{i+1}) \Delta x_{i+1}} - \frac{C_f \rho_f (V_x - |V_x|)}{2 K_c \Delta x_{i+1}},$$

$$c_{ij} = \frac{2}{(\Delta y_j + \Delta y_{j+1}) \Delta y_j} + \frac{C_f \rho_f (V_y + |V_y|)}{2 K_c \Delta y_j},$$

$$d_{ij} = \frac{2}{(\Delta y_j + \Delta y_{j+1}) \Delta y_{j+1}} - \frac{C_f \rho_f (V_y - |V_y|)}{2 K_c \Delta y_{j+1}},$$

$$e_{ij} = a_{ij} + b_{ij} + c_{ij} + d_{ij},$$

$$g_{ij} = Q_{ij} / K_c.$$

With the calculation of these quantities accomplished, the iteration proceeds according to the equation

$$T_{ij}^{m+1} = \frac{a_{ij}}{e_{ij}} T_{i+1,j}^m + \frac{b_{ij}}{e_{ij}} T_{i-1,j}^{m+1} + \frac{c_{ij}}{e_{ij}} T_{i,j+1}^m + \frac{d_{ij}}{e_{ij}} T_{i,j-1}^{m+1} + \frac{g_{ij}}{e_{ij}},$$

where the superscript indicates the stage of the iteration. This continues until the numerical solution is presumed to be very close to the analytic solution. One criterion of convergence is to stop the iteration when the

largest change in temperature between successive iterations is less than a prescribed value. While this is a convenient criterion to evaluate, it doesn't provide any information on how close the numerical solution is to the analytic one. Comparing the temperature at corresponding nodes for two different solutions, one whose initial guess is above the analytic solution and another whose initial guess is below, is one means of estimating the error of the numerical solution.

The numerical scheme is generally stable even for very large groundwater velocities. However, there are situations in which the scheme will diverge. The method is occasionally unstable when the ratio between the sizes of successive grid elements near the boundary is larger than 1.3. Instability is also occasionally noticed at very high groundwater velocities although the limiting velocity depends on the specific problem. When instability occurs a smaller, more uniform grid should be used. Another way to increase the stability is to use under-relaxation (Burggraf, 1966).

The rate of convergence of the method depends very strongly on the types of boundary conditions that are used. Gosman (1969) explains that errors in the interior of the grid are propagated by conduction and convection to the boundaries where they are reduced by the imposition of the boundary conditions. With constant temperature boundaries the error is reduced to zero during each iteration. All other types of boundaries leave a small residual error that has to be reduced in successive iterations. Therefore, constant temperature boundaries result in more rapid convergence than constant flux or mixed boundaries.

Two standard methods of increasing the convergence rate are over and

under-relaxation (Varga, 1962). Unfortunately, there is no way to calculate what the optimum relaxation parameters are (Gosman et al., 1969). The use of incorrect relaxation parameters can slow the convergence drastically and, in the case of over-relaxation, actually provoke instability. Thus if a very small under-relaxation parameter is arbitrarily chosen to add stability to the method, the convergence may be slow and expensive. The rate of convergence also depends on the initial guess. A good initial guess results in a substantial reduction in the number of iterations required to reach a specific error criterion.

Besides calculating the temperature distribution, the program must also calculate the groundwater velocity field. So far the velocity field has been assumed a priori and calculated in an external subroutine. It is possible, however, to write this subroutine as a finite difference solution for the piezometric head, ϕ (Davis and DeWiest, 1966). The velocity field is subsequently determined from Darcy's law, $V_x = -k_x \frac{\partial \phi}{\partial x}$ and $V_y = -k_y \frac{\partial \phi}{\partial y}$, where k_x and k_y are the hydraulic conductivities in the x and y directions. The advantage here is that free convection problems can be handled. Elder (1965) gives a definition of the piezometric head that includes bouyant effects. Using Hubbert's (1940) concept of force potential, this definition generalizes to

$$\phi = \int_{P_0}^P \frac{dp}{\rho} + \int_{Z_0}^Z g(1 - \beta(T(z') - T_0)) dz',$$

where P is the hydrostatic pressure at a height, z , in the fluid, P_0 is the pressure on the reference datum, Z_0 , and β is the thermal expansivity of water.

