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THE DEEPER PARTS OF THE GEYSERS THERMAL SYSTEM - IMPLICATIONS FOR HEAT RECOVERY

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ABSTRACT

High-temperature hydrothermal convection systems are driven by the energy from shallow igneous bodies. The transfer of heat to the hydrothermal convection zone takes place across a zone of heat conduction from the intrusive to the overlying fluid circulation system. The transition from the convective to conductive thermal regimes is an important interface whose character and location changes with the evolution of the system. In some cases, this boundary appears to be related to the transition from brittle to ductile mechanical behavior. In other cases the boundary lies above the brittle-ductile boundary and results from situations where the rocks are under compressive stress or where the lithology is not able to sustain fracture permeability. Evidence also suggests an important physical boundary may be present at temperatures of about 340° C, below the temperature of the brittle-ductile transition. We believe it may be possible to mine heat from areas below the hydrothermal convective zone through deep injection.

1. INTRODUCTION

High-temperature geothermal systems are supported by heat transfer from an underlying magma chamber (Smith and Shaw, 1975). The transfer of heat to the convective volume is governed by conduction. The transition from a conductive to convective regime can be influenced 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 (Fournier, 1991). In other systems, for example the Tiwi geothermal system, there are downward decreases in permeability at temperatures less than those required for the ductile flow of rock (Gambill and Beraquit, 1993; Nielson *et al.*, 1999).

Above the conductive-convective interface, temperatures within the geothermal system are buffered by the boiling point to depth curve. If the temperature at the conductive-convective interface is greater than the boiling temperature of water at that depth, boiling will take place, and we expect to observe hydrothermal breccias, alteration assemblages diagnostic of boiling, and vapor-rich fluid inclusions. Therefore, boiling will buffer the temperature of the convective zone.

Geothermal 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 they cool, and this is a logical location for engineered heat recovery. The conductive zone is where much (80% in the case of Kakkonda; Nielson and Rose, 1999) of the thermal energy is located. However, at the present time, there are only limited data available from the area beneath the zone of active hydrothermal circulation. At The Geysers geothermal field, the hydrothermal zone has been superimposed on rocks that were ductile during the early history of the system. These rocks have been sampled by some of the deep wells and are available for study.

2. CONTACT METAMORPHISM AT THE GEYSERS

Although The Geysers is currently vapor-dominated, fluid inclusion and mineral relationships demonstrate the early geothermal system was liquid-dominated. Based on temperature and fluid composition determinations, this early system can be divided into two, physically distinct regimes (Moore and Gunderson, 1995; Moore *et al.*, 1999). The upper regime extended from the surface to the top of the biotite hornfels, which represents strongly metamorphosed metagraywacke (Fig.1). Within the upper regime, increasing thermal metamorphism as the pluton is approached is most clearly reflected in the progressive changes in the mineralogy of the veins. In the outermost portion of the thermal aureole, veins within the metagraywacke are characterized by quartz + calcite ± potassium feldspar. As the intrusion is approached, calcite disappears and epidote and then actinolite become stable vein minerals. Figure 2A shows the relationship between the homogenization temperatures of the inclusions with respect to distance from the intrusion. These temperatures must have developed shortly after intrusion of the Geysers pluton. The relationship between the homogenization temperatures and a boiling point to depth curve based on the estimated position of the topographic surface is shown in Fig. 3. The overall correspondence between the boiling point and fluid-inclusion temperatures argues that the pressures within this regime were hydrostatic.

The salinity of the inclusion fluids and the composition of the gaseous species indicate that the early hydrothermal fluids were crustal waters with moderate salinity of 3 to 6 weight percent NaCl equivalent. In the southern and central parts of the field, the early crustal fluids have mixed with meteoric waters. Fluid-inclusion gas compositions demonstrate that incursion of meteoric fluids occurred during the transition from an early liquid-dominated to the modern vapor-dominated regime. $^{40}Ar/^{39}Ar$ dating places the age of this event between 280 and 250 ka, some 800,000 years after formation of the system occurred (Hulen *et al.*, 1997).

