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CONCEPTUAL GEOLOGIC MODEL AND NATIVE STATE MODEL OF THE ROOSEVELT HOT SPRINGS HYDROTHERMAL SYSTEM

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ABSTRACT

A conceptual geologic model of the Roosevelt Hot Springs hydrothermal system was developed by a review of the available literature. The hydrothermal system consists of a meteoric recharge area in the Mineral Mountains, fluid circulation paths to depth, a heat source, and an outflow plume. A conceptual model based on the available data can be simulated in the native state using parameters that fall within observed ranges. The model temperatures, recharge rates, and fluid travel times are sensitive to the permeability in the Mineral Mountains. The simulation results suggests the presence of a magma chamber at depth as the likely heat source. A two-dimensional study of the hydrothermal system can be used to establish boundary conditions for further study of the geothermal reservoir.

INTRODUCTION

The Roosevelt Hot Springs (RHS) hydrothermal system was the site of an active exploration program starting in 1974. A 500°F liquid-dominated reservoir was discovered through exploration drilling in 1975. The Roosevelt Hot Springs Unit (RHSU) was formed in April 1976 and was the first geothermal unit approved by the United States Department of Interior. A 25 MW_e geothermal power plant started operations in 1984. The location of the study area is shown in Figure 1.

The Roosevelt Hot Springs area has been used as a natural laboratory for the development and testing of geothermal exploration and evaluation methods, involving geologic, geophysical, geochemical, and reservoir testing. A literature review for the RHS system reveals over 180 geoscience titles. These many sources were used to develop a conceptual geologic model of the hydrothermal system.

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A two-dimensional reservoir model of the hydrothermal system is used to investigate the conceptual model and the physical constraints of the system. The native state simulation study tests the conceptual geologic model and establishes reservoir boundary conditions. As the simulation study progresses, the conceptual geologic model provides a reference for adjusting reservoir parameters.

GEOLOGY

The RHS geothermal system is located on the eastern edge of the Basin and Range physiographic province and at the transition between the Colorado Plateau and the Basin and Range. The geothermal system lies to the west of the batholith of the Mineral Mountains, the first range west of the Wasatch Front. The Mineral Mountains are a north-south trending horst bounded by Basin and Range normal faults. A geologic map of the area is given in Figure 2.

Eruptive History

The Mineral Mountains intrusive complex has a history of magmatic activity since the Oligocene time (Neilson et al., 1986). The oldest phase began about 25 Ma with intrusives into Precambrian rocks. This pluton was then intruded by the main intrusive complex about 22 Ma. About 9.0 to 9.6 Ma an igneous sequence was emplaced. The earliest volcanic activity occurred 7.9 Ma along the west side of the range. The final volcanic episode started in the Twin Peaks volcanic complex about 2.7 Ma with the eruption of rhyolite domes and widespread basalt flows. The last rhyolitic volcanism occurred between .8 and .5 Ma and resulted in twelve domes in the central Mineral Mountains and the Bailey Ridge rhyolite flow just east of the reservoir. Chemical similarity of all the domes suggests they were derived form the same magma source (Ward et al., 1978).

Structure

The structural geology of the RHS has been studied by many workers. A brief description of the features that follows draws upon the work of Nielson et al. (1978), Ward et al. (1978), Bruhn et al. (1982), Ross et al. (1982), Nielson et al. (1986), and Nielson

(1989). The commercial geothermal reservoir is closely associated with the Negro Mag and Opal Dome Faults. Structural features are important in controlling the reservoir characteristics and boundaries.

The Negro Mag Fault is an east-striking, high angle, oblique slip with significant right lateral shear fault. This range cutting fault is the major driving fault defining local active structures and is active into the deep basement. The Negro Mag Fault is located along the axis of a complex graben structure 4 miles across. This graben forms a low in the crest of the Mineral Mountains, separating a Pleistocene rhyolite dome complex to the south from lower and more dissected ground containing no rhyolite domes to the north. The Bailey Ridge rhyolite appears to have erupted from faults associated with this graben, suggesting the structure has been present since at least the early Pleistocene.

The highly conspicuous Opal Mound Fault is a north-south normal fault marked by alluvial scarps, surface alterations, and opaline deposits which attest to geologically recent activity and extensive leakage of the reservoir along this feature. The Opal Dome Fault separates a graben to the east from a narrow horst to the west.

Low- to moderate-angle denudation faults occurs throughout the Mineral Mountains, but are most common in the geothermal area. The faults dip between 5° and 35° to the west with an estimated maximum depth of formation of 16,000 feet (Bruhn et al., 1982). These low-angle faults developed after the emplacement and consolidation of the Tertiary pluton complex and pre-date rhyolite domes and flows dated at 0.5 Ma.

The older, low-angle faults consist of up to 650 feet zone of cataclasis separating rocks of the Mineral Mountains intrusive complex from overlying sedimentary rocks. The Cave Canyon Fault represents this style of faulting.

A second series of listric normal faults occurs cutting principally rocks of the Mineral Mountains intrusive complex. The Wildhorse Canyon and Salt Cove Faults are representative of this style of faulting. The Wildhorse Canyon Fault is a continuous feature on the west side of the Mineral Mountains. This feature contains a number of NW, high angle cataclasite zones up to 12 feet thick in hills south of Big Cedar Cove. The Salt Cove Fault is a similar, parallel structure, east of the Wildhorse Canyon Fault.

