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Transient Thermal Conditions in Basin and Range Extensional Geothermal Systems

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ABSTRACT

The relationship of high temperature extensional geothermal systems to young normal faults has been a subject of empirical comparison. We examined the cause for the relationship by investigating the time varying behavior of a suite of numerical models of Basin and Range geothermal systems utilizing a range of bulk rock permeabilities and a typical geometry. The modeling focused on the evolution of the system temperature with time. The geometry of the model consists of two mountain ranges (~1 km relief from the valley floor) separated by a thick sequence (about 4 km) of clastic sediments derived from the adjacent ranges, and a relatively permeable, high angle fault that functions as a conduit for subsurface fluids. The reservoir conditions were characterized by utilizing several parameters, including temperature along the producing fault, maximum temperature away from the fault in the probable reservoir region, mass transport, and the predicted surface heat flow. A previous study of the steady state case of models with similar properties was not able to match the high temperature observed at Dixie Valley. However, transient models are able to do so over a limited range of permeability, time, and heat flow. Thus there are insights into the exploration approach for such systems from the modeling and an explanation of the empirical relationships observed.

Introduction

Active geothermal systems in which subsurface temperatures are in excess of 200°C, and even 250°C by 2-3 km depth, are associated with Quaternary normal faulting in the Basin and Range. These systems are non-magmatic in origin based on the Helium 3 values observed (Kennedy, 2002). Meteoric water enters via the range or valley, warms during deep circulation, and ascends along high permeability pathways, usually range-bounding faults. In a number of Basin and Range geothermal systems temperatures are over 200°C. Given the typical background gradients of about 25 to 40 °C/km these temperatures imply circulation to depths of 5-10 km. One such geothermal system with unusually hot temperatures: in excess of 280°C by 3 km depth, is located in Dixie Valley, Nevada (see Blackwell et al., 2000). Dixie Valley is the site of numerous hot springs and fumeroles. Along a distance of at least 10 km along the contact of the Stillwater Range and Dixie Valley the temperatures are over 200°C by 3 km depth. Thus the conditions indicate high subsurface temperatures and a 2-dimensional flow system perpendicular to the range valley contact.

Our objective in this study is to determine under what conditions a reservoir temperature near 280°C is generated and if possible, sustained. We characterize the reservoir by utilizing several parameters, including temperature along the producing fault, maximum temperature away from the fault in the probable reservoir region, mass transport, and the predicted surface heat flow. The models we present are similar to the natural geometry of extensional geothermal systems: two mountain ranges (~1 km relief from the valley floor) separated by a valley filled with a thick sequence (about 4 km) of clastic sediments derived from the adjacent ranges. The boundary between one range and the valley is a relatively permeable, high angle fault (65°) that functions as a conduit for rising hot fluids. This research extends recent models of the flow regime in these types of systems by Wisian (2000). Utilizing the TOUGH2 (Pruess et al., 1999) porous media fluid flow modeling code, the parameter space of the numerical models necessary to establish sufficient upflow along a permeable fault to match 1) the observed discharge at the fault/valley contact, and 2) the steady-state temperature distribution was investigated. In this paper we describe the transient numerical results as a potential explanation for reconciling higher observed reservoir temperatures (from precision temperature logs) than predicted by steady-state models.
A modified version of the equation-of-state for pure water (EOSI) for TOUGH2 that allows for supercritical conditions in pure water (EOSIsc, Brikowski, 2001) was used in the modeling. Both the EOS1 and EOSIsc modules yield almost identical results; however, because even these non-magmatic geothermal systems are near the critical point of pure water by 6-9 km depth, the latter module provides more realistic characterization of reservoir pressure-temperature conditions. Earlier steady-state models are described in more detail by Wisian and Blackwell (2003).