The lower regime was developed entirely within the biotite hornfels and the underlying intrusive rocks. The biotite hornfels, which represents extreme thermal metamorphism of the metagraywacke, extends from the intrusive contact outward for a distance of up to 600 m. The earliest veins within the biotite hornfels and intrusive rocks are characterized by the presence of tourmaline and/or biotite + potassium feldspar + actinolite. Fluid inclusions trapped in quartz within these veins indicate that the mineralizing fluids were hypersaline and had temperatures exceeding 400°C. These fluids had salinity up to 44 weight percent NaCl equivalent (Fig. 2B). Moore and Gunderson (1995) concluded that pressures must have been lithostatic during vein formation and that the fluids were magmatic in origin. Pressures between 0.8 and 0.9 kb were estimated from the data for the formation of tourmaline and biotite veins in some of the deepest samples of the biotite hornfels. We suggest that the hornfels was formed when rocks were within the ductile zone.

As the rock cooled magmatic fluids were replaced by crustal and locally derived meteoric water. In well DV-2, which penetrated a shallow portion of the pluton, crustal waters overprinted the early tourmaline + quartz veins and deposited veins of quartz + epidote. Fluid inclusions trapped within the quartz indicate that the late mineralization occurred at a temperature of about 300°C. These inclusions record salinities of up to 6.0 weight percent NaCl. In L'Esperance –2, $^3He/^4He$ ratios of 0.5 Ra record the incursion of crustal waters into the deep portions of the biotite hornfels where present-day temperatures exceed 240°C. Although the temperature of this incursion is not known, minimum trapping temperatures of the fluid inclusions, assuming hydrostatic pressures as in the case of DV-2, would have been approximately to 290°C.

3. HIGH-TEMPERATURE ZONE

The High-Temperature Reservoir (HTR) at The Geysers is an intriguing thermal zone that may have implications for processes in the deeper parts of other geothermal systems. Walters *et al.* (1992) have described the relationships within the HTR, and the following points are taken from that paper. Although the temperature relationships are poorly defined, the HTR is characterized by high temperature, with flowing steam at 347° C recorded in one well. In addition, the HTR is characterized by a relatively high gas concentration. The HTR is most commonly encountered in the biotite hornfels, although normal reservoir temperatures and pressures can also be present in the hornfels zone. Interestingly, heat transport within the HTR appears to have a large conductive component. However, steam appears to be the pressure-controlling phase and there are no significant pressure differences between the HTR and normal graywacke reservoir. In addition, there are no mineralogical differences between the HTR and the normal graywacke reservoir. Figure 7 of Walters *et al.* (1992) shows that the depth to the HTR is deeper along NW and NE trends that define the principal structural trends in the reservoir (Nielson and Nash, 1997). It was the conclusion of Walters *et al.* that the HTR was a relict of the previous liquid-dominated system where there was still a significant component of conductive heat transfer.

Although temperatures within the HTR are high, where measured they are too low to support the ductile flow of rock. The most important evidence of this is that the wells into the HTR continue to sustain geothermal production over time. However, the top of the HTR is an important physical boundary at The Geysers, and we speculate that it may be significant in other systems as well. Indeed, we note the coincidence of the highest temperature measured at The Geysers with the maximum solubility of quartz (Fournier, 1985, 1991). The temperature profile of the WD-1 well at Kakkonda (Nielson and Rose, 1999) suggests that the bottom of hydrothermal circulation occurs at temperatures of ~340° C. In the Tiwi system, maximum fluid inclusion homogenization temperatures of about 340° C are found at the level where both vein intensity and temperatures decrease (Nielson *et al.*, 1996).

4. CONCLUSIONS

In summary, the fluid-inclusion and mineralogical data demonstrate that there were major differences in the behavior of the early shallow and deep regimes. The most significant differences are reflected in the pressures and fluid compositions. The absence of hypersaline or high-salinity fluids in the rocks above the biotite hornfels argue that there was little interaction between the early magmatic and overlying crustal waters during the period when the deep regime was ductile. We believe that ductile conditions can generally be expected at temperatures above 370° C.

