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THE THERMAL CONDUCTIVITY OF ROCK UNDER HYDROTHERMAL CONDITIONS: MEASUREMENTS AND APPLICATIONS

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ABSTRACT

The thermal conductivities of most major rock-forming minerals vary with both temperature and confining pressure, leading to substantial changes in the thermal properties of some rocks at the high temperatures characteristic of geothermal systems. In areas with large geothermal gradients, the successful use of near-surface heat flow measurements to predict temperatures at depth depends upon accurate corrections for varying thermal conductivity. Previous measurements of the thermal conductivity of dry rock samples as a function of temperature were inadequate for porous rocks and susceptible to thermal cracking effects in nonporous rocks. We have developed an instrument for measuring the thermal conductivity of water-saturated rocks at temperatures from 20 to 350 °C and confining pressures up to 100 MPa. A transient line-source of heat is applied through a needle probe centered within the rock sample, which in turn is enclosed within a heated pressure vessel with independent controls on pore and confining pressure.

Application of this technique to samples of Franciscan graywacke from The Geysers reveals a significant change in thermal conductivity with temperature. At reservoir-equivalent temperatures of 250 °C, the conductivity of the graywacke decreases by approximately 25% relative to the room temperature value. Where heat flow is constant with depth within the caprock overlying the reservoir, this reduction in conductivity with temperature leads to a corresponding increase in the geothermal gradient. Consequently, reservoir temperatures are encountered at depths significantly shallower than those predicted by assuming a constant temperature gradient with depth. We have derived general equations for estimating the thermal conductivity of most metamorphic and igneous rocks and some sedimentary rocks at elevated temperature from knowledge of the room temperature thermal conductivity. Application of these equations to geothermal exploration should improve estimates of subsurface temperatures derived from heat flow measurements.

INTRODUCTION

Heat flow measurements require accurate determinations of both the vertical geothermal gradient and the in situ thermal conductivity. Although most major rock-forming minerals exhibit substantial changes in thermal conductivity with increasing temperature and modest changes with increasing pressure (Birch and Clark, 1940; Van Buskirk et al., 1985), these temperature and pressure effects are generally insignificant over the approximately 100 meter depth range of most heat flow holes. Unless near surface conditions are anomalously cold (e.g. permafrost) or warm (e.g. hot springs) room temperature thermal conductivity measurements are adequate.

For boreholes deeper than 1 km, particularly in regions of high geothermal gradients, temperature effects become important. In addition, the role of both temperature and pressure becomes significant when near-surface temperature measurements are extrapolated to greater depth. For example, failure to account for changing thermal conductivity with depth can lead to errors in estimating the depth to particular reservoir temperature or mask changes in heat flow with depth which could lead to misinterpretations of heat and fluid flow conditions within an hydrothermal system.

With the exception of the work by Van Buskirk et al. (1985) on porous tuffs, previous studies on the variation of thermal conductivity with temperature and pressure have focused on igneous and metamorphic rocks with porosities less than 5% and air as the confining and saturating medium (e.g. Buntebarth, 1991; Kawada, 1964). Air, with a thermal conductivity of 0.024 W/m•K at room temperature and pressure, presents a significant contrast to the rock matrix, which has typical conductivities ranging from 2 to 5 W/m•K. Even at low porosities, this leads to significant changes in thermal conductivity (Zimmerman, 1989) which are not introduced when water, with a conductivity of 0.61 W/m•K, is the saturating medium. Birch and Clark (1940) used helium (0.15 W/m•K) and nitrogen (0.025 W/m•K) as saturating fluids but the substantial thermal conductivity contrast relative to in situ conditions was not completely eliminated.

Despite these limitations, the earlier studies established that a general relationship between temperature and thermal conductivity of the form

$$\lambda(T) = \frac{\lambda_0}{(a + bT)} \tag{1}$$

where λ_0 is the thermal conductivity at 0 °C, a and b are experimentally-derived constants, and T is the temperature, could be applied to almost all rock types.