Thermal and Flow Regime

The geothermal systems are assumed to thermally evolve from a conductive regime as the range-bounding fault permeability abruptly increases during/after an earthquake, opening a short circuit that allows rapid ascent of heated fluid. Figure 2 shows the steady-state temperature and velocity distribution obtained after a total simulation time of 32 Myr ("steady-state") utilizing a bulk rock permeability of 1 x 10^-20 m^2 (i.e., domain 1, see Figure 1). At such a small permeability, almost all the fluid velocities are low (with the exception of the range tops). The only observed thermal effect is a modest conductive increase in temperature between the ranges (domains 6-9 in Figure 1) due to refraction of heat by the low thermal conductivity of valley fill. From the conductive steady-state model, temperature and pressure conditions are utilized as the initial conditions for all transient flow models.

A steady state thermal model for two other bulk rock permeability values, 1 x 10^-16 m^2, and 1 x 10^-16 m^2 are shown in Figures 3 and 4. Under these conditions, models of bulk rock permeability > 1 x 10^-16 m^2 are completely dominated by fluid flow in steady-state and temperatures predicted by the models are unrealistically low. At permeabilities less than 1 x 10^-16 m^2 models are nearly

Steady-State Results

Model Geometry

The geometry utilized in the numerical modeling is illustrated in Figure 1. The modeling parameters specific to each numbered domain are listed in Table 1. The valley fill incorporates anisotropic permeability to minimize the discharge in the valley in accord with observations. The cell size away from the fault zone is 230 m x 90 m, whereas immediately adjacent to the fault, the cell size is halved (130 m x 45 m) to provide sufficient resolution of both the thermal and flow regime near the fault. The total number of elements is 7837. The models were developed utilizing PetraSim by Thunderhead Engineering Consultants. Unless stated otherwise, all the models developed in this study utilize surface temperature and pressure boundary conditions of 20°C, and 1.01 x 10^5 Pa, respectively. A basal heat flow of 90 mW/m^2 directed into the model (chosen so that the model accurately represents the high heat flow Basin and Range Province) is specified for the basal row of cells. The heat flow used here is a reasonable average for the western part of the Basin and Range although several different heat flow values were used by Wisian and Blackwell (2003).

Table 1. TOUGH2 modeling parameters. A constant density and heat capacity of 2650 kg/m^3, and 1000 J/kg°C, respectively were utilized in the modeling. The bulk rock permeability is varied in each model. Unless noted otherwise, a basal heat flow of 90 mW/m^2, surface temperature of 20°C, and surface pressure of 1.01 x 10^5 Pa were specified as boundary conditions.

<table>
<thead>
<tr>
<th>Domain</th>
<th>Material</th>
<th>Porosity</th>
<th>Wet Thermal Conductivity (Wm°K^-1)</th>
<th>Permeability</th>
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<tr>
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<td>Bulk Rock</td>
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<td>2.50</td>
<td>Variable</td>
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<td>Fault</td>
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<td>2.50</td>
<td>1.0E-14, 1.0E-14</td>
</tr>
<tr>
<td>3</td>
<td>Fault, Right-Side</td>
<td>1.0E-01</td>
<td>1.25</td>
<td>1.0E-15, 1.0E-15</td>
</tr>
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<td>4</td>
<td>Fault, Left-side</td>
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<td>1.25</td>
<td>1.0E-16</td>
</tr>
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<td>Bottom</td>
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<tr>
<td>6</td>
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<td>1.25</td>
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<tr>
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<td>Valley-Fill 2</td>
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<td>1.0E-16</td>
</tr>
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</table>

Figure 2. Predicted fault temperature for several bulk rock permeabilities and simulation times. Conductive steady-state is shown for bulk rock permeabilities 1x10^-17 and 1x10^-20 m^2.
Figure 3. Steady-state thermal and flow regime obtained utilizing a bulk rock permeability of $1 \times 10^{-16}$ m$^2$, and a basal heat flow of 90 mWm$^{-2}$.

Figure 4. Steady-state thermal and flow regime obtained utilizing a bulk rock permeability of $5 \times 10^{-16}$ m$^2$, and a basal heat flow of 90 mWm$^{-2}$.

Figure 5. Predicted surface heat flow for several bulk rock permeabilities (top) and simulation times (bottom). Note the vertical-heat flow-scale change between the two panels.