The presence of the HTR at The Geysers and decrease of permeability at depth in other geothermal systems, suggests that there may also be an important physical boundary at temperatures of about 340° C. The cause and character of this boundary is speculative at the present time. We note the general coincidence with the maximum solubility of quartz; however, studies necessary to document important phenomena at this temperature have not been done.

We are also intrigued by the energy potential of zones beneath active hydrothermal circulation where heat conduction is the dominant energy transfer process. Extraction of energy from these zones will require deep injection. It is likely that injection will have to be initiated at ambient temperatures of less than 340° C to access interconnected fracture systems.

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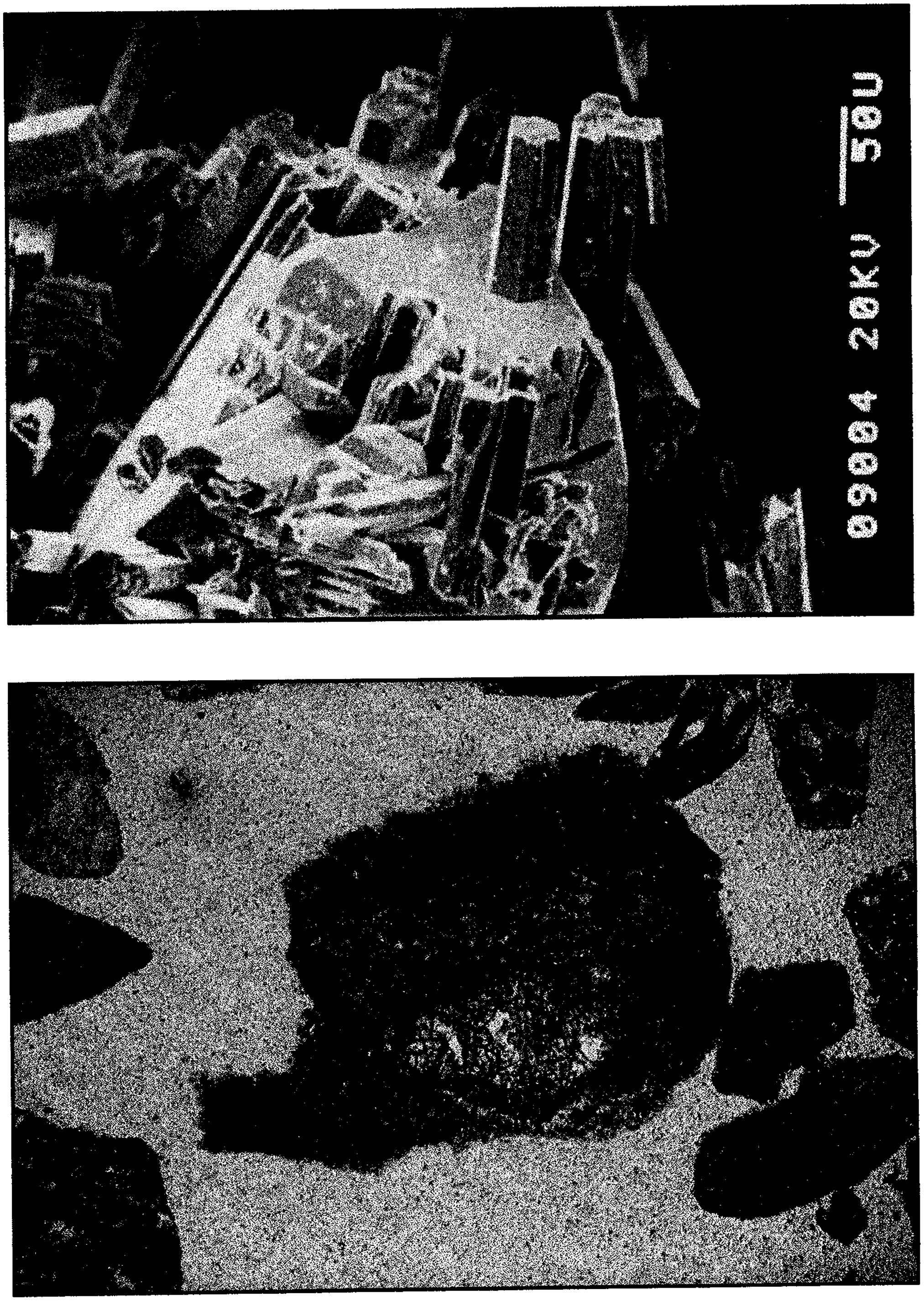