The joint system through the central Mineral Mountains is relatively homogenous and consists of three major joint sets. Two sets of steeply dipping, sub-orthogonal extension joints trend northward and eastward, occurring roughly parallel and perpendicular to the strike of the contacts between igneous and country rocks. The joint spacing varies from 3 to 95 feet to less than two inches in areas of intense faulting. A third joint set consists of gently to moderately westward dipping joints generally having smooth planar surfaces with a joint spacing varying from greater than 3 feet to 4 inches in highly faulted areas. The joint system in the Precambrian rocks is similar to the pluton.

Geophysics

The surface heat flow map of the area clearly shows the location of the shallow geothermal reservoir, (Figure 3). Surface heat flow above the known reservoir is greater than 1000 mW/m^{-2}, with a large plume extending to the northwest, (Wilson and Chapman, 1980). Continuation with depth of the heat flow data shows an eastward extension along the Negro Mag fault. The large plume northwest of the intersection of the Negro Mag and Opal Dome Faults is associated with outflow from the geothermal reservoir. The regional heat flow is 92 mW/m^{-2}, while heat flow measured at depth from the Acord 1-26 well was 146 mW/m^{-2} (East, 1981).

The total aeromagnetic intensity residual map of the RHS area shows the dominance of east-west features that cut the Mineral Mountains and extend east into the Beaver Valley, reflecting the structure at depth.

Gravity modeling and filtering by Becker (1985) indicates an anomalous gravity low centered 13,000 - 20,000 feet below the reservoir with a density contrast of approximately -.15 g/cc. This result closely corresponds to work by Robinson and Iyer's (1981) investigation of P-wave structure of the crust and uppermost mantle. Their work showed a clear pattern of relatively low velocity (5 to 7 per cent less than the surrounding rock) material extending up from the upper mantle to a depth of about 16,000 feet under the west side of the Mineral Mountains. This plume is centered near the geothermal area, but extends to the north and south at depth. The degree of velocity change modeled would indicate a temperature increase of about 1080° to 1,530°F, indicating for typical crustal rocks some degree of melting.

Pre-production microseismic monitoring detected several episodic east-west swarms south of the Negro Mag Fault, (G. Zandt and D. Nielson, written communication). Focal depths were clustered at two distinct depths of 10,000 feet and 26,000 feet. The interval in-between was aseismic. The microseismicity demonstrates the Negro Mag graben system is still active, see Figure 2.

Geochemistry
The thermal waters were by characterized Capuano and Cole (1982) as a dilute sodium chloride brine, with approximately 7000 mg/l total dissolved solids. The Na-K-Ca and SiO_2 geothermometers indicate deep geothermal temperatures of 466° and 550°F for the Roosevelt seep and deep well fluid samples, respectively. Analysis of fluid samples from wells and springs in the RHS area suggests that they are derived from a common reservoir source, with variations due largely to mixing with shallow groundwater. The least mixing of thermal fluids with the shallow groundwater occurs for wells in close proximity with the Negro Mag Fault (Vuataz and Goff, 1987).

The stable isotopic composition of the thermal fluids indicates they are of meteoric origin with the water derived primarily from the Mineral Mountains with perhaps some contribution from the Tushar Mountains to the east, (Bowman and Rohrs, 1981). The thermal fluids have a δD value of -116 and a $\delta^{18}O$ value of -13.7. Stable isotopic studies of the age of Great Basin thermal waters by Flynn (1990) suggest the RHS waters may be 10,000 to 15,000 years old, (Flynn, personal communication). Tritium content of the thermal fluids are very low, ≤ 1 TU, (Vuataz and Goff, 1987).

Several attempts have been made to date the age of the hydrothermal system. Paleo-magnetic dating of opal from the Opal Dome suggest a minimum age of 12,000 years and a length of activity from 35,000 to 70,000 years. There were correlative problems and the maximum age could be as old as 350,000 years, (Brown, 1977). Hydration dating of obsidian in the alluvium yielded ages of 220,000, 257,000, and 330,000 years, (Bryant et al., 1977). These dates are maximum times for obsidian hydration in alluvium based on a hydration rate at 40°F, (Parry et al., 1978). The range of all these dates are poorly constrained.

Hydrology
The shallow alluvial aquifer west of the RHS area has a potentiometric surface dipping northwest toward the Milford Valley (Mower and Cordova, 1974,). Flow in the aquifer is stratigraphically controlled by flow in horizontal, permeable Quaternary alluvial deposits of sands and gravels. Concentrations of boron and chloride in the aquifer clearly show a large outflow plume from the high temperature thermal system trending northwest down the hydrologic gradient (Vuataz and Goff, 1987). The outflow plume is leaking over the Opal Dome horst, with the plume centered at the intersection of the Opal Dome and Negro Mag faults. The Opal Dome horst acts as a hydrologic barrier to the geothermal reservoir.

A aquifer test was made in well 26-9-18. The test results indicate a permeability of 1560 mD, assuming an aquifer thickness of 320 feet (Vuataz and Goff, 1987). The thickness of the principal aquifer west of Negro Mag Wash varies from greater than 500 feet west of the reservoir to 100 to 300 feet in the center of Milford Valley (Mower and Cordova, 1974).

The central Mineral Mountains receive an average annual precipitation of 16 to 25 inches, at elevations of 6400 to 8600 feet (Mower and Cordova, 1974). Vautaz and Goff concluded the shallow aquifer was recharged by precipitation in the Mineral Mountains with a minimum residence time of 70 years. The total recharge in the Mineral Mountains was estimated by Smith (1980) at 7650 acre-ft/year. An estimated 650 acre-ft of this total is available for the RHS area. The elevation difference of more than 980 feet between Beaver Valley and Milford Valley may allow some component of inter-basin flow.

