PROCEEDINGS, Twenty-fourth Workshop on Geothermal Reservoir Engineering
Stanford University, Stanford, California, January 25-27, 1999
SGP-TR-162

LOWER LIMITS OF HYDROTHERMAL CIRCULATION IN THE TIWI GEOTHERMAL FIELD, LUZON

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ABSTRACT

In high-temperature geothermal systems, the lower limit of hydrothermal circulation is controlled by a decrease in permeability. This limit is a fundamental aspect of the circulation volume and the heat budget, and, as a consequence, it will influence the ultimate heat recovery from geothermal systems and the procedures that are applicable for heat mining below the permeability interface. In this paper, we consider the apparent termination of permeability with depth as documented in corehole Matalibong-25 in the Tiwi geothermal field. The bottom of convective circulation at Tiwi cannot be related to either the transition from brittle to ductile behavior which occurs at a temperature of about 400°C, or to changes in rock lithology. We suggest that the reduction in permeability is due to a stress reorientation with depth. The termination of permeability with depth, whatever the cause, requires that heat be transferred to the convective portion of the system by conduction. Temperatures within the overlying convective zone are buffered by the boiling point to depth curve. We use time-temperature constraints imposed by $^{40}Ar/^{39}Ar$ age spectra on adularia, hydrothermal alteration mineralogy, and fluid-inclusion geothermometry to constrain the intrusive history of the system and the transfer of heat across an impermeable barrier. In the Tiwi system, three intrusive episodes are required to match the observed time-temperature relationships.

INTRODUCTION

All high-temperature geothermal systems are supported by heat transfer from an underlying magma chamber (Smith and Shaw, 1975). The transfer of heat from the magma chamber to the convective volume is governed by conduction. This interface between conductive and convective regimes can be determined by several factors. Perhaps the most discussed is the transition from brittle to ductile behavior of the rocks (Fournier, 1991; Nielson, 1996). Termination of permeability at the brittle-ductile transition is important in fields such as Larderello (Manzella *et al*., 1995) and Kakkonda (Yagi *et al*., 1995) where high temperatures (370°- 400°C) are associated with pressures exceeding hydrostatic. At lower temperatures, the presence of lithologies that are not able to support open fracturing under hydrostatic pressures may contribute to decreased permeability. The thick section of Mesozoic shale in the Valles caldera which separates two major zones of hydrothermal circulation is an example (Nielson and Hulen, 1984).

The transition from conductive to convective thermal regimes can occur over very short distances and can be marked by steep temperature gradients. Within the convective zone, temperature will closely follow the boiling point to depth curve (e.g. Henley, 1985).

Although research has focused on the upper parts of the hydrothermal circulation volume where permeability is highest, the lower parts are of interest because they will determine how the systems will naturally evolve. As technology improves, more developers will be interested in the recovery of heat from the lower parts of active hydrothermal systems (Nielson, 1996).

THE TIWI GEOTHERMAL SYSTEM

The Tiwi geothermal field is a hot-water dominated system located in southern Luzon, Philippines. Geothermal activity is closely related to Mt. Malinao, a volcano whose initial eruptions have been dated at 500 ka. The thermal structure of the field and its main features have been described by Gambill and Beraquit (1993). Information on the thermal history of the system and the fracture distribution in the reservoir have come mainly from studies of core from Matalibong-25 (Nielson *et al*., 1996). This well was continuously cored from depths of 789 to 2439 m. Matalibong-25 is located in the western portion of the field, and has a maximum measured temperature of 275° C.