Published data for most common rock types show considerable variation in a and b, precluding any generalization of equation (1) to individual rock types. However, Sass et al. (1992) noted that the constant b could be described as a function of the thermal conductivity at 0 °C (λ_0) when applied to the Birch and Clark (1940) data on crystalline rocks, such that

$$\lambda(T) = \frac{\lambda_0}{(a + cT\cdot[1 - \frac{d}{\lambda_0}])} \tag{2}$$

where $c(1 - d/\lambda_0)$ corresponds to b in equation (1). The advantage of this formulation is established by the constant values of a, c and d for most crystalline rocks. According to Sass et al. (1992), crystalline rock thermal conductivities are well-characterized by

$$\lambda(T) = \frac{\lambda_0}{1.007 + 0.0036T\cdot(1 - \frac{2}{\lambda_0})} \tag{3}$$

Consequently, with λ_0 established from room temperature measurements, equation (3) yields the thermal conductivity at any temperature up to 300 °C (the maximum temperature reached by Birch and Clark, 1940). Despite the usefulness of equation (3) for a range of crystalline rocks, it is based upon air- and gas-saturated measurements which, as noted above, are not representative of in situ conditions. A technique for measuring the thermal conductivity of porous and nonporous rocks under water-saturated conditions of elevated temperature and pressure is clearly needed.

EXPERIMENTAL APPROACH

The device employed by the USGS in Menlo Park for high temperature and pressure thermal conductivity measurements is shown in Figure 1.

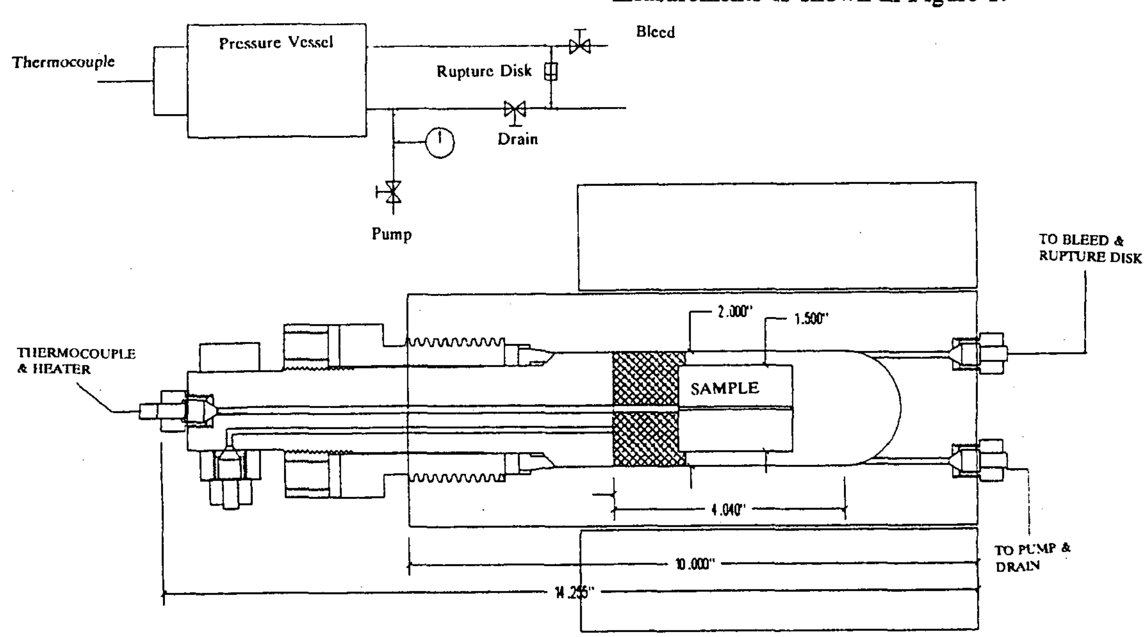

Figure 1 - Schematic of USGS apparatus for thermal conductivity measurements at elevated temperature and pressure. Detail view in lower right shows contents of pressure vessel.

The measurement technique is the transient line-source or "needle probe" method (Von Herzen and Maxwell, 1959), and the general approach is similar to those followed by Morin and Silva (1983) for deep sea sediments and Van Buskirk et al. (1985) for nuclear waste disposal studies. The primary difference with these earlier studies is the larger range of temperature and pressure and the ability to separately vary pore and confining pressure. The experimental set-up consists of a stainless steel pressure vessel with an access port for the needle probe, a ceramic-insulated needle probe in a steel housing with a loop of heater wire and a copper-constantan thermocouple, separate pressure lines for controlling pore and confining pressure, a cylindrical electric heater surrounding the outside of the vessel, and a pump and associated valves for controling fluid pressure. The pressure vessel and associated valves and tubing are rated to 350 °C and 100 MPa. Thermocouple readings are referenced to an electronic zero point outside the pressure vessel and recorded on a computer through use of a digital nanovoltmeter. Sample preparation and experimental approach are summarized as follows.