A

B

Fig.1. A) SEM image of tourmaline and potassium feldspar assigned to vein paragenesis 1. The vein was deposited in the pluton. B) Drill chips of biotite hornfels. The reddish brown mineral is biotite; the high relief green mineral in the central chip is diopside that was deposited in a vein. DLN99044

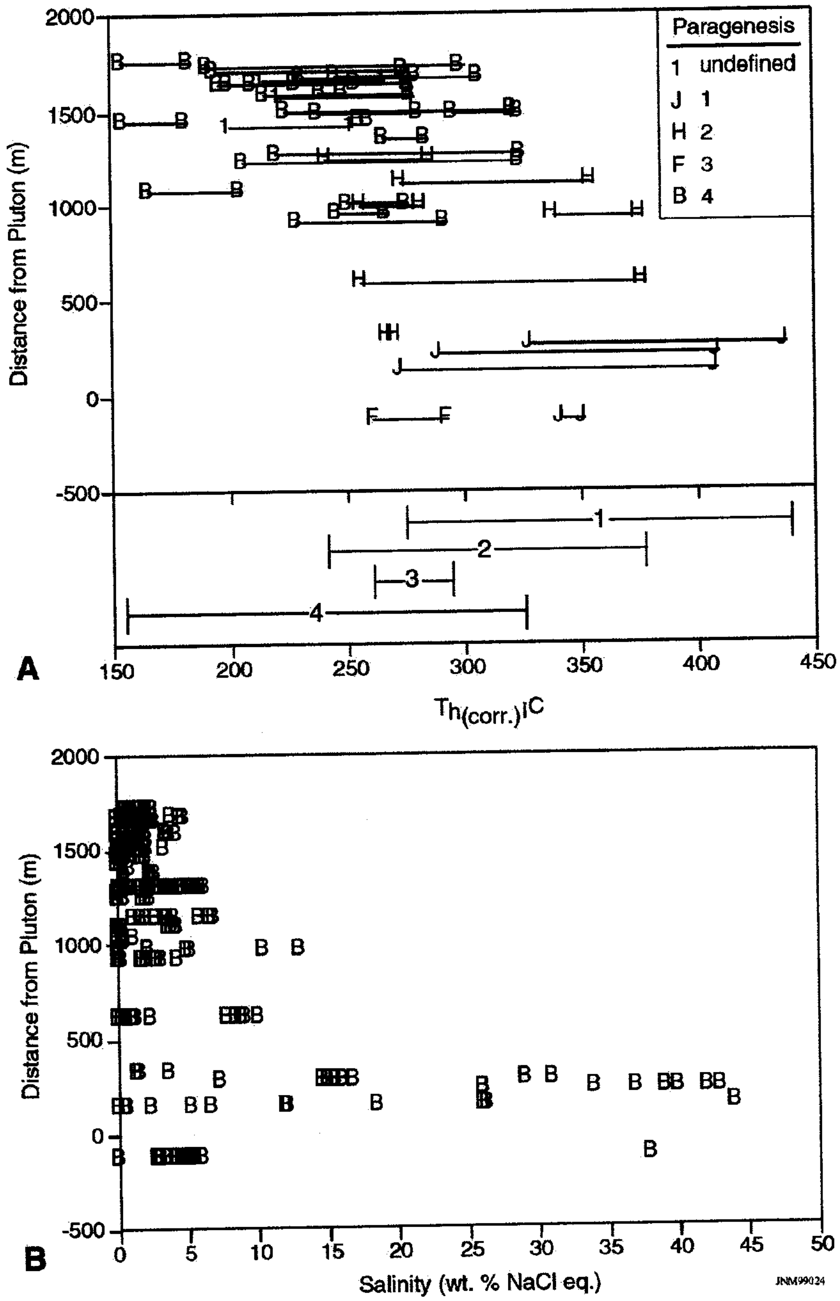

Fig. 2. A) Pressure corrected homogenization temperatures ($Th_{(corr.)}$°C) plotted with respect to distance from the pluton. Veins assigned to paragenesis 1 contain tourmaline or biotite deposited in the biotite hornfels or intrusive rocks. Paragenesis 2 contains actinolite; paragenesis 3 contains epidote; paragenesis 4 contains calcite. See Moore and Gunderson (1995) for a full mineralogic description. B) Salinity in weight percent NaCl equivalent (wt. % NaCl eq.) as a function of distance to the pluton.