Constraints on the Thermal History

The hydrothermal history of the rocks in Matalibong-25 was established by combining the succession of hydrothermal alteration assemblages with $^{40}Ar/^{39}Ar$ age spectra on secondary adularia and fluid-inclusion homogenization temperatures (Moore *et al.*, 1997, 1998). Six stages of alteration and vein mineralization were documented in the cored portion of the well. The earliest (stage 1) is characterized by the deposition of chalcedony and clays and represents the initial heating of the system. No fluid-inclusion data were obtained on this stage. However, chalcedony is commonly found at temperatures between 140° and 180°C; at higher temperatures quartz is the typical silica polymorph (Fournier, 1985). For the purpose of modeling the time-temperature history of the system, we have chosen a temperature of 150°C to approximate the initial wall rock conditions during this stage. Stages 3-5 represent the main period of hydrothermal alteration in Matalibong-25. During stage 3, quartz ± epidote ± adularia ± sulfides were deposited in veins by boiling fluids. These assemblages frequently cement hydrothermal breccias confirming the vigorous nature of the boiling. Fluid inclusions trapped in the associated quartz demonstrate that deposition of adularia occurred at temperatures close to 330°C which is consistent with temperatures defined by the boiling point to depth curve.

$^{40}Ar/^{39}Ar$ age spectra were obtained on adularia from three depths between 1808 and 1851 m. These data constrain the formation of adularia to the period between 314 and 279 ka. Subsequent alteration included the formation of sericite, and wairakite + epidote (stage 4), followed by calcite and then actinolite during stage 5. The presence of actinolite indicates that temperatures in the lower part of the well must have exceeded 300°C at the end of the main stage of mineralization. $^{40}Ar/^{39}Ar$ age spectra indicate that these temperatures could have persisted until about 200 ka when the system underwent significant cooling. Secondary fluid inclusions suggest that minimum temperatures stabilized at about 235° C; although, inclusions record temperatures as low as 191° C. The present temperature of 275° C implies reheating by recent subvolcanic intrusions. Modeling of the $^{40}Ar/^{39}Ar$ age spectra suggests that the present temperature profile is not likely to have persisted for more than the last 50,000 years. Figure 1 presents a thermal model that is compatible with the measured adularia age spectra and the petrologic data.

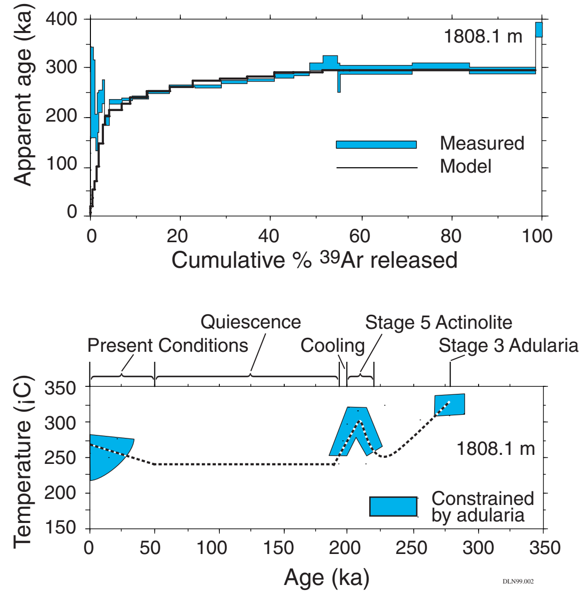

Fig. 1. Measured and modeled age spectra (a) and thermal history (b) for adularia from a depth of 1808.1 m. The regions shaded in gray represent the portion of the model constrained by the adularia age spectra. The timing of key events during the evolution of the Tiwi system are shown in b.

Permeability Structure

An important feature of the Tiwi system is an apparent termination of permeability with depth. Gambill and Beraquit (1993) reported that the basement rocks beneath the field reach temperatures of more than 315°C, but are relatively impermeable and unproductive. This reservoir bottom in the Naglagbong area is reported by them to be at an elevation of about 1500 m below sea level.