(1) Cylindrical rock samples 38 mm (1.5 inches) in diameter and approximately 51 mm (2.0 inches) in length are cut from larger samples. A 1.3 mm (0.05 inch) hole is then drilled down the central axis of the sample.
(2) The sample is lowered over the needle probe and loaded inside the pressure vessel. The sample can be jacketed to allow for differential pore and confining pressures.
(3) The vessel is filled with water and heat is applied to the outside of the vessel. Heating rates are kept below 1 °C/minute and pressure maintained at a minimum of 10 MPa (1500 psi) to reduce the risk of thermal cracking.
(4) Tests begin once the target temperature and pressure are reached and temperature drift has dropped below 1 °C per hour. An 80 mA current is applied to the heater wire and the resulting temperature rise in the sample is measured. Tests typically last 2 minutes and are repeated multiple times at each temperature point.
(5) The temperature-time data are fit to the asymptotic solution of the transient line source, which is given by

$$T(t) = \frac{I^2 R}{4\pi\lambda} \ln t \qquad (4)$$

where t is time, I is the applied current, and R is the specific resistance of the heater wire (approximately 0.7 ohms per mm). The resulting fit provides the value of λ.

Calibration of the device was established both at room temperature and over the temperature range of 0 to 300 °C. At room temperature, tests were performed on a series of water-saturated rock samples ranging in thermal conductivity from 1.0 to 3.5 W/m•K. The resulting values were compared with measurements by the room temperature needle probe device described by Sass et al. (1984). The results of the room temperature cross-calibration are shown in Figure 2.

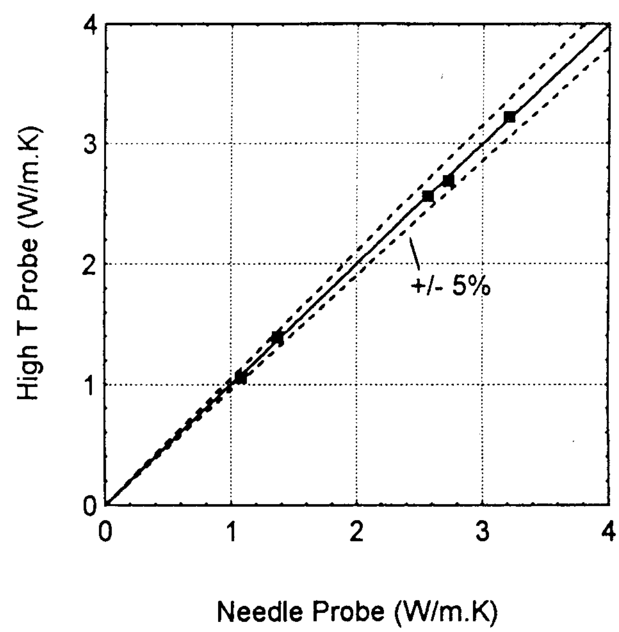

Figure 2 - Comparison of high temperature probe calibration with standard needle probe (Sass et al., 1984). Measurements made at 25 °C.

Accuracy at elevated temperatures was established by comparing measurements on fused silica with the curve established by Carwile and Hoge (1968) after a careful compilation of reliable values from the literature. Carwile and Hoge (1968) estimate an uncertainty for their results of approximately ±7%, and the USGS measurements fall within 3% (Figure 3).

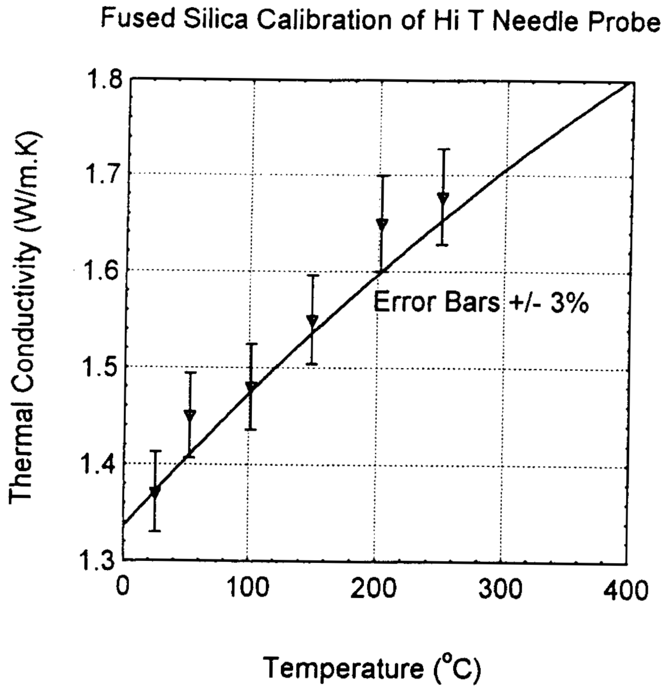

Figure 3 - Comparison of USGS thermal conductivity measurements on fused silica with curve established by Carwile and Hoge (1968).