RESERVOIR ENGINEERING

The primary reservoir has a reported fluid volume of 19 billion barrels from two long-term flow tests (three months and nine months) of a single well, RHSU 54-3 (Kerna and Allen, 1984). Assuming a primary reservoir dimension of 10,000 by 23,000 feet (from heat flow data) and 10,000 feet deep (2500 feet below the deepest well), an average total porosity of 4.7% is calculated. Well test information indicates the wells are able to flow from 300 Klbm/hr to over 1,000 Klbm/hr (Butz and Plooster, 1979) or receive injectate at rates of up to 1,850 Klbm/hr (Rosser et al., 1984), indicative of a highly fractured reservoir .

Applying the concept of a permeability window (Forster and Smith, 1989) and assuming a bulk permeability of ~.01 mD representative of granite, maximum temperatures will from a fracture permeability of ~1000 mD. This estimate is in agreement with the qualitative assessments of reservoir permeability above.

The pre-exploitation reservoir pressure in RHSU 14-2 is 1365 psia at 2950 ft below the surface, (Butz and Plooster, 1979). This pressure is assumed to be representative of the geothermal reservoir.

Temperature data was compiled from several sources, (Wilson and Chapman, 1980, Shannon et al., 1983). A comparison of several wells in the RHS area show three types of profiles, two conductive and one convective, (Figure 4). The Acord 1-26 well measured a bottomhole temperature of 446°F at 12,645 feet, and a gradient at depth of 2.95°F/100 feet. The Acord 1-26 profile was extrapolated from these measurements. This profile closely matches the RHSU 24-36 profile to about 3000 feet, with RHSU 24-36 showing hotter temperatures and a slightly higher gradient below this depth, suggesting a closer proximity to a heat source. RHSU 24-36 is located north-east of the known geothermal reservoir and at a higher elevation. The RHSU 9-1 and RHSU 52-21 profiles are intermediate in temperatures, with a similar gradient as the Acord, but hotter. These two wells are located in close proximity to the geothermal reservoir but are non-productive. The RHSU 14-2 and RHSU 72-16 profiles show a convective system at 500°F. Extrapolation of the RHSU 24-36 and RHSU 14-2 profiles indicates an intersection at approximately 12,000 feet. This depth is similar to the shallow depth of micro-seismicity.

CONCEPTUAL MODEL

A conceptual model of a hydrothermal system needs to address four components of a dynamic flow system: fluid recharge, fluid circulation paths, a heat source, and an outflow plume. The geothermal reservoir at Roosevelt Hot Springs can be modeled as a dynamic flow system into and out of a intensively fractured graben.

Meteoric recharge is primarily from precipitation on the crest and west flank of the central Mineral Mountains. A small contribution from inter-basin transfer from Beaver Valley to the east to Milford Valley through the Negro Mag graben structure is possible.

Downward circulation of recharge waters occur in the Mineral Mountains east of the geothermal reservoir. The extensive joint and fracture system coupled with the complex east-west graben associated with the Negro Mag fault allows meteoric waters to circulate to depths controlled by the presence of open fractures. Microseismicity suggests open fractures may exist at depths of 10,000 and 26,000 feet. The waters heat up at depth and flow hydrologically down gradient until they encounter deep-seated normal faults such as the Opal Dome and/or Negro Mag Faults. Upwelling occurs along these high permeability features, with lateral flow into the Opal Dome graben. The intersection of the Opal Dome and Negro Mag grabens and the low angle faults provides an intensively fractured geothermal reservoir for the thermal fluids. Circulation in the reservoir takes place in the complex, well- developed, three-dimensional permeability structure.

Two sources of heat for the hydrothermal system need to be considered: deep circulation in an area with high regional heat flow, or a shallow magma chamber. Both possibilities can be investigated through simulation studies. The preferred heat source at RHS is the plume of partial melt material under the western, central Mineral Mountains. As modeled by Robinson and Iyer (1981), and Becker (1986), this plume extends from a depth of about 16,000 feet to at least as far as the upper mantle. The 16,000 feet depth has been inferred by Nielsen et al. (1986) to be at approximately the brittle-ductile boundary. Thus, open fractures could be supported to this depth and provide permeable pathways for convecting water. Cooling of the magma chamber with time could result in the development of deeper fractures.

The outflow of thermal water occurs over the Opal Dome horst and is centered at the intersection of the Opal Dome and Negro Mag faults. A plume of hot water then mixes and dilutes with cooler water in the shallow aquifer as it flows down the hydrologic gradient to Milford Valley.

A schematic of the conceptual model is presented in Figure 5. The location of select wells is projected on the figure. Key features of the conceptual model are identified.

The deep structure of the Roosevelt hydrothermal system is a matter of conjecture, with the possibility that a hotter, deeper system may exist below the known reservoir. The microseismic gap between 10,000 and 26,400 feet suggests the presence of an intermediate impermeable seal, allowing a second, deeper, hotter system to develop with leakage between the two systems occurring through the deep-seated normal faults. A hotter system is also suggested by extrapolation of the temperature profile data with an intersection at about 12,000 feet. A continuation of the RHSU 24-36 profile suggests a deeper system could exist with some leakage into the shallow system, with insufficient residence time be detected by geochemistry. The presence of deeper, hotter systems below geothermal reservoirs has been documented for The Geysers (Walters, 1988) and suggested for Valles Caldera (Hulen et al., 1989) and Dixie Valley (Doughty et al., 1990).

