A HYDROTHERMAL MODEL OF THE ROOSEVELT HOT SPRINGS AREA, UTAH, USA

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SUMMARY - The hydrothermal system at Roosevelt Hot Springs Known Geothermal Resource Area (KGRA), located on the west side of the Mineral Mountains, Utah, is somewhat unusual in that it appears to be related to a young granitic intrusive and associated faults and fractures within a horst block. The shape and size of the heat source are based on gravity, seismic, and geological data while the details of the model of the circulation system are constrained by drilling information. The temporal evolution of the system seems to require continued intrusive activity after the latest dated surface flows (500 ka). The shallow part of the hydrothermal system appears to be controlled to some extent by the details of the permeability structure in the immediate vicinity if the high surface heat flow region.

1. INTRODUCTION

The Roosevelt Hot Springs Known Geothermal Resource Area (KGRA) has undergone active geothermal exploration and is the site of a 20 MW (electrical) power development. This paper describes the results of hydrothermal modeling undertaken to investigate the nature of the regional hydrothermal system. The thermal model is based on a model of the heat source developed from gravity analysis (Becker and Blackwell, 1993)

Location and Geographic Setting The Mineral Mountains are located along the eastern edge of the Basin and Range Province in southwest Utah. A geologic map (after Steven and Morris, 1983) is shown in Figure 1. Precambrian metamorphic rocks are exposed on the west side of the Mineral Mountains. Paleozoic and Mesozoic sedimentary rocks crop out on the northern and southern ends of the Mineral Mountains. Most of the Mineral Mountains are underlain by a Cenozoic quartz monzonite intrusive complex referred to as the Mineral Mountain batholith.