Nielson *et al.* (1996) documented the lithologic and fracture distributions in Matalibong-25. That paper concluded that the fractures controlling permeability are predominantly steeply dipping. It was also found that both fracturing and permeability decreased markedly in the lower part of the well. It appears that both the permeability and the frequency of veins decrease below about 2194 m depth (1643 m below sea level). There are a number of possible explanations for this. The first possibility would be a brittle-ductile transition. However, measured temperatures are too cool and analysis of fluid inclusions suggests that the rocks were never hot enough for this explanation to be valid. The second possibility would be a lithologic effect. However, no major lithologic changes were observed in the lower part of the core, and permeability decreases related to rock type are unlikely. Within Matalibong-25, below depths of 1600 m, the rocks consist mainly of volcanic sandstones. At greater depths, limestones, shales, and metamorphic basement are expected (Gambill and Beraquit, 1993). The third possibility would be a change in stress from extension above to compression below.

THERMAL MODELING

Modeling of heat transfer can be useful for placing constraints on the size and age of intrusive events that are responsible for the observed thermal history of a geothermal system. Although we recognize that convection is important at Tiwi, particularly in its upper parts, the data suggest that conductive heat transfer dominates below about 1500 to 1650 m below sea level. Thus, in this paper, only the conductive cooling of an intrusive body and the surrounding wall rocks are considered. Because convection will cool rocks more rapidly, the conductive models presented here provide estimates of the maximum longevity of the system. The use of conductive cooling models has been discussed by Lovering (1935) and Carslaw and Jaeger (1959), and we are using heat conduction equations from those papers in this study. However, we have the advantage of having $^{40}Ar/^{39}Ar$ age spectra that provide constraints on temperature, age, and duration of the heating events, hydrothermal alteration mineralogy that shows the character of the water-rock interaction, fluid-inclusion analyses that define temperatures and salinities associated with the alteration episodes, and fracture distributions that provide information on permeability.

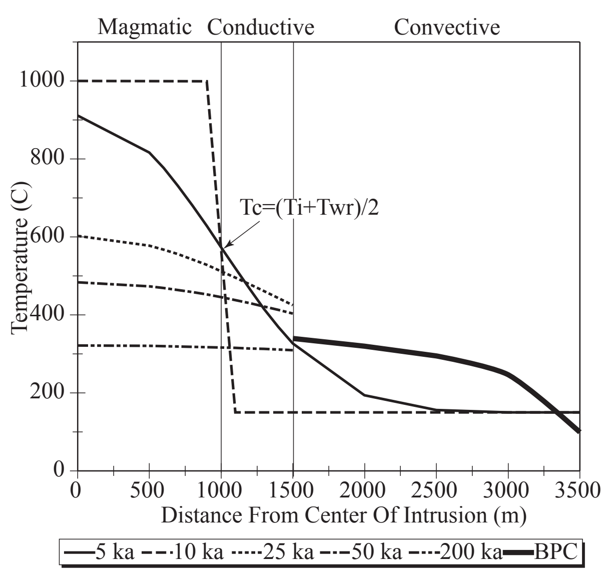

Fig. 2. Conductive cooling model of a one-dimentional intrusion with an initial temperature of 1000°C emplaced in 150°C wallrock. Temperature in the convective zone are limited by the boing point curves to depth (BPC).

The thermal zones of a high-temperature geothermal system and the changes they experience with time are illustrated in Figure 2. The lowest zone is the magmatic heat source that is being modeled as a dike of infinite areal extent with a total thickness of 2000 m and an initial temperature of 1000°C. This is instantaneously emplaced into wall rock with a temperature of 150°C. Conduction models (Lovering, 1935) show that the temperature at the contact between the intrusive and the wall rock is $T_c = (T_i + T_{wr})/2$, where T_i is the initial temperature of the intrusive and T_{wr} is the initial wall-rock temperature. Thus, the temperature of the wall rocks will not rise above T_c without release of fluids or another intrusive event. The conduction of heat away from the contact is a function of the temperature gradient and the thermal diffusivity (h^2) for which we used a value of 0.012 $cm^2 sec^{-1}$.