RESULTS

The initial series of measurements was devoted to samples of Franciscan graywacke from wells at The Geysers geothermal field in northern California. Because of the predominantly conductive nature of heat transfer within The Geysers graywacke caprock, shallow heat-flow measurements are often used to estimate the depth to the top of the 240 °C vapor-dominated reservoir (Urban et al., 1976). In addition, profiles of heat flow within the caprock can indicate the magnitude and timing of thermal processes within the reservoir (Williams et al., 1993).

Williams et al. (1993) and Williams and Sass (1994) applied the thermal conductivity model of Sass et al. (1992) to the variation in conductivity of graywacke over the temperature range encountered in the caprock (15 to 250 °C) and calculated a corresponding decrease of more than 30% in thermal conductivity. This result led to the determination that heat flow was not constant with depth for those scattered locations where equilibrium temperature profiles cover the entire depth extent of the caprock. As noted above, the Sass et al. (1992) model is based upon crystalline rocks, so graywacke measurements were required to verify the Williams et al. (1993) results.

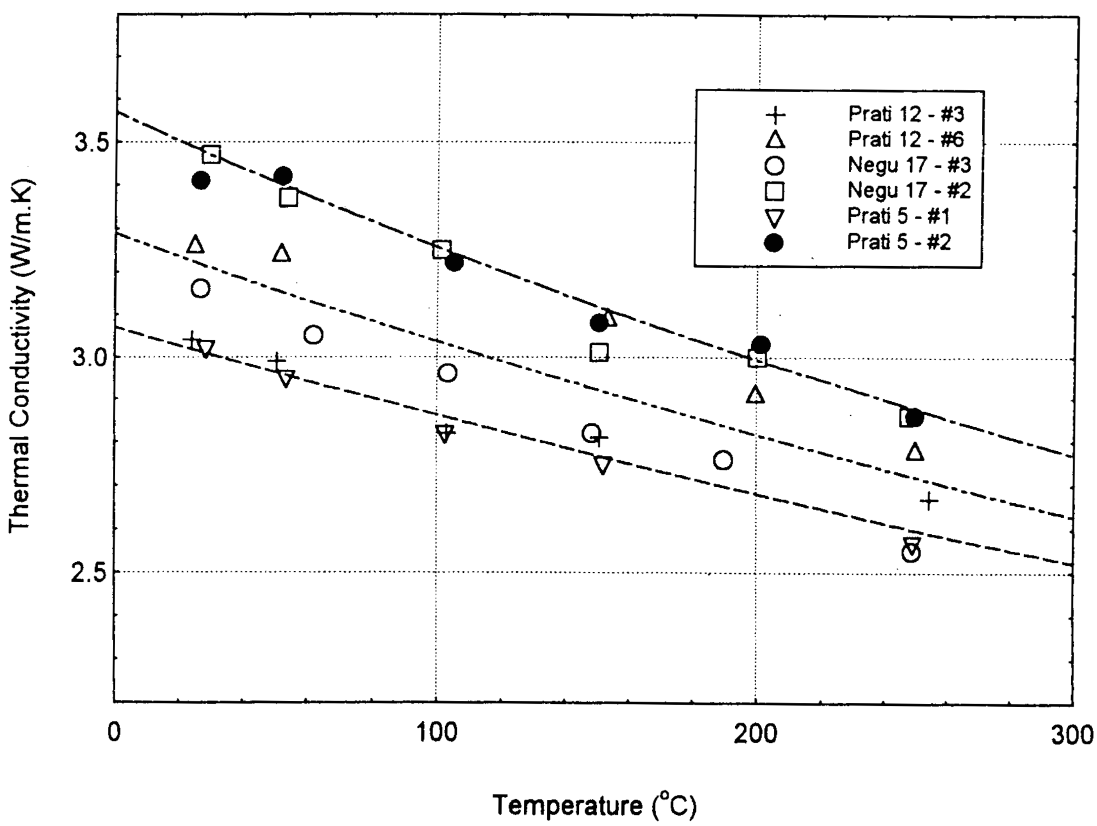

Figure 4 - Variation of thermal conductivity with temperature for six core samples of Franciscan graywacke from the Northwest Geysers. Dashed lines are illustrative curve-fits based on equation (5).