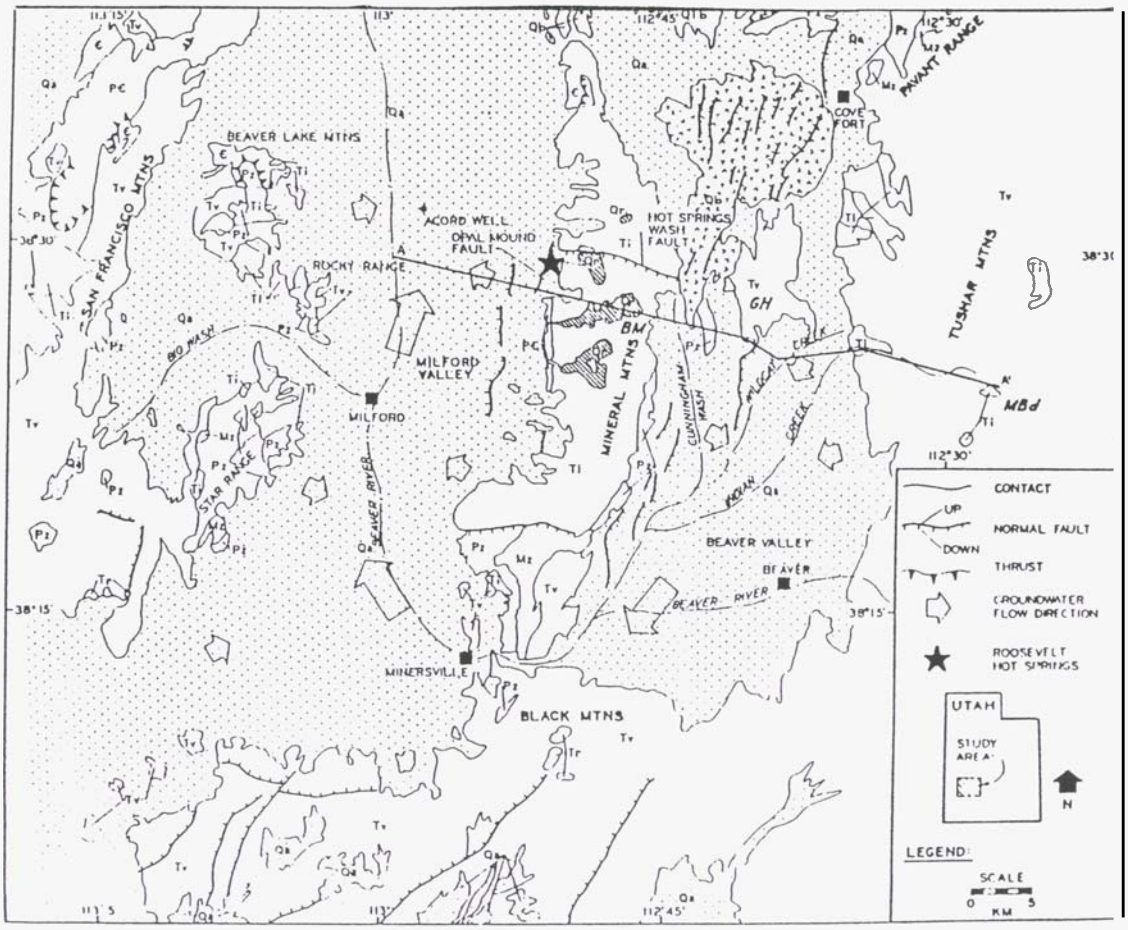

Figure 1. Geology of the Roosevelt Hot Springs area. Arrows show primary shallow groundwater flow directions. Abbreviations: BM-Bearskin Mountain, GH-Gillies Hill, MBd-Mount Baldy. Stippled area has surface heat flow >400 mWm^{-2}.

Nielson et al. (1986) dated initial intrusive activity at 27 Ma, with continued intrusion to about 12 Ma. Late Cenozoic rhyolite-basalt volcanism in the vicinity of the Mineral Mountains is represented by rhyolite flows and domes extruded at about 9 Ma at Gillies Hill (Evans and Steven, 1982), a quartz latite flow extruded at 7.9 Ma on the southeast side of the range (Sibbert and Nielsen, 1980), basalt activity from 1.0 to 0.3 Ma 15 to 20 km to the northwest of Roosevelt Hot Springs (Ross et al., 1982), and rhyolite domes and flows from 0.8 to 0.5 Ma along the crest of the Mineral Mountains (Evans and Nielson, 1982). The youngest volcanics in the region are basalts in the Cove Fort volcanic field northeast of the Mineral Mountains estimated by Nielson et al. (1986) to be Pleistocene to Holocene in age. The 0.5 Ma rhyolite flows may come from a parent magma chamber at a depth of 8 to 11 km.

Milford Valley, and the other surrounding valleys, are underlain by Late Cenozoic sediments. The Acord 1-26 well was drilled near the center of Milford Valley (see Figure 1) and penetrated about 1000 m of lacustrine clastics, 1100 m of Cenozoic sands, 300 m of volcanics, and 700 m of Cenozoic conglomerate above quartz monzonite and Precambrian metamorphics (Shannon et al., 1983).

The structure of the area is dominated by Basin and Range faulting. The Mineral Mountains are bounded on the west by a number of west-dipping normal faults, including the Opal Mound fault (Figure 1). The Hot Springs Wash fault is a major east-trending fault that cuts through the central part of the Mineral Mountains batholith and the geothermal field. Normal faults have also been mapped on the east side of the range by Steven and Morris (1983). Nielson et al. (1986) hypothesized west-dipping low-angle detachment faulting affecting the western half of the range. The Opal Mound fault and other faults on the west side of the range offset Quaternary alluvium. Micro-seismic activity has been recorded near Roosevelt Hot Springs and near Cove Fort, 40 km to the northeast (Nielson et al., 1986).

The hydrology of the Milford Valley was investigated by Mower and Cordova (1974) and that of Beaver Valley south of Gillies Hill was described by Mower (1978). Because of the difference in elevation between Beaver Valley and Milford Valley there may be cross-range flow under the Mineral Mountains. Smith (1980) modeled the regional ground water flow regime and found cross-basin flow to be unlikely assuming reasonable permeability anisotropy. He did not investigate thermal effects, which may enhance the possibility of cross-range flow.