NATIVE STATE SIMULATION OF THE HYDROTHERMAL SYSTEM

A two-dimensional, east-west vertical cross-section was used to simulate the conceptual geologic model. The model length is 41,250 feet covering the crest of the Mineral Mountains to the center of Milford Valley, a distance with a surface relief of 3,200 feet. A numerical grid of 20 by 8 blocks was used in these simulations. The ground surface defines the top of the model with a depth of 19,800 feet below the ground surface. The simulation studies were performed using TETRAD, a fully implicit, finite difference geothermal simulator, (Vinsome, 1990). The simulator has been validated against a number of geothermal problems and yields results comparable to those published elsewhere (Stanford Special Panel on Geothermal Model Intercomparison Study, 1980).

A basal heat flux of 150 mW/m^2 was assumed. The initial temperature distribution was based on the Acord 1-26 temperature profile. The Mineral Mountains side was modeled as a no-flow boundary at depth. A constant pressure boundary on the Milford Valley side, with a pressure gradient assigned according to the temperature profile of the Acord 1-26 well was used. The high water table configuration of Smith (1980) was used. This configuration closely matched the elevations of springs in the Mineral Mountains noted by Vuataz and Goff (1987). Meteoric recharge at a temperature of 40°F was approximated using steady state aquifers at the water table to the east of the geothermal reservoir. The actual recharge rate is unknown, but a range of 1% to 5% of the annual precipitation in the Mineral Mountains is considered to be realistic.

The permeability of the Mineral Mountains massif was estimated from the joint system distribution of the Central Mineral Mountains. The fracture width is unknown so a width of 100 µm was used. A fracture width of 100 µm and a joint spacing 3 to 100 feet, results in fracture permeability range from Reiss (1980) of 167 mD to 5.6 mD, respectively. A matrix permeability of 0.01 mD was assumed and calculating the system average permeability by the method of Aguilera (1980) results in an average permeability of 0.027 mD to 0.010 mD. Values of 0.5 mD and 0.05 mD were used for the mountain massif. These values are higher than calculated above, but were used to establish an upper limit. The massif is located east of the reservoir and extends to a depth of about 15,000 feet. The average permeability in the Mineral Mountains influences the fluid travel time from meteoric recharge to the geothermal reservoir. A porosity of 1% was used.

The structural complexity of the geothermal reservoir was reduced by assuming the Opal Dome Fault is a high permeability zone extending to about 17,000 feet. This simplification ignored the role of the Negro Mag Fault as the major driving fault defining local active structures. This deep seated high permeability zone is a synthesis of the Opal Dome and Negro Mag faults features. A permeability of 500 mD, reducing with depth to 10 mD, was used for this feature. The Opal Dome horst was assigned a permeability of 0.001 mD.

The shallow aquifers west of the geothermal reservoir was given a permeability of 1000 mD and 500 mD, decreasing with depth to 0.01 mD. The horizontal and vertical permeability structure used in the study is presented in Figures 6 and 7.

A straight line relative permeability relationship was used with a residual water saturation of 25% and a residual gas saturation of 1%.

Results

Simulation runs were made to adjust the model, calibrate the location of the water table, and investigate model sensitivity to the permeability structure. The model is extremely sensitive to the permeability in the mountain massif. A uniform basal temperature of 639° F, based on the Acord 1-26 profile was used. A tracer component with the same properties of water was used to track recharge flow paths and travel times.

A model run using a permeability of 0.5 mD in the mountain massif clearly shows the cooling of the mountain massif, the upwelling in the high permeability fault zone, and the shallow outflow plume. Thus the essential features of a hydrothermal system were present. The model temperatures and pressures reached steady-state conditions after about 30,000 years. The temperatures in the shallow geothermal reservoir range from 169° to 254°F (Figure 8) and had a fluid travel time of about 800 years. The steady state recharge rate was 35 Klbm/hr. This implies an annual meteoric recharge of 1.11 inches (113 acre-ft) for a model width of 3,000 feet.

A second simulation used a reduced permeability in the mountain massif of 0.05 mD. The model temperatures reached steady-state conditions after 50,000 years. The temperatures in the shallow reservoir range from 204° to 353°F (Figure 9) and had a fluid travel time of about 8,000 years. The steady state recharge rate was 12 Klbm/hr, implying an annual meteoric recharge rate of .38 inches.

The 0.05 mD case generally results in simulated temperatures within 30°F measured in outlying wells Acord 1-26, RHSU 24-36, and RHSU 9-1. However, the temperature in the geothermal reservoir is several hundred degrees cooler than measured. It appears that the high temperatures observed in the geothermal reservoir are unlikely to be due solely to a high regional heat flow and suggests the presence of a magma chamber.

The simulated recharge rate implies a recharge of about 2% of the annual precipitation in the central Mineral Mountains. This is a reasonable value for the arid climate in the study area.

CONCLUSIONS

A conceptual model based on available data can be simulated using values of parameters that fall within observed ranges.

Model temperatures, recharge rates and fluid travel times are sensitive to the permeability in the Mineral Mountains.

The boundary conditions for further simulation of the geothermal reservoir can be established by two-dimensional native state simulation.

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Figure 1 Location of Roosevelt Hot Springs hydrothermal system.