Thermal conductivity measurements on six water-saturated graywacke samples over the temperature range of 25 to 250 °C are shown in Figure 4. As expected, thermal conductivity decreases with increasing temperature, from values ranging from 3.0 to 3.5 W/m•K at 25 °C to 2.5 to 2.85 W/m•K at 250 °C. The rate of decrease with temperature is less than the rate estimated from the Sass et al. (1992) curve-fit to the Birch and Clark (1940) data (equation 3). According to equation 3, the thermal conductivities should decrease to values ranging from 2.3 to 2.6 W/m•K. A series of measurements on gneiss and amphibolite with the USGS instrument by Pribnow et al. (1996) also shows a smaller rate of decrease in thermal conductivity with temperature.

A fit of equation 2 to all of the data in Figure 4 yields

$$\lambda(T) = \frac{\lambda_0}{1.0 + 0.0024T \cdot (1 - \frac{2.16}{\lambda_0})} \quad (5)$$

Curves derived from equation 5 and bracketing values of λ_0 are shown in Figure 4. Deviations from equation 5 average less then 0.05 W/m•K. The difference between these graywacke measurements and the Birch and Clark (1940) crystalline rock measurements is not due to porosity within the graywacke. The graywacke sample porosity ranges from 0.5 to 3.5%, which largely overlaps the porosity range for crystalline rocks, and the variation of water thermal conductivity with temperature is relatively modest. Between its freezing point and critical point, the thermal conductivity of liquid water varies over a fairly small range (Figure 5; Bolz and Tuve, 1971). In the context of crystalline rock with porosity in the range of 0 to a few percent, the temperature variation has a very small effect on the whole rock thermal conductivity and is very easy to include in the calculation. Of far greater importance is the presence or absence of water in pores and cracks during the measurement process.

The most likely explanations for the discrepancy are either (1) problems with the Birch and Clark (1940) measurements and others made on dry rock samples or (2) fundamental differences between the thermal properties of quartz-rich igneous rocks and quartz-rich metamorphic and sedimentary rocks. The available information favors explanation 1, in that there seems to be no obvious reason for the constituent minerals to have different conductivities. On the other hand, thermal cracking is commonly observed when multimineral assemblages are heated up, even under confining pressure. These new cracks tend to lower conductivity, and the effect is much more pronounced if they are filled with low conductivity vapor rather than liquid water (Pribnow et al., 1996). Measurements on a series of igneous rocks corresponding to those studied by Birch and Clark (1940) should resolve the issue and will be conducted in the near future.

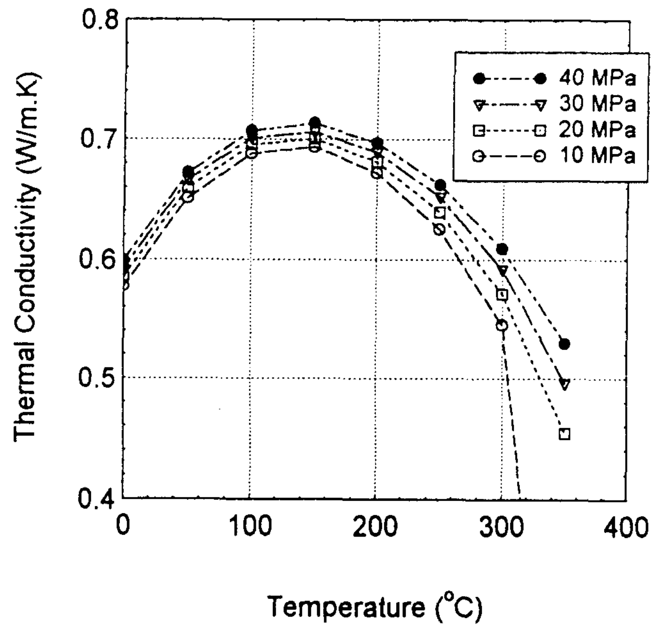

Figure 5 - Variation of the thermal conductivity of water with increasing temperature and pressure as tabulated by Bolz and Tuve (1971).

The high temperature graywacke measurements were applied to the temperature and conductivity data collected by Williams et al. (1993) in the northwest Geysers. Figure 6 shows the temperature gradient, thermal conductivity, and resulting heat flow for well Prati 31. Although the decrease in thermal conductivity with depth based upon actual measurements is less than that predicted by equation 3, the heat flow profile still reflects a substantial decrease in heat flow with depth. These results confirm the importance of accurately correcting thermal conductivity measurements for the influence of temperature.