The Geothermal Field Roosevelt Hot Springs KGRA is a fluid dominated geothermal field centered near the intersection of the E-W trending Hot Springs Wash fault and the N-S trending Opal Mound fault (Figure 2). Large amounts of siliceous sinter and altered alluvium attest to a greater discharge in the past (Ward et al., 1978). Production is currently limited to the area east of the Opal Mound fault and west of the range front as shown in Figure 2. Wells drilled west of the Opal Mound fault have been hot, but dry. The primary reservoir appears to be confined to fractures within the small horst east of the Opal Mound fault. Typical production depths range from 820 m to 2286 m with a maximum reported temperature of 271°C (Benoit and Butler, 1983). Fluid chemistry studies are consistent with the main deep reservoir having a temperature near 285°C (Capuano and Cole, 1982). Isotopic studies indicate that the water is of meteoric origin (Rohrs, 1980).

Previous Geophysical Investigations There have been several studies of the thermal regime near Roosevelt Hot Springs. A surface heat-flow map adapted from Wilson and Chapman (1980) is shown on Figure 2. They estimated an anomalous surface heat loss of 64 MW.

Robinson and Iyer (1981) analyzed teleseismic P-wave delays in the Mineral Mountains area. A three dimensional inversion of travel time residuals delineated an anomalous low velocity zone that extends from the upper mantle to within 5 km of the surface and has a velocity 2 to 7% lower than the surrounding rock. They estimated a density contrast of -150 kg m^{-3} for the zone based on Birch's Law. However, they attributed the low velocities to a high-temperature zone containing a small fraction of partial melt rather than compositional differences.

The gravity field in the vicinity of the Mineral Mountains was analyzed by Carter and Cook (1978). Becker and Blackwell (1993) used this data to isolate a gravity anomaly that coincides with the volcanism and

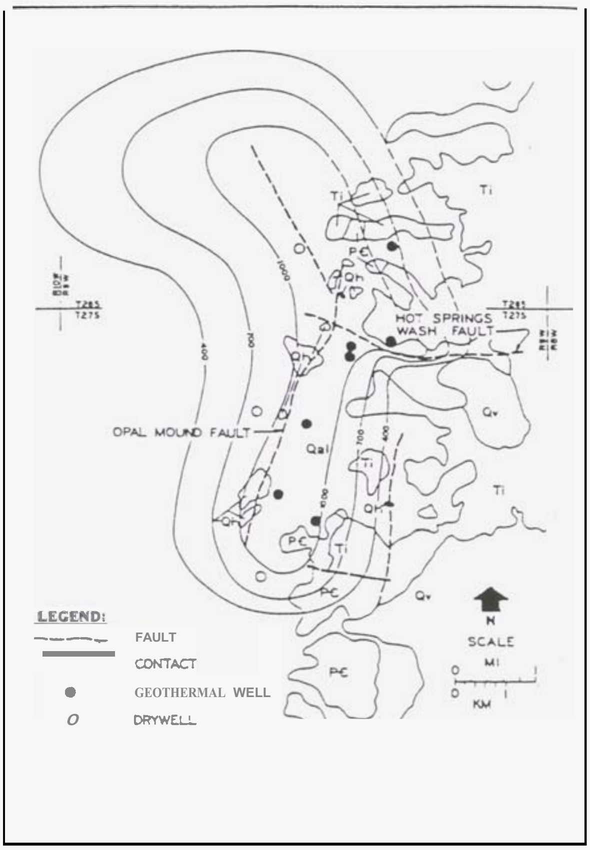

Figure 2. Geologic map of the Roosevelt Hot Springs geothermal field showing locations of productive geothermal wells (from Benoit and Butler, 1983). Surface heat flow contours are from Wilson and Chapman (1980).

hydrothermal activity in the range. They interpreted the interpreted anomalous body as a magma chamber. The interpreted mass has 1) a density contrast with the surrounding rock of approximately -150 kg m^{-3}, (which is in agreement with seismic evidence) 2) a roughly cylindrical shape which extends from perhaps as deep as the Moho to within 4-6 km of the surface, and 3) a typical diameter of 15 km. The depth estimates for the body are consistent with the depths of origin for the rhyolites found at the range crest as inferred from geochemical evidence. The body is interpreted to be the source of heat for the Roosevelt Hot Springs, and may be related to the source of the Quaternary rhyolites. A comparison of the gravity interpretation to the seismic P-wave delay interpretation is shown in Figure 3. Unlike caldera terrains, granite is exposed at the surface so the near-surface composition is clearly not involved as the cause of the gravity or seismic anomalies.