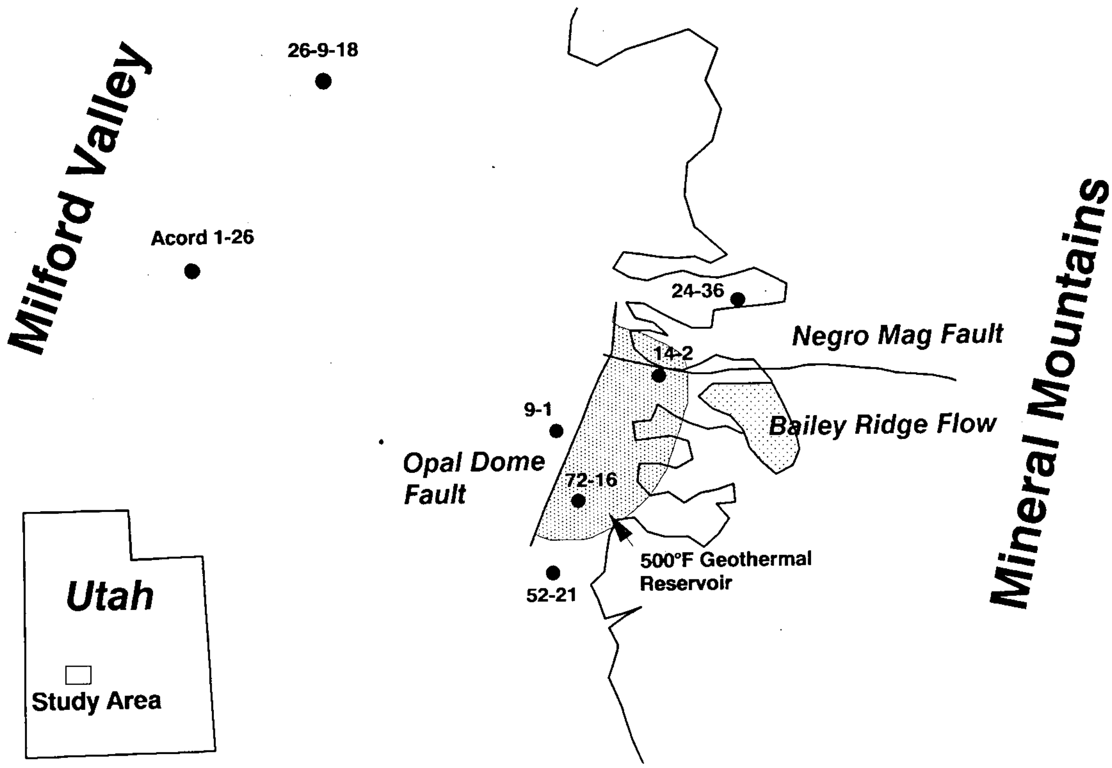
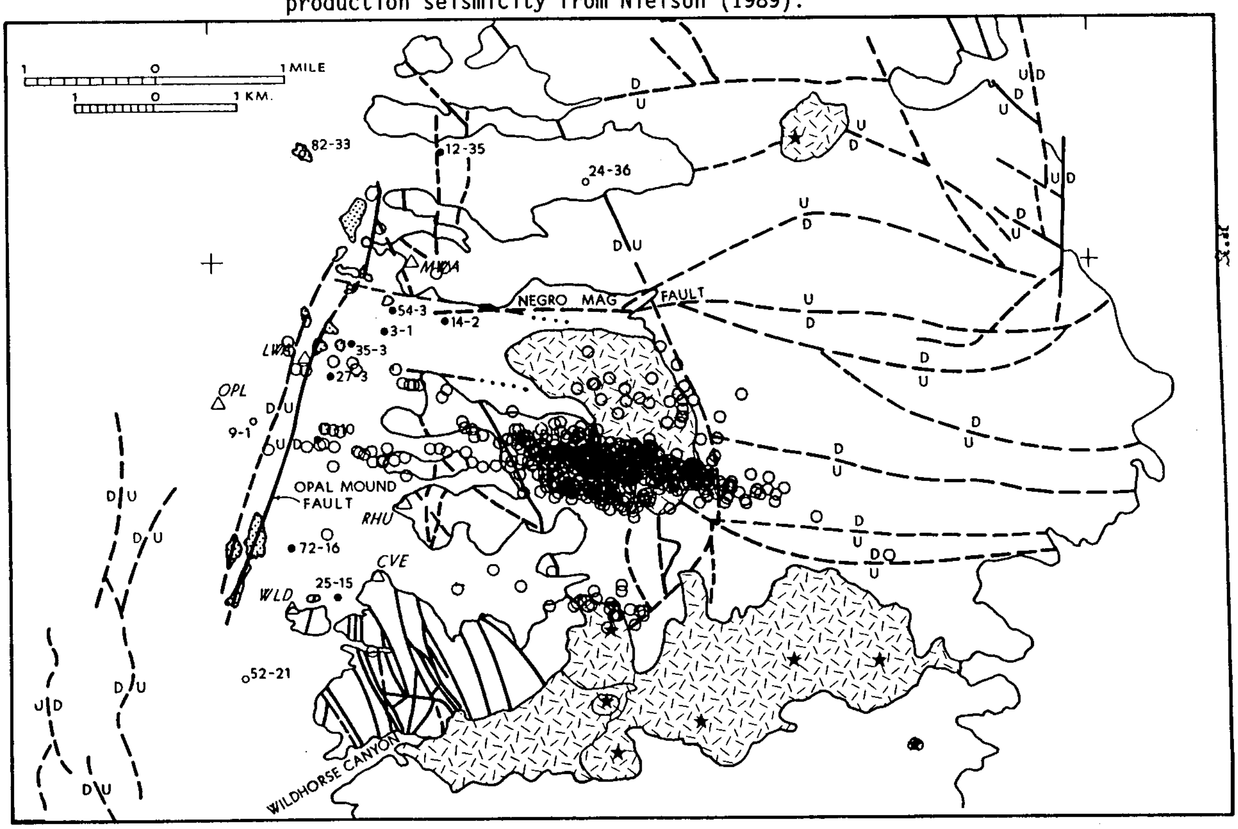