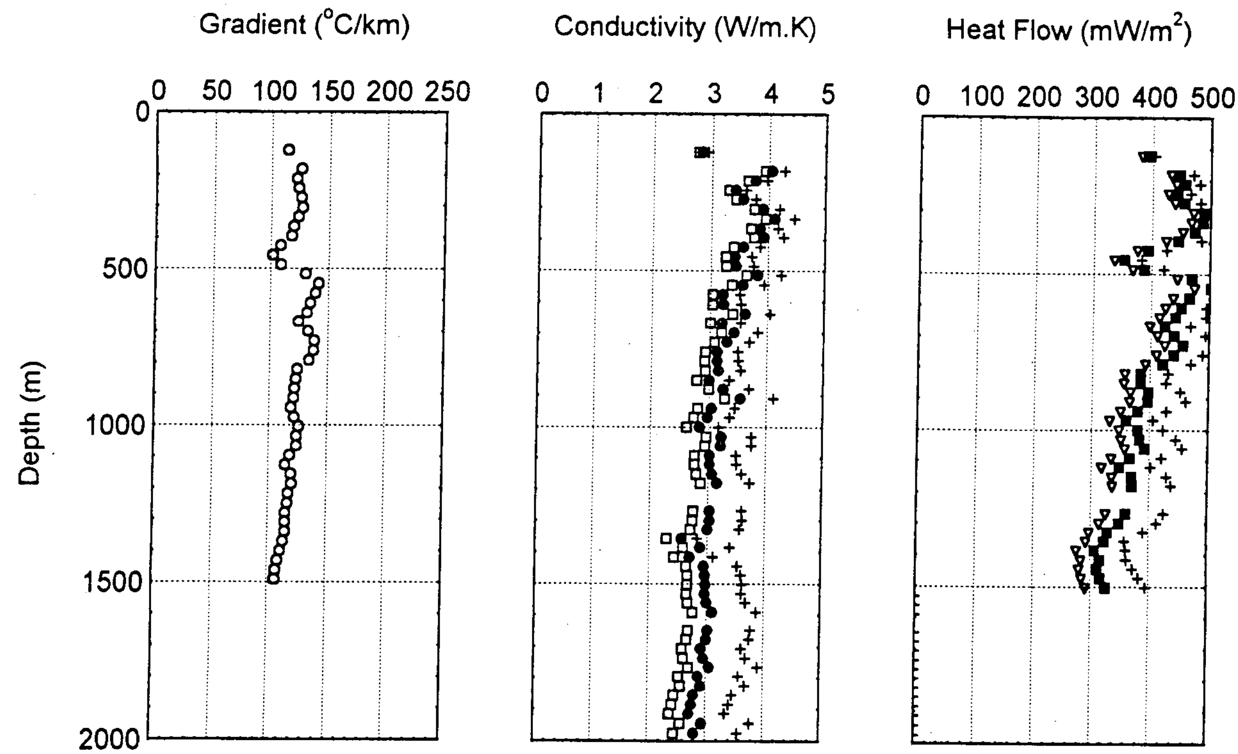

Figure 6 - Temperature gradient, thermal conductivity and heat flow from well Prati State 31. Crosses are room temperature measurements, open squares are temperature-corrected results from Williams et al. (1993) and solid symbols are corrected values based on equation (5).

CONCLUSIONS

A device for measuring the thermal conductivity of rocks under conditions of elevated temperature and pressure has been developed. Measurements on samples of Franciscan graywacke (summarized in this report) and metamorphic rocks (summarized in Pribnow et al., 1996) confirm the general decrease in conductivity with increasing temperature. This temperature-dependence can be fit to a modified form of the equation presented by Sass et al. (1992) (equation 5).

Consistency of results from the USGS apparatus suggest that equation 5 should be applicable to a wider range of rock types. For rocks with more substantial porosity, equation 5 should be used to determine the matrix conductivity after correcting for the effects of porosity via a geometric mean model of

$$\lambda = \lambda_m^{(1-\Phi)} \cdot \lambda_w^{\Phi} \qquad (6)$$

where λ_m is the matrix thermal conductivity, λ_w is water fluid conductivity, and Φ is the fractional porosity.

The downward continuation of surface heat-flow measurements, whether for geothermal exploration or for estimating temperatures in the middle and lower crust, should take these temperature effects into account.

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