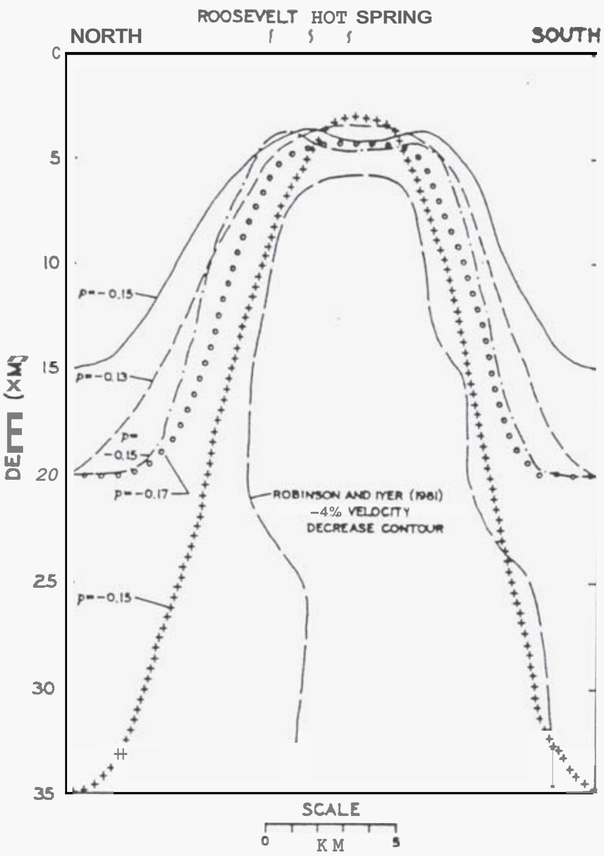

Figure 3. Various geometries inferred for the anomalous body. Shapes based on various density contrasts and depths to base are shown as is the 4% velocity decrease contour from teleseismic P-wave delays (Robinson and Iyer, 1981).

2.0 MODELING METHODOLOGY

Program Used The CCC (Conduction, Convection, and Consolidation) pro-gram, developed by Lawrence Berkeley Laboratories (Mangold and Lippmann, 1980), was used to develop a two-dimensional model of the hydrothermal system. Sorey et al. (1978) used the CCC program to analyze the hydrothermal system in Long Valley caldera, California for example. The program uses an integrated finite difference technique to solve the coupled mass (fluid flow) and energy (heat flow) balance equations. The matrix values of porosity, permeability, specific storage, thermal conductivity, heat capacity, and density, and the fluid density, viscosity, specific heat, and fluid compressibility are discussed briefly below. More details may be found in Becker and Blackwell (1993).

The Two-dimensional Model A more or less east-west vertical cross section was used. The line of profile, as shown on Figure 1, was chosen to be parallel to either the gradient of the piezometric surface,' where known, or the topographic contours. A diagram of the model is given in Figure 4. The capital letters indicate the various areas in the model included in Table 1. The base of the model profile was chosen to be at an elevation of -6.0 km below mean sea level (m.s.l.), giving the profile a maximum thickness of almost 9 km at the eastern end. This thickness was chosen so that several kilometers of the top of the hypothesized intrusion would be included in the model.

The configuration of the deep intrusion was based on the gravity modeling results. Milford and Beaver Valleys were assumed to be filled with no more than two kilometers of sediment. The Mineral Mountains batholith was assumed to be limited mainly to areas east of the Opal Mound fault zone and west of Gillies Hill based on the interpretation of Evans and Nielson (1982, p. 1135). The Opal Mound fault zone was assumed to be one km in width, and essentially vertical. The fault zone was assumed to extend to an elevation of -3.0 km (m.s.l.). Special attempts were made to provide as much detail as possible for the immediate area of Roosevelt Hot Springs.