Figure 2 Geologic map of Roosevelt Hot Springs showing mapped faults and pre-production seismicity from Nielson (1989).



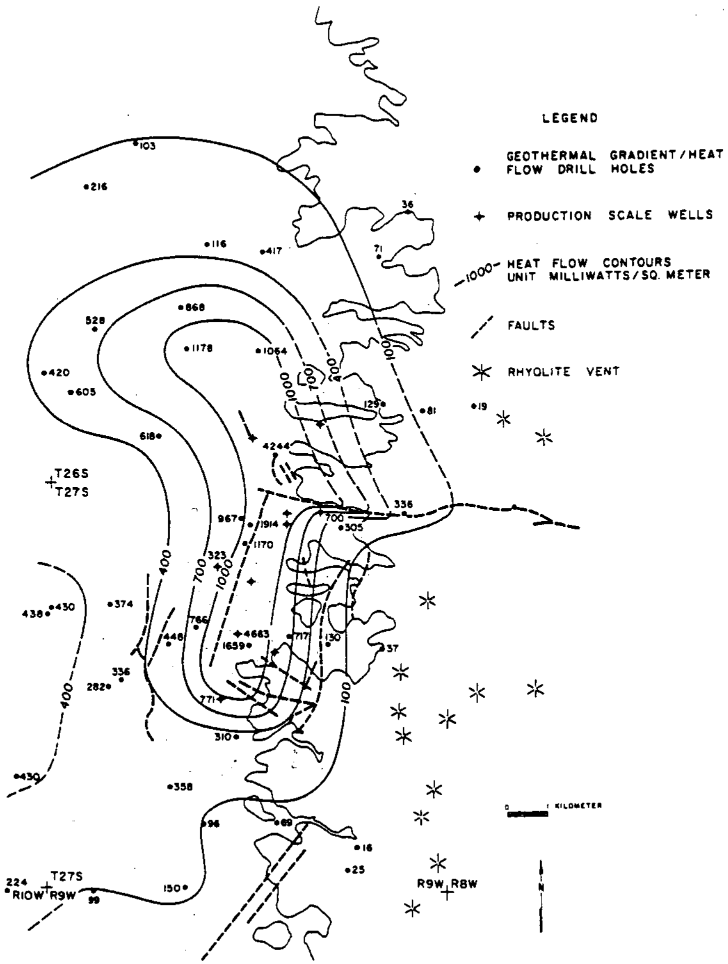


Figure 3

Surface conductive heat flow for Roosevelt Hot Springs from Wilson and Chapman (1980).

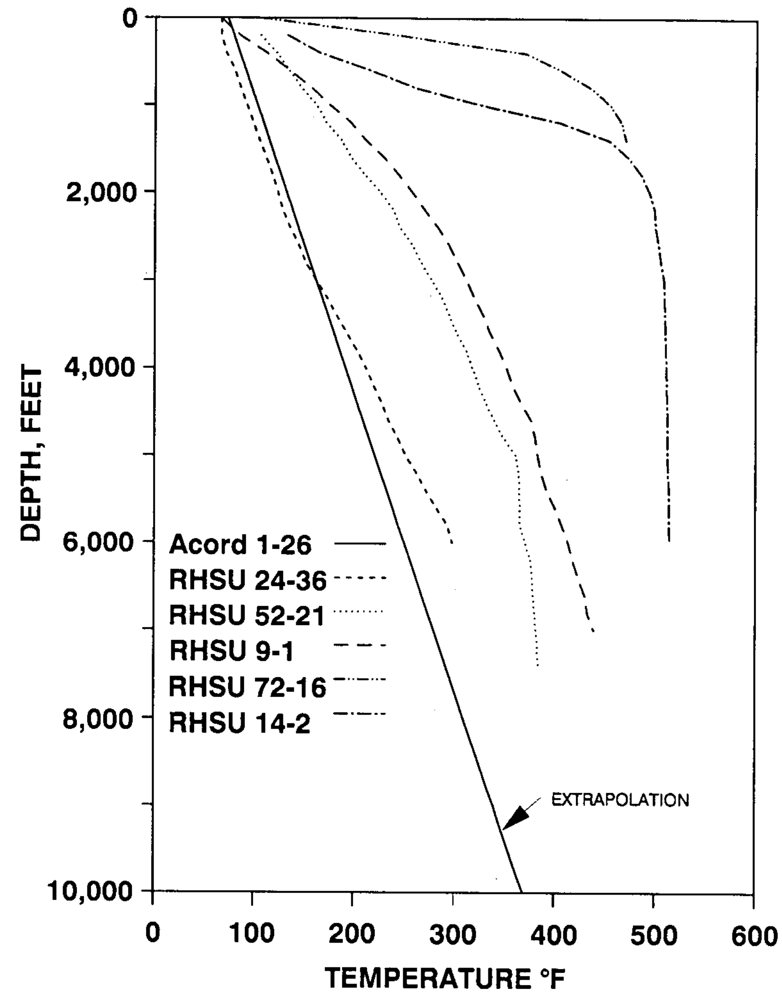


Figure 4 Temperature profiles of select wells.