Heat is conducted away from the intrusive across a zone whose thickness will change with time. This is termed the conductive zone, and is drawn as being 500 m thick on Figure 2. This zone has modeled temperatures of greater than 400°C between about 5 ka and 50 ka in this example. In natural systems, magmatic fluids with salinities exceeding 26 weight percent NaCl equivelent are likely (Moore and Gunderson, 1995, Hedenquist *et al.*, 1998) and contact metamorphism may lead to the development of hornfelsic texture. Pressures greater than hydrostatic, which mark the change from brittle to ductile behavior can be expected (Fournier, 1991).

In many cases, the brittle-ductile boundary will be coincident with the transition from the conductive to convective regime which is shown as a sharp interface in Figure 2, but it is probably gradational in nature. Above the conductive-convective interface, the temperatures within the geothermal system are buffered by the boiling point to depth curve. If the temperature at the interface is initially greater than the boiling point curve, boiling will take place until the temperature is cooled to that defined by the curve for a particular depth. Under this situation, we expect to observe hydrothermal breccias, alteration assemblages diagnostic of boiling, and vapor-rich fluid inclusions.

Model 1

The geothermal reservoir at Tiwi underlies an area of approximately 1800 hectares and has dimensions of about 5 km x 3.6 km (Gambill and Beraquit, 1993). Although the size and temperature of the intrusive body that triggered geothermal activity are unknown, the presence of advanced argillic alteration, the temperatures of the early hydrothermal fluids and their gas compositions provide unequivocal evidence of its presence (Moore *et al.*, 1998). For these calculations, we have arbitrarily assumed that the areal extent of the intrusive heat source is equivalent to the productive area of the field and that it lies directly beneath the reservoir. It was also assumed that the permeability decrease with depth resulted from a change in the stress orientation due to doming above an intrusive as proposed by Johnson (1970). A similar model was applied to the decrease in permeability in the Valles caldera geothermal system by Nielson and Hulen (1984). Johnson (1970) demonstrated that if a homogeneous plate is domed, a neutral plane located half way through the plate separates an extensional zone above from a zone under compression below. Using this simple model, one would predict that if permeability were lost at 2200 m depth (as it is in Matalibong-25), the causative intrusive would be located at a depth of 4400 m.

We chose to apply a three-dimensional heat conduction model with an intrusive geometry that is shown in Figure 3. Sensitivity analyses were run using both 2-D and 1-D models, but both of these options showed significantly longer cooling times and produced much higher temperatures than did the 3-D case. Figure 2 shows the well Matalibong-25 in its proper location with respect to the production zone. The intrusive in the model was assigned a thickness of 10 km, a length of 5 km, and a width of 3.6 km. Initial simulations were run at a wall rock temperature of 150°C (based on mineralogical data) and intrusive temperatures of up to 1000°C. Given our initial assumptions, these simulations were not able to produce temperatures at the interface between the ductile and convective zone high enough to fit the assumptions of the domed model. Through trial and error, it appeared that the intrusive had to have been hotter and the distance from the intrusive less than was originally assumed. The distance was modified to 500 m (as shown in Fig. 3) and the intrusive temperature increased to 1200°C. This high a temperature would imply an andesite melt (Myers and Johnson, 1996), an assumption that is not unreasonable in this tectonic environment. We emphasize however, that the intrusive has not been observed and we are inferring its size and temperature. In fact, we should avoid using these models to draw conclusions concerning the details of the heat source.

Figure 4 shows the conductive cooling history of Model 1 that represents an intrusive with a temperature of 1200°C emplaced into 150°C wall rock at time 0. From the $^{40}Ar/^{39}Ar$ age spectra and fluid inclusion analyses, we know that the geothermal system requires sufficient heat to support temperatures on the boiling point curve at the sample depth of 1800 m. From Fig. 4, it appears that this is possible if the interface between the ductile and convective zones is