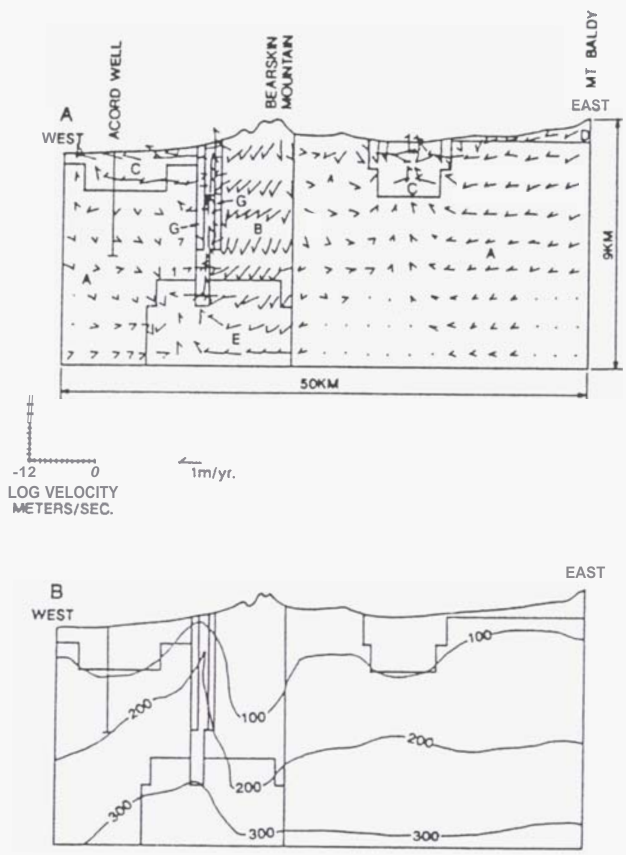

Figure 4. Flow and temperature regime computed for pre-ferred model, including paired low-permeability bound-aries adjacent to the Opal Mound fault, after 148,000 y.

Boundary Considerations The boundary conditions chosen for the model included no-flow along the vertical sides and a no-flow along the base of the model with a background heat flux of 80 mWm^{-2}. The upper surface of the model was assumed to be a constant pressure and temperature boundary. The constant pressure boundary corresponds to a fixed water table surface and implies recharge in the mountains and discharge at the low points in the valleys.

The initial temperature and pressure at each node must be specified prior to starting the simulations. The initial pressure estimates need not be refined because the pressure adjusts rapidly to the model geometry compared to length of the simulations. The initial temperatures were established using a uniform gradient of 40°C km^{-1}.

Rock Properties The values chosen for rock permeabilities, densities, thermal conductivities, specific storage, and specific heat were gathered from published values, measured values in the Mineral Mountains vicinity, values computed from previous gravity modeling, or values representative of similar lithologies. For conciseness only the permeability is discussed here.

The ranges of intrinsic permeabilities used in the model are shown in Table 1. Most of the material represented in the model is crystalline rock. The range of permeability of a sample of unfractured granite under pressure is 10^{-16} to 10^{-20} m^2 with most measurements less than 10^{-19} m^2 (Brace et al., 1968). The higher permeability values were assigned at pressures of less than 3 x 10^6 Pa, corresponding to depths of less than 500 m. Fractured crystalline rock displays a large range in permeability. Though there are relationships defining permeabilities of individual fractures, the emphasis of this model is on gross properties. It is assumed that the fractured rock mass behaves, on this large scale, as an equivalent porous media. Fryer et al. (1981) suggest a range of permeability of 10^{-14} to 10^{-18} m^2, with a relative decrease in permeability of 20 to the 1.5 power with each 3800 m increase in depth. They estimated that typical flow systems can extend to depths of 7.5 to 10 km. Brace (1984) also summarized the available data on in-place permeability of fractured crystalline rocks as typically less than 10^{-15} m^2, except in the case of highly fractured rock masses.