Figure 5   Conceptual geologic model of the Roosevelt Hot Spring hydrothermal system.

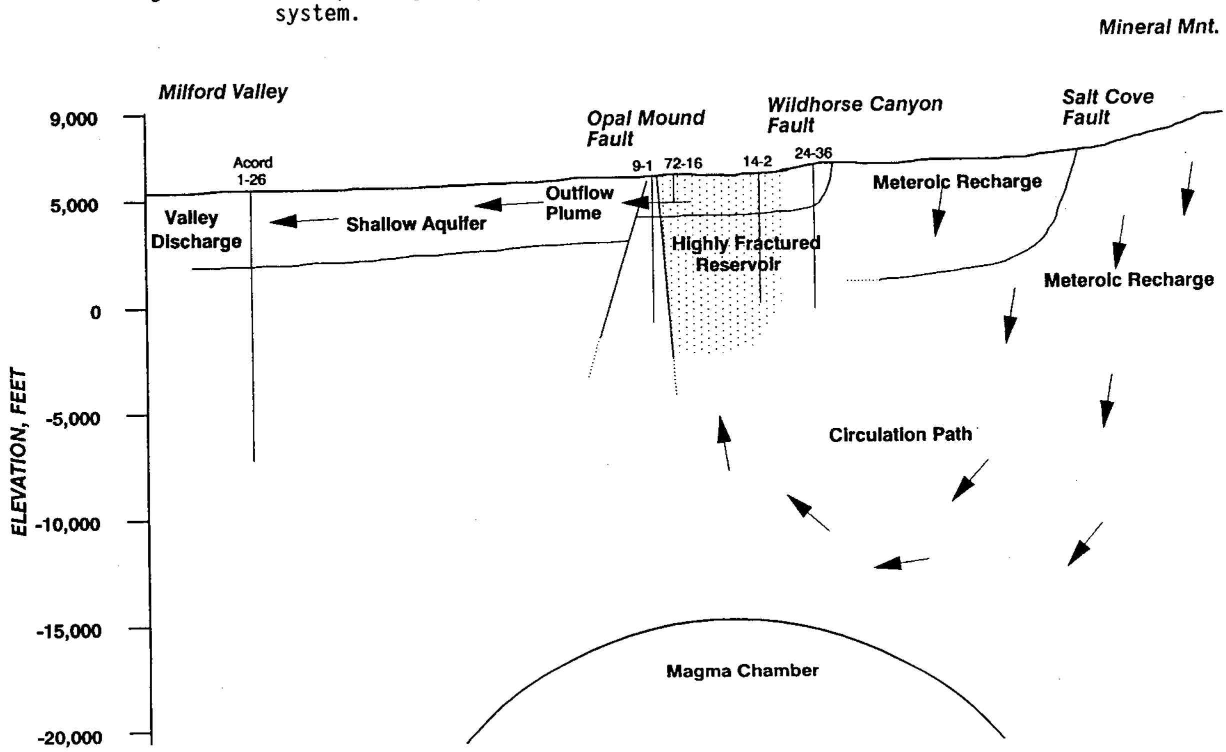


Figure 6   Horizontal permeability (mD) used in the simulation, note 0.5 mD in mountains massif.

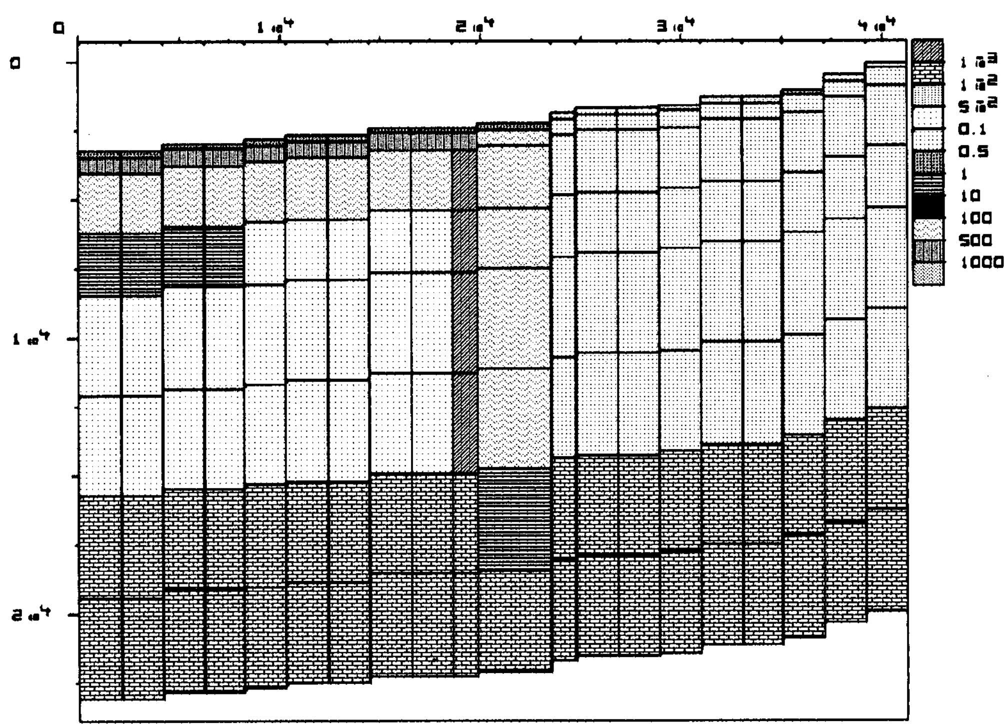


Figure 7 Vertical permeability (mD) used in the simulation study, note 0.5 mD in mountain massif.