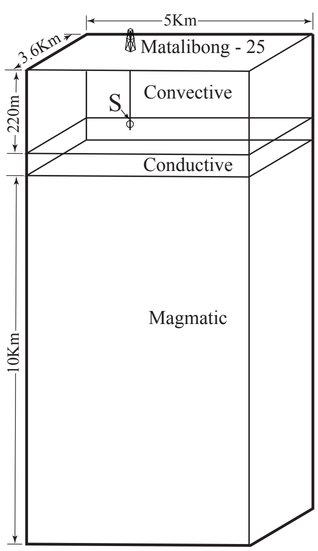

Fig. 3. Geometry of 3-D cooling model of Tiwi used for models 1, 2, and 3. S shows location of adularia samples

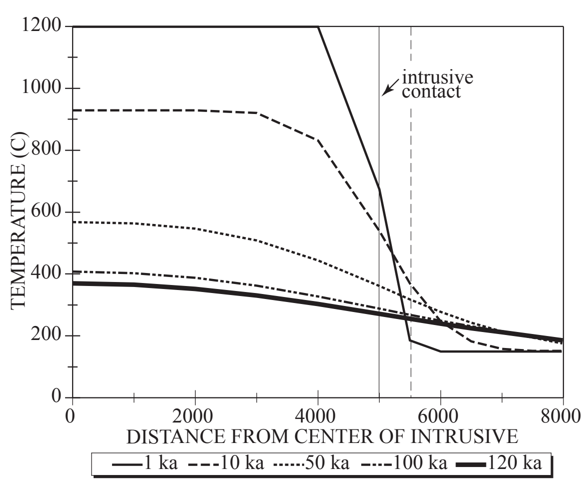

Fig. 4. Conductive cooling of model 1. Dashed line shows assumed location of conductive-convective boundry.

about 500 m from the contact. By rearranging the heat conduction equation, a time-temperature history at 500 m from the contact can be constructed (Fig. 5). This calculation shows a rapid increase in temperature within the first 5,000 years following an intrusive event; a temperature peak at about 10,000 years, and then gradual cooling. Figure 5 shows that temperatures of 340°C and above are present at the interface between 5 and 35 ka following the intrusive event. This is sufficient to maintain temperatures along the boiling point curve and produce vigorous boiling within the convective volume, consistent with the observations in Matalibong-25.

An interesting feature of Figure 5 is the relatively short duration of this heating event. Although we can match the temperatures and duration of heating during deposition of adularia between 314 and 279 ka, the model shows that temperatures would have cooled below 340°C after 35,000 years and below the present temperature of 275°C after 90,000 years. Thus, this intrusive event cannot provide sufficient heat to maintain a temperature in excess of 300°C during the latter half of the main period of mineralization (stages 4+5), as implied by the time-temperature relationships, or the present temperature profile.

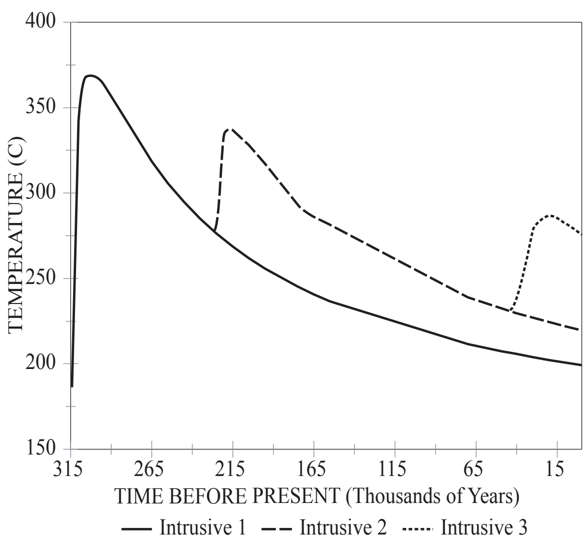

Fig. 5. Time-temperature profiles at 500 m from the intrusive contact for models 1, 2, and 3.