The typical permeabilities assumed for the crystalline rocks in the models ranged from 10^{-15} to 10^{-18} m^2. A decrease in permeability with depth was assumed in all cases. The permeability in severely fractured fault zones was assumed to vary between 10^{-12} and 10^{-15} m^2. For certain simulations, the vertical permeability of the Opal Mound fault was assumed to be 1000 times the horizontal permeability. The horizontal permeability of the sediment in the Milford and Beaver graben was initially estimated to lie in a range of 10^{-12} to 10^{-13} m^2. These values are within the ranges for silts and silty sands. The ratio of horizontal to vertical permeability was assumed to be three for the graben fill. The permeability values assumed here are generally consistent with values used in a recent model of the immediate vicinity of the Roosevelt Hot Springs by Faulder and Shook (1991).

TABLE 1. Rock Properties Used in Preferred Hydrothermal Model (Tertiary volcanics omitted).

PARAMETER	MATERIAL					
	Basement Rock	Min.Mtns Batholith	Cenozoic Sediment	Deep Intrusion	Opal Mound Fault	Fault Edges
Thermal Cond. ($W\ m^{-1}\ K^{-1}$)	3.2-2.1	3.2-2.1	0.625 V. 1.250 H.	3.2-2.1	3.2-2.1	3.2-2.1
Porosity (%)	1	1	20	0.1	1-2	1
Specific Storage (m^{-1})	$3x10^{-7}$	$3x10^{-7}$	$5x10^{-7}$	$1.9x10^{-7}$	$3.8x10^{-7}$	$3.8x10^{-7}$
Permeability (m^2)	10^{-16} to 10^{-18}*	10^{-15} to 10^{-16}*	10^{-13} V/H=1/3	10^{-16}	10^{-12} V/H=100	10^{-18}

Heat Flux at base of model starts at 80 mW m^{-2}, rising to 150 mW m^{-2} as time progressed.

***Permeability decreased with depth.**

Porosity and density (not listed) have little effect on the simulation. The values for thermal conductivity, are also summarized in Table 1. The specific storage of the rocks (Table 1) was computed as described by Becker and Blackwell (1993). The values are grouped around 10^{-6} m^{-1} and ranged from 5 x 10^{-6} m^{-1} for the sediments filling the grabens to 3 x 10^{-7} m^{-1} for crystalline rock. The values for the assumed specific heats were all taken from representative values for similar materials. The values range from 880 J/kg-C to 800 J/kg-C.

Fluid Properties The equation of state for the fluid was defined within the CCC program. The calculated density is close to the values given in the steam tables of Keenan et al. (1978) for the typical temperatures and pressures in this model. The equation used does not allow for two-phase fluid or for the effect of dissolved solids or gases.

3.0 HYDROTHERMAL MODELING RESULTS

Numerous simulations of the system were performed attempting to match the conditions observed at the geothermal field. A variety of assumptions and geometries were tested, including: various permeabilities in the Opal Mound fault zone; high and low water table conditions in the Mineral Mountains; the presence of Lake Bonneville filling Milford Valley; and various permeabilities assigned to the sediments filling the grabens. None of the model runs recreated the high temperatures in the near-surface unless the modeling simulated the presence of the inferred intrusion and low-permeability vertical boundaries adjacent to the Opal Mound fault zone. The incorporation of these features resulted in a satisfactory match to known conditions. In this preferred model, two low-permeability zones bordering the Opal Mound fault are introduced as shown in Figure 4. These zones are assumed to have a permeability of 10^{-18} m^2 and to extend to depths of 10^{-18} m^2 and to extend to depths of greater than 4 km (see Table 1). This model is the preferred representation of the system and is the only run which will be discussed at length.

Hydrothermal deposits are found along the Opal Mound fault and along a fault roughly corresponding to the eastern boundary of the thermal anomaly (Figure 2). It is likely that the temperature and pressure changes outside of the permeable fault zone would favor the deposition of silica and other material which would lead to low permeability "boundaries" along the fault zone. Sporadic movement along the fault would tend to maintain its permeability. Ross et al. (1982) found evidence for higher resistivity zones bounding the low resistivity zone corresponding to the Opal Mound fault. This condition may be related to deposition of material in zones peripheral to the fault. Lippmann and Bodvarsson (1985) suggested a similar cause for the low-permeability zones at the edges of the Heber, California geothermal field.