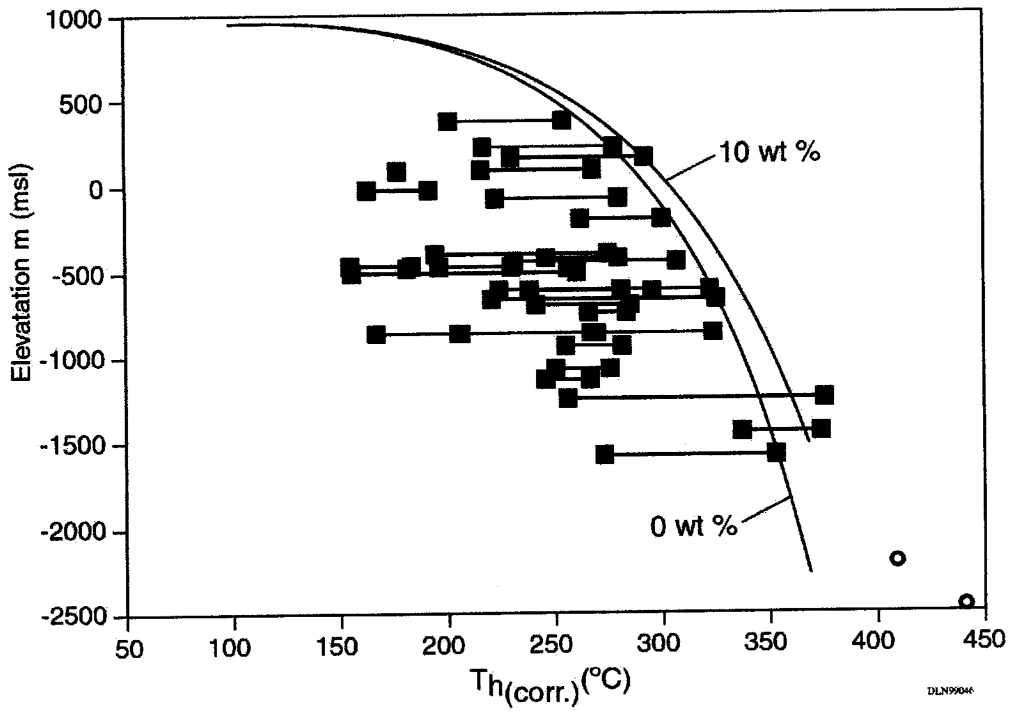

Fig. 3. Pressure corrected homogenization temperatures (Th(corr.)°C) of fluid inclusions plotted in meters against the sample elevations relative to mean sea level (msl). Pressures used to correct the temperatures were assumed to be hydrostatic for the case of inclusions formed above the biotite hornfels (squares). Minimum and maximum temperatures are shown for these samples. Boiling point curves for gas-free 0 and 10 weight percent NaCl solutions are shown. The water table was assumed to be at an elevation of 950 m above mean sea level. Circles show maximum temperatures of inclusions from the biotite hornfels. Trapping pressures were estimated independently from inclusions containing halite and vapor bubble. It is apparent that these inclusions must have been trapped at pressures greater than hydrostatic. Modified from Moore and Gunderson (1995).