Model 2

Although temperatures exceeding 300° C between 314 and 200 ka could be achieved by increasing the size of the intrusion or its temperature (such a model is not specifically precluded by the $^{40}Ar/^{39}Ar$ age spectra), the presence of wairakite, epidote, and calcite prior to the formation of actinolite implies that temperatures first declined below 300°C and then increased. Figure 5 shows the effects of a second intrusion emplaced at 225 ka. At the time of this second intrusive event, the initial intrusive would have cooled to 400°C. Consequently, it was necessary to increase the emplacement temperature of the second intrusion by 200°C to 600°C in order obtain a satisfactory temperature at the conductive-convective boundary. This thermal pulse delivers a maximum temperature of 337°C to the conductive-convective interface. This would raise the temperature throughout the convective volume nearly to the boiling point curve but would be insufficient to produce more than local boiling in the reservoir. Continued conductive cooling to the present would result in a temperature of 219°C at the conductive-convective interface.

At about 200 ka, the $^{40}Ar/^{39}Ar$ data suggests that the convective portion of the system experienced cooling that we assume was produced by cool water flowing into the reservoir as a result of tectonic activity. We do not have a good understanding of this event, but find no evidence in Matalibong-25 that it affected the conductive-convective interface.

Model 3

A third intrusive event is required to raise temperature to that presently observed at 1800 m depth (275°C). Temperature distributions presented by Gambill and Beraquit (1993) show that the present upflow zone is located near the southwestern edge of the field. This suggests that the most recent intrusion is located in the southwest, although it may extend outside the current area of production. In formulating Model 3, we found it necessary to introduce a third intrusive event 41,000 years ago that increased the temperature of the intrusive body to 500°C to obtain the results shown in Figure 5. During this event, temperature was not sufficient to maintain boiling throughout the reservoir.

DISCUSSION

The transfer of heat from the underlying intrusive to the convective volume is governed by conductive heat transfer. To support this statement, we offer these observations. First, meteoric water will not circulate within an intrusive until it has crystallized and cooled to a temperature of less than approximately 400°C where open fractures can be maintained. Second, an intrusive will elevate the temperature of its wall rocks to above 400°C for some period of time during which these rocks will not be capable of supporting fracturing and thus fluid circulation (convective heat transfer). This was demonstrated by Fournier (1991), and similar conclusions have been reached by Marsh *et al*. (1997). As argued by Nielson (1996), cooling of the wall rocks and the intrusive with time will lead to the eventual collapse of the convective regime into the magmatic heat source.

The interface between the conductive and convective zone is very important in the thermal budget of the geothermal system. In a system that is lithologically and structurally homogeneous, this boundary will coincide with the brittle-ductile boundary at a temperature of about 400°C. In the Tiwi system, this boundary is found at T < 400°C, and we suggest that it is related to a change in the state of stress. The thermal energy delivered to the boundary by conduction will determine the behavior of the overlying convective volume with respect to the boiling point curve. If the temperature at the boundary is equivalent to the boiling point curve at that depth, the boiling point curve will be supported throughout the convective volume (barring influx of cold water into the system). If the temperature is less than the boiling point curve for the depth of the interface, at least part of the lower portion of the hydrothermal system will be less than the boiling point curve. If the temperature at the interface is greater than the boiling point curve, the hydrothermal system will boil, and vigorous boiling and the formation of hydrothermal breccias are expected.

The time-temperature profile determined from conductive cooling models suggests that intrusive events are associated with a rapid increase in temperature followed by gradual cooling. This suggests an episodic behavior of geothermal systems where intrusive events are manifested by active convection (perhaps boiling) and deposition of hydrothermal minerals followed by cooling and stagnation of the convective volume.

Another implication of the models presented here is that long-lived geothermal systems, such as Tiwi, require multiple episodes of igneous activity to sustain high temperature. Figure 5 shows that the initial intrusive episode at Tiwi would have cooled to less than the temperature required to support the boiling point curve (340°C) after 40,000 years and the temperature today at the conductive-convective interface would be about 200°C.

ACKNOWLEDGMENTS

This research was supported by DOE contract No. DE-AC07-95ID13274.

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