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**AUTHOR(S):** W. L. Sibbitt, J. G. Dodson, and J. W. Tester

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**los alamos scientific laboratory**
**of the University of California**
LOS ALAMOS, NEW MEXICO 87545

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THERMAL CONDUCTIVITY OF ROCKS ASSOCIATED WITH ENERGY EXTRACTION FROM HOT DRY ROCK GEOTHERMAL SYSTEMS*

W. L. Sibbitt, J. G. Dodson, and J. W. Tester

Los Alamos Scientific Laboratory, Univ. of California

Los Alamos, New Mexico 87545

## Abstract

Because the lifetime of and heat extraction rate from a hot dry rock (HDR) geothermal reservoir can be substantially controlled by the in situ rock thermal conductivity, information concerning the dependence of thermal conductivity on moisture content and temperature is important for proper design and management of the reservoir. Results of thermal conductivity measurements are given for 14 drill core rock samples taken from two exploratory HDR geothermal wellbores (maximum depth of 2929 m (9608 ft) drilled into Precambrian granitic rock in the Jemez Mountains of northern New Mexico. These samples have been petrographically characterized and in general represent fresh competent Precambrian material of deep origin. Thermal conductivities, modal analyses and densities are given for all core samples studied under dry and water-saturated conditions. Additional measurements are reported for several sedimentary rocks encountered in the upper 760 m (2500 ft) of that same region. A cut-bar thermal conductivity comparator and a transient needle probe were used for the determinations with fused quartz and Pyroceram 9606 as the standards. The maximum temperature range of the measurements was from the ice point to 250°C. The measurements on wet, water-saturated rock were limited to the temperature range below room temperature. Conductivity values of the dense core rock samples were generally within the range from 2 to 2.9 W/mK at 200°C. Excellent agreement was achieved between these laboratory measurements of thermal conductivity and those obtained by in situ measurements used in the HDR wellbores. By using samples

*Work done under the auspices of the U.S. Department of Energy.

of sufficient thickness to provide a statistically representative heat flow path, no difference between conductivity values and their temperature coefficients for orthogonal directions (heat flow parallel or perpendicular to core axis) was observed. This isotropic behavior was even found for highly foliated gneissic specimens. Estimates of thermal conductivity based on a composite dispersion analysis utilizing pure minerallic phase conductivities and detailed modal analyses usually agreed to within 9% of the experimental values.

## INTRODUCTION AND SIGNIFICANCE

The objective of this work was measurement of the thermal conductivity of deep crustal crystalline rock obtained from cores taken at depths to 3 km. This material is unique in the sense that it is free of surface weathering, reasonably competent with well-sealed microfractures (permeability <1 μdarcy) and is part of a hot dry rock (HDR) geothermal reservoir system in Precambrian granite at 200°C. Detailed information on the dependence of thermal conductivity on temperature and moisture content as well as on mineralogy was required for predictive performance modeling of an HDR reservoir for the following reasons [Smith, 1975].

The operational feasibility of hot dry rock (HDR) geothermal systems requires that heat be transferred efficiently from hot reservoir rock to a circulating fluid (water). In most of the concepts being pursued that involve fractured rock reservoirs, the thermal conductivity of the formation critically affects the lifetime and thermal capacity of the reservoir. [Tester and Smith, 1977]. Of equal importance are the amount of surface area accessible to fluid circulating across the fracture faces and the rate of fluid flow. As given by Murphy [Tester and Smith, 1977], the recoverable power, P(t), in watts, for a single ideal disc shaped fracture system in granite can be estimated for uniform water flow conditions as:

$$P(t) = \dot{m}_w C_w (T_i - T_{min}) \operatorname{erf}\left(\sqrt{\frac{(\lambda\rho C)_r}{t}}\ \frac{\pi R^2}{\dot{m}_w C_w}\right) \qquad (1)$$

where

$R$ = fracture radius, m
$C_r$ = heat capacity of granite, ∿1000 J/kgK [Clark, 1966]
$C_w$ = heat capacity of water, ∿4200 J/kgK
$\dot{m}_w$ = water flow rate through fracture, kg/sec
$t$ = time, sec
$T_i$ = mean initial rock temperature, °C