The results of the simulation after 148,000 years of simulated time are presented as Figure 4. The flow is largely downward beneath Milford Valley and there is a strong upward flow in the fault zone. The overall appearance is that of a paired convection system beneath the valley and the Mineral Mountains, with the total flow rising along the fault zone. An expanded plot of the temperatures in the vicinity of the fault zone is given in Becker and Blackwell (1993). Temperatures at maximum drilled depths in the geo-thermal field are over 220°C and are 140°Cjust below the surface.

After 400,000 y the flow regime decays to a more widespread upward flow beneath much of Milford Valley regardless of the conditions in the fault zone (Becker and Blackwell, 1993). The flow in the fault zone, though still upward, decreased in magnitude and the temperatures rose below the valley and decreased in the fault zone. The temperature contours in the vicinity of the fault zone are generally vertical, again indicating that the fault zone approximates a boundary between flow regimes. The temperatures in the fault zone decrease by as much as 50°C from the 148,000-year results. Temperatures at the location and depth of the Acord well are near 260°C, about 60°C higher than temperatures computed in the simulation at 148,000 y. No cross-basin flow was observed, though eastward flow approaches stagnation under the east flank of the Mineral Mountains.

Other runs were also performed to test the sensitivity of the parameters used in the preferred model. The results appear to be insensitive to the water table height, including the presence of Lake Bonneville, and the permeability of the graben fill. The results were not reproducible, however, if the permeability of the deep intrusion was reduced to 5 x 10^{-18} m^2 or if the deep intrusion was omitted from the model.

4.0 DISCUSSION

Other studies have explored the convective systems associated with specific geothermal systems. Many of these studies focus on the details in and around the systems and so are local in scale (on the order of kilometers). The emphasis is on the possible effects of exploitation based on the state of the initial system as inferred from modeling (i.e. Lippmann and Bodvarsson, 1985). The model presented here is a regional scale model constrained by geophysical and geological observations. The flow regime displayed in the early stages of the preferred simulation may be used as an approximation to the native flow conditions at Roosevelt Hot Springs.

The regional nature of this modeling suggests that a zone of elevated temperatures associated with a possible intrusion at depth are required to achieve the observed temperatures in the geothermal field. Deep cross-range circulation of meteoric water in areas of normal Basin-and-Range heat flow, as proposed for the Socorro, New Mexico geothermal system by Barroll and Reiter (1990), appears insufficient to cause the observed anomalous temperatures at shallow depths. The gravity interpretation indicates that the Roosevelt Hot Springs area is in fact underlain by an abnormally hot zone, perhaps containing a small component of partial melt. This zone extends upward from as deep as the Moho to a depth of only 5 km. The lateral extent of this body ranges from about 15 km in the east-west direction to 10 km in the north-south direction. This deep body is approximately centered under the geothermal field.

The time scale for the decay of the system is unexpectedly short compared to the age of the Quaternary rhyolites present at the crest of the range. The area has had a significant history of igneous and volcanic activity continuing from the Miocene to the Pleistocene-Holocene (Nielson et al., 1986). Thus the presence of a periodically replenished heat source at depth is certainly conceivable and seems to be required by the results of the simulations. However, further refinement of the model to consider effects of the heat of solidification in a partial melt as was done by Sammel et al. (1988) may be useful.

The modeling presented here suggests significant permeability extending to mid-crustal depths below the geothermal field. The retention of this level of permeability would require repeated movement along pathways such as the Opal Mound and Hot Springs Wash faults. The apparent offset on Quaternary sediments and microseismic activity suggest tectonic activity on the faults in the Roosevelt Hot Springs vicinity continued perhaps through the Quaternary to the present. The effect of the low-angle faults on the hydrothermal system is not clear, but is probably significantly less than high-angle faults such as the Opal Mound fault. Finally, additional exploration for permeable zones beneath the sediments filling Milford Valley might result in the discovery of additional geothermal resources. Thermal indications are masked beneath the sediments by large flow of non-thermal water in the sediments.

5.0 ACKNOWLEDGMENTS

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