Because the dependent variable, T , occurs in the calculation of the

force potential, the solution of a free convection problem has to iterate between the subroutine where the flow is calculated and the main program where the temperature distribution is calculated.

There are many versions of codes to solve equation 1 on a digital computer. An example of the simplest code is listed on the next few pages. It solves the two-dimensional, steady-state form of equation 1 without sources on a regularly spaced grid. The code is especially simple because of the use of the NAMELIST statement to enter data. The use of this option is explained in fortran language manuals. Included in the listing are all of the subroutines except FLOW and GUESS which calculate the velocity field and the initial guess, respectively. These two subroutines have to be written specifically for the problem to be solved.

[illegible]

```

      NFH=300
C   READ TITLE
      READ(INPUT,1) TITLE
      1 FORMAT(20A4)
      READ(INPUT,DATA)
C   ECHO THE PROBLEM CONSTRAINTS
      WRITE(POUT,5) TITLE
      5 FORMAT(1H1,20A4,/,/, ' ALL VALUES ARE IN MKSA UNITS')
      WRITE(POUT,6) KF,PF,CF,KS,N,DH
      6 FORMAT(/,1X,'PHYSICAL CHARACTERISTICS OF ADVECTING MATERIAL',/,/,
        $1X,'THERMAL CONDUCTIVITY= ',F10.4,/,/,
        $1X,'DENSITY= ',F10.4,/,/,1X,'SPECIFIC HEAT= ',F10.4,/,/,1X,'PHYSICAL CH
        $ARACTERISTICS OF THE SOLID MATERIAL',/,/,1X,'THERMAL CONDUCTIVITY= '
        $,F10.4,/,/,1X,'PERMEABLE POROSITY (FRACTION)= ',F10.4,/,/,1X,
        $'DEEP CONDUCTIVE HEAT FLOW= ',F10.4,/,/)
      WRITE(POUT,7) ST,H,DX,DY,ERROR
      7 FORMAT(1X,'THERMAL CHARACTERISTICS OF THE SURFACE',/,/,
        $1X,'MEAN AIR TEMPERATURE= ',F10.4,/,/,1X,'TRANSFER COEFFICIENT (SURFA
        $CE-AIR)= ',F10.4,/,/,1X,'FINITE DIFFERENCE IN X = ',F10.4,/,/,
        $1X,'FINITE DIFFERENCE IN Y = ',F10.4,/,/,1X,'PROBABLE ACCURACY OF TH
        $E TEMPERATURES IS PLUS OR MINUS ',F10.6)
C   CALCULATE BULK THERMAL CONDUCTIVITY
      IF(TEST.EQ.0) GO TO 51
      CALL MAXW(KS,KF,N,COND)
      GO TO 16
      51 CALL RECP(KS,KF,N,COND)
      16 CALL FLOW(VX,VY)
      GRAD=DH/COND
C   INITIAL GUESS
      711 CALL GUESS(GRAD,U,BT,BR,BL,BB,DY,ST)
      712 NN=1
        XI=1./(DX**2.)
        YI=1./(DY**2.)
        ZX=CF*PF*N/(2.*COND*DX)
        ZY=ZX*DX/DY
C
C   GAUSS-SEIDEL ITERATION
C
      55 CHECK=0.
      DO 59 J=1,40
      DO 60 I=1,40
        VXX=VX(I,J)
        VYY=VY(I,J)
        AVX=ABS(VXX)
        AVY=ABS(VYY)
        UU=BT(J)
        UB=BB(J)
        UL=BL(I)
        UR=BR(I)
        IF(I.EQ.1) GO TO 101