$T_{min}$ = fluid temperature as it enters the reservoir, °C
$\lambda_r$ = thermal conductivity of granite, W/m K
$\rho_r$ = rock density, $kg/m^3$

The error function term in eq (1) describes the finite thermal resistance for heat transport by transient conduction through the rock to the circulating fluid. Even in situations with non-uniform flows and non-ideal fracture geometries, recoverable power levels will depend on the parametric form of eq (1). [Murphy and McFarland, 1976; Harlow and Pracht, 1972.] Thus in order to predict reservoir performance, one must have reasonable values for the thermal conductivity of the rock formation.

In addition to influencing reservoir performance, and therefore reservoir-related costs, thermal conductivity also affects the economics of HDR systems in another way. For a given regional heat flow, the thermal conductivity of the overlying rocks will control the mean geothermal temperature gradient observed. Thus the drilling depth required to reach rock of a given temperature will vary inversely with the mean gradient. Because drilling costs increase exponentially with depth [Tester and Smith, 1977] thermal conductivity-heat flow characteristics are important in determining costs associated with developing the reservoir.

The conventional definition of thermal conductivity $\lambda$ using the Biot/Fourier formulation of unidirectional conductive heat flow has been applied to mineral rock systems [Birch and Clark, 1940]. Because rocks are all diathermanous materials, the thermal conductivity $\lambda$ of laboratory-sized samples can depend upon specimen thickness as well as upon specimen composition and its intrinsic physical condition. All rocks transmit infrared radiation; therefore, there are always two parallel paths for heat. Thus $\lambda$ has been called an "apparent conductivity" since it is not a true physical property of the material. The apparent conductivity is the sum of the true conductivity (a physical property) and the "radiation conductivity." The contribution due to radiation however is usually negligible for temperatures below 800 K and does not influence the data presented in this paper.

Heat flow in rock composites is rather complex: it is always three-dimensional on a macroscopic scale although in many cases it can be treated as unidirectional flow. In a typical granitic rock of interest to our geothermal project, microcapillaries occur along mineral contacts and along cleavage planes of feldspars and iron-magnesium silicates (biotites). Quartz tends to form microcracks around and across quartz grains. The volume fraction of these voids is usually small (less than 0.001) but the interfacial surface is very large and substantially reduces the apparent radiation conductivity but only moderately reduces the apparent lattice conductivity. For situations where confining pressures are sufficiently high, microcracks closure can be complete and result in almost no change in phonon transport. In almost any case the intercrystalline resistance to heat flow in relatively fresh, wet igneous rocks is low at moderate temperatures. [Walsh and Decker, 1966].

If apparent conductivity values are to be used to estimate the geothermal heat flux, then experimental rock specimens should be of such a size that average conductivity values will be characteristic of large masses of rock eliminating the effects of microscopic heterogeneities. The thickness of the specimen should be of at least an order of magnitude greater than the grain size.

Practical problems, however, influence the choice of a method for measuring the thermal conductivity of rocks. Larger sized specimens usually result in longer measurement times. In addition, igneous rocks are difficult to machine: complicated shapes and especially long holes of small diameter present problems. Flat circular discs can be obtained by coring, sawing, grinding, and then lapping the surfaces to some degree of flatness. However, when the temperature gradient is imposed on an aggregate of non-cubic crystals, surface warping may limit sample flatness [Birch and Clark, 1940].

In measuring the surface temperature of a flat rock, good thermal contact between the temperature sensor and the non-metallic surface of the rock must be maintained. Errors also may be introduced by local cooling in the area of temperature sensor contact because of its generally higher conductivity than the rock. Furthermore, temperature averaging effects result when a finite sized sensor is placed in a temperature gradient.

## APPARATUS DESCRIPTION

Most of the problem areas cited above were eliminated completely or reduced in magnitude so that their effects were negligible by employing a properly designed cut-bar thermal conductivity comparator for steady state and a needle conductivity probe for transient measurements with carefully prepared samples and fully characterized standard materials.