The predicted surface heat flow for the steady-state conductive ($1 \times 10^{-17}$ m$^2$) and higher bulk rock permeability models ($1 \times 10^{-16}$ and $5 \times 10^{-16}$ m$^2$) are illustrated in the upper panel of Figure 5. In the conductive case, the modeled surface heat flow is equal to the basal heat flow, although the overall shape of the curve is complicated by heat refraction effects from the dipping fault/valley fill contact. The models that utilize the larger bulk rock permeability both show a >450 mWm$^{-2}$ heat flow anomaly centered on the fault/valley fill contact, although the less permeable model predicts a heat flow anomaly about 100 mWm$^{-2}$ greater than the more permeable model (upper graph Figure 4). Additionally, the predicted heat flow in the ranges is somewhat higher for the less permeable bulk rock model than that utilizing more permeability due to the lower fluid recharge rate in the ranges.
Fault Temperature

The fault temperatures for the steady-state thermal regimes modeled appear in Figure 2. In each case, the temperature is extracted along the fault (the solid line dipping at 65° in each thermal cross-section) and plotted as a function of depth. The maximum steady-state fault temperature is about 220°C for higher bulk rock permeability model and about 160°C for the lower bulk rock permeability model. These temperatures are significantly lower than the ~280°C temperatures measured in Dixie Valley. The shape of the temperature-depth curves offers one possible explanation. In geothermal systems the fault temperature at approximate depths of 0 to 3 km is significantly warmer than a purely conductive model at an identical depth. However, even with modest upflow, the fault should remain at least as warm as the conductive model over the entire fault depth. The severe departure of the 5 x 10^{-16} m^2 bulk rock model from this ideal suggests that heat mining by the convection is important in this particular model, and in regional flow models with high permeability. Hence low regional heat flow is predicted to be associated with large scale flow systems of this sort. This is not the case in Dixie Valley where the heat flow is high everywhere measured.

Transient Results

Geothermal systems in extensional settings go through cycles of fracture (associated with earthquakes) and chemical deposition/sealing of the permeable fault. Hence the fault temperature might reflect the recurrence interval or the age of the last event on the fault. Typical Basin and Range faults appear to sustain large earthquakes every 1,000-20,000 years. The southern Dixie Valley fault last ruptured in 1954, but the area where the geothermal system is located has not ruptured for about 3,000 years (J. Caskey, personal communication, 2002). The problem of reconciling the very high observed reservoir temperatures in Dixie Valley with the lower temperatures modeled at steady-state might be related to the steady state assumption. Hence in the effects of transient conditions in the geothermal system were investigated based on the models described.

Transient Fault Temperature

The temperature-time history for a cell located at the downdip edge of the fault, approximately 3.85 km below and 2.15 km to the right of the fault/valley contact (see Figure 1) utilizing several bulk rock permeabilities are shown in Figure 6. The cell records the maximum fault modeled temperature. For models utilizing bulk rock permeabilities smaller than 1 x 10^{-17} m^3, the thermal regime is essentially conductive. For the larger bulk rock permeabilities, however, the temperature at the base of the fault varies strongly as a function of time. The maximum temperature for the cases investigated is 255-275°C and does not occur at steady-state, rather at about 60 kyr for the 5 x 10^{-16} m^2 bulk rock permeability model, and at about 359 kyr for the 1 x 10^{-16} m^2 bulk rock permeability model.

Thus, it appears that much like actual geothermal systems, the models presented here "mine" heat over time causing the system to cool significantly. Therefore the large difference between the steady-state temperatures predicted by the 1 x 10^{-16} and 5 x 10^{-16} m^2 models can be explained by the mining of the heat. So much so, that the overall thermal regime is cooler as shown by the background heat flows in Figure 5 away from the fault.

Because both permeability models predict temperatures that are still somewhat lower than the ~280°C temperatures measured in Dixie Valley, several models were created to test the effect of deepening the fault and valley fill by 2 km. Figure 5 also shows the results from these models. The maximum temperature for the higher bulk rock permeability case occurs at about 200 kyr earlier, whereas the maximum temperature for the lower bulk rock permeability case occurs about 20 kyr later. The increase in fault temperature is only about 10-15°C, which albeit a small increase, is sufficient to match the measured temperature.