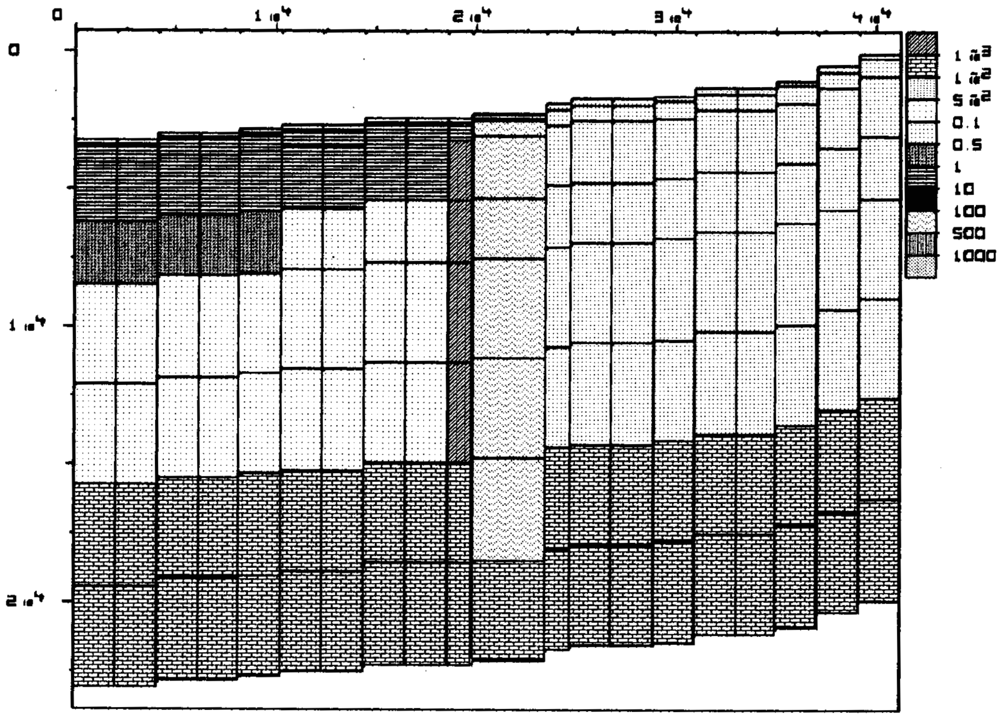


Figure 8 Model temperatures (°F) after 30,000 years for 0.5 mD in mountain massif.

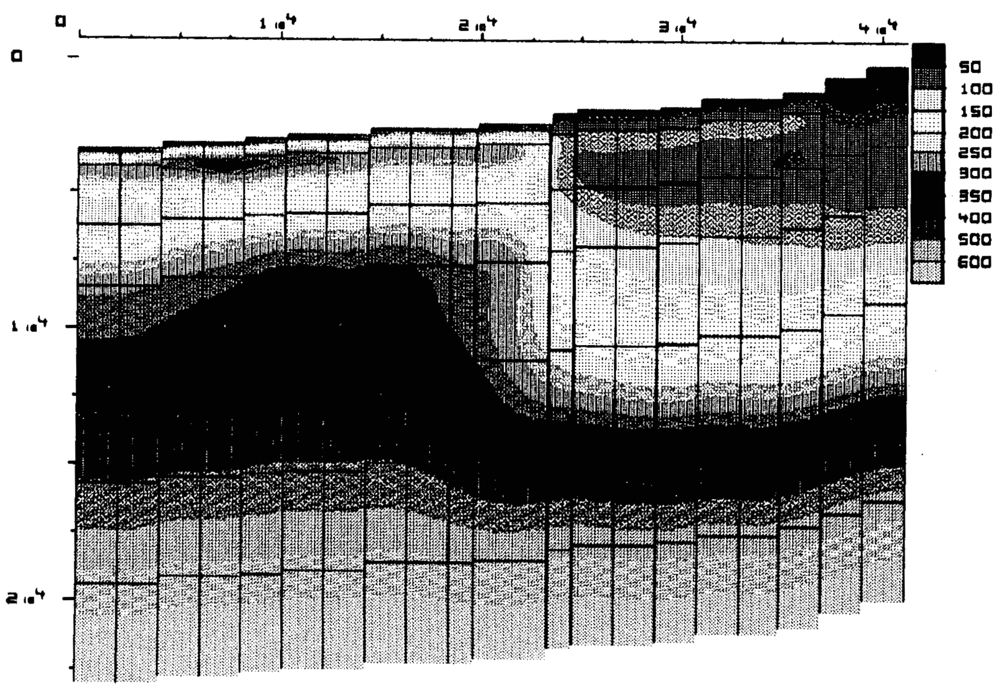


Figure 9 Model temperatures (°F) after 50,000 years for 0.05 mD in mountain massif.