```

```

      UU=U(I-1,J)
101 IF(I.EQ.40) GO TO 102
      UB=U(I+1,J)
102 IF(J.EQ.1) GO TO 103
      UL=U(I,J-1)
103 IF(J.EQ.40) GO TO 104
      UR=U(I,J+1)
104 A=XI-ZX*(VXX-AVX)
      B=XI+ZX*(AVX+VXX)
      C=YI-ZY*(VYY-AVY)
      D=YI+ZY*(AVY+VYY)
      E=A+B+C+D
      DELTA=(A*UR+B*UL+C*UB+D*UU)/E
      CHECK1=ABS(U(I,J)-DELTA)
      IF(CHECK1.GT.CHECK) CHECK=CHECK1
      U(I,J)=DELTA
60 CONTINUE
59 CONTINUE
      DO 198 J=1,40
      XCV=(COND/DY-VY(1,J)*CF*PF*N+H)
      XVB=(COND/DY-VY(1,J)*CF*PF*N)*U(1,J)+H*ST
      SURFT(J)=XVB/XCV
      XC=COND/DY+H
      XV=COND*U(1,J)/DY+H*ST
      BT(J)=XV/XC
      BB(J)=U(39,J)+DH*DY*2./COND
198 CONTINUE
455 DO 456 I=1,40
      BL(I)=U(I,2)
      BR(I)=U(I,39)
456 CONTINUE
C
C TEST ERROR CRITERION
C
      NN=NN+1
      IF(NN.GT.NFH) GO TO 74
      IF(CHECK.LE.ERROR) GO TO 75
      GO TO 55
74 WRITE(POUT,76) NFH
76 FORMAT(1H1,' ERROR CRITERION WAS NOT MET WITHIN THE LIMIT OF',I4,
  $ ' ITERATIONS',//)
C WRITE VALUES OF FLAGS
75 WRITE(POUT,756)
756 FORMAT(1H1,1X,' VALUES OF FLAGS',/,3X,'FLAG',3X,'VALUE',/)
      WRITE(POUT,896) (I,FLAG(I),I=1,10)
896 FORMAT(1X,I4,3X,I4)
      WRITE(POUT,77) NN,CHECK
77 FORMAT(1H1,' NODAL TEMPERATURES FOR STEADY STATE SOLUTION',//,
  $1X,I5,' ITERATIONS WERE COMPLETED. THE LARGEST CHANGE IN THE LAST
  $ITERATION WAS ',F10.6,' DEGRESS KELVIN.',//
  $1X,' DEPTH NODE',3X,' LATERAL NODES ',//)
      NK=1

```

```

      NH=10
500 WRITE(POUT,89) (NM,NM=NK,NH)
      89 FORMAT(9X,10(3X,I4,4X),/)
      DO 81 I=1,40
      WRITE(POUT,87) I,(U(I,J),J=NK,NH)
      87 FORMAT(1X,I4,6X,10E11.4)
      81 CONTINUE
      NK=NK+10
      NH=NH+10
      IF(NH.GT.40) GO TO 121
      WRITE(POUT,88)
      88 FORMAT(1H1)
      12 FORMAT(15,2E12.4)
      GO TO 500
121 WRITE(POUT,707)
707 FORMAT(1H1,' SURFACE TEMPERATURES',/,1X,'NODE',4X,'TEMPERATURE',
  35X,' HEAT FLOW',5X,' WATER ',/,
  $T9,'(CONDUCTIVE)',T26,'(CONDUCTIVE)',T41,'TEMPERATURE')
      DO 202 J=1,40
      FHEAT=H*(BT(J)-ST)
      WRITE(POUT,708) J,BT(J),FHEAT,SURFT(J)
708 FORMAT(1X,I4,5X,F10.3,4X,F10.6,5X,F10.5)
202 CONTINUE
709 FORMAT(15,F10.0)
568 STOP
      END
      SUBROUTINE MAXW(G,R,S,T)
C
C THIS SUBROUTINE CALCULATES THE THERMAL CONDUCTIVITY FROM A MAXWELL
C MODEL OF THE SOLID
C
      RATIO=R/G
      XN=1.-RATIO
      XD=1.+RATIO
      CONST=1.-XN*S/XD
      CPNST=1.+XN*S/XD
      T=G*CONST/CPNST
      RETURN
      END

      SUBROUTINE RECP(G,R,S,T)
C
C THIS SUBROUTINE CALCULATES THE THERMAL CONDUCTIVITY FROM A RECIPROCAL MEAN
C CONDUCTIVITY
C
      RATIO=R/G
      CONST=RATIO/((1.-S)*RATIO+S)
      T=G*CONST
      RETURN
      END

```

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