Transient Surface Heat Flow

The predicted surface heat flow for the models at the simulation time corresponding to the maximum fault temperature for three different bulk rock permeability models (1 x 10^{-20}, 1 x 10^{-16}, and 5 x 10^{-16} m^2) are shown in the lower panel of Figure 4. The predicted heat flow at the fault/valley contact is almost 200 mW/m^2 (1 x 10^{-16} m^2 case) and 500 mW/m^2 (5 x 10^{-16} m^2 case) higher at the time of maximum temperature than for the respective steady-state cases. Furthermore, the predicted heat flow in the range is somewhat higher due to the lower recharge rates at the earlier simulation times and the lack of time for heat mining to occur. The double peak in the predicted heat flow near the fault/valley contact for the (5 x 10^{-16} m^2 bulk rock permeability case is an artifact of the manner in which the heat flow is calculated and is not meaningful.

Wisian et al. (2001) calculated a heat loss of 1 x 10^7 W for the Dixie Valley geothermal system. Normalizing their results by the
actual geothermal production area (12 km²; 6 km along strike x 2 km along dip), the Dixie Valley heat loss is about 833 mWm⁻². Clearly, the best match to the observed heat loss is obtained at the time of maximum temperature (60 kyr, and 359 kyr for a bulk rock permeability of 5 x 10⁻¹⁶ m², and 1 x 10⁻¹⁶ m², respectively).

Conclusions

The numerical models discussed are highly simplified and stylized and the details of the results may not apply to any real geothermal system. However, the results do point to some valuable conclusions with respect to the exploration for Basin and Range geothermal systems.

The most important observations obtained from our simulations are that transient effects in these types of systems can dramatically modify the maximum predicted reservoir temperature. For example, the maximum fault temperature of about 275°C obtained utilizing a bulk rock permeability of 5 x 10⁻¹⁶ m² does not occur at steady-state, but rather at 60 kyr, and is about 110°C hotter than at steady-state. The 1 x 10⁻¹⁶ m² bulk rock permeability model behaves similarly. It is interesting to note that a 2 km deeper fault yields similar behavior and temperatures. Furthermore to satisfy the observed fault temperature, flow rates, and heat flow, we conclude that high permeability persists to at least 6 km depths and that much of the fluid flow penetrates through the brittle crust.

The flow regimes modeled suggest that the heat present in the system is “mined” over time causing the system to cool significantly, but nonetheless, the system may persist if not sealed for millions of years at commercially exploitable temperatures. The problem of reconciling higher observed reservoir temperatures with the lower temperatures modeled at steady-state, may be a problem of determining where in the temporal-evolution of the geothermal system production is situated. If higher temperatures are required, they are easily obtained from the models simply by assuming that the present thermal and flow regime is early in the system cycle. In fact, the present day heat loss from the Dixie Valley geothermal system suggest that the system is not at steady-state, but rather early in its temporal evolution (perhaps only a few hundred thousand years into the cycle). The spectrum of geothermal system temperatures in the Basin and Range can be matched by a variety of time varying points in the system temperature evolution. As the systems evolve with time more and more heat is mined from the surrounding country rock if sealing does not stop the deep circulation. Therefore the observation that low heat flow is not seen on a regional basis in the areas of greatest heat transfer by geothermal systems, such as in the Carson Sink area of Nevada, implies that the systems are not in steady state equilibrium.

However, some process must still prevent the system from mining enough of the heat so that it “shuts-off” completely. The relatively high reservoir temperatures commonly observed in geothermal systems (> 280°C) must be a function of oscillating high/low fault permeability maintained by seismicity along the range-bounding fault. Thus the upflow temperature and flow simply wanes with decreasing fault permeability, and/or perhaps is rerouted elsewhere. The models presented here are 2-d whereas many systems are 3-d (unlike Dixie Valley which is close to 2-d if not truly 2-d). Higher system temperatures and less regional heat mining are probable for the 3-d systems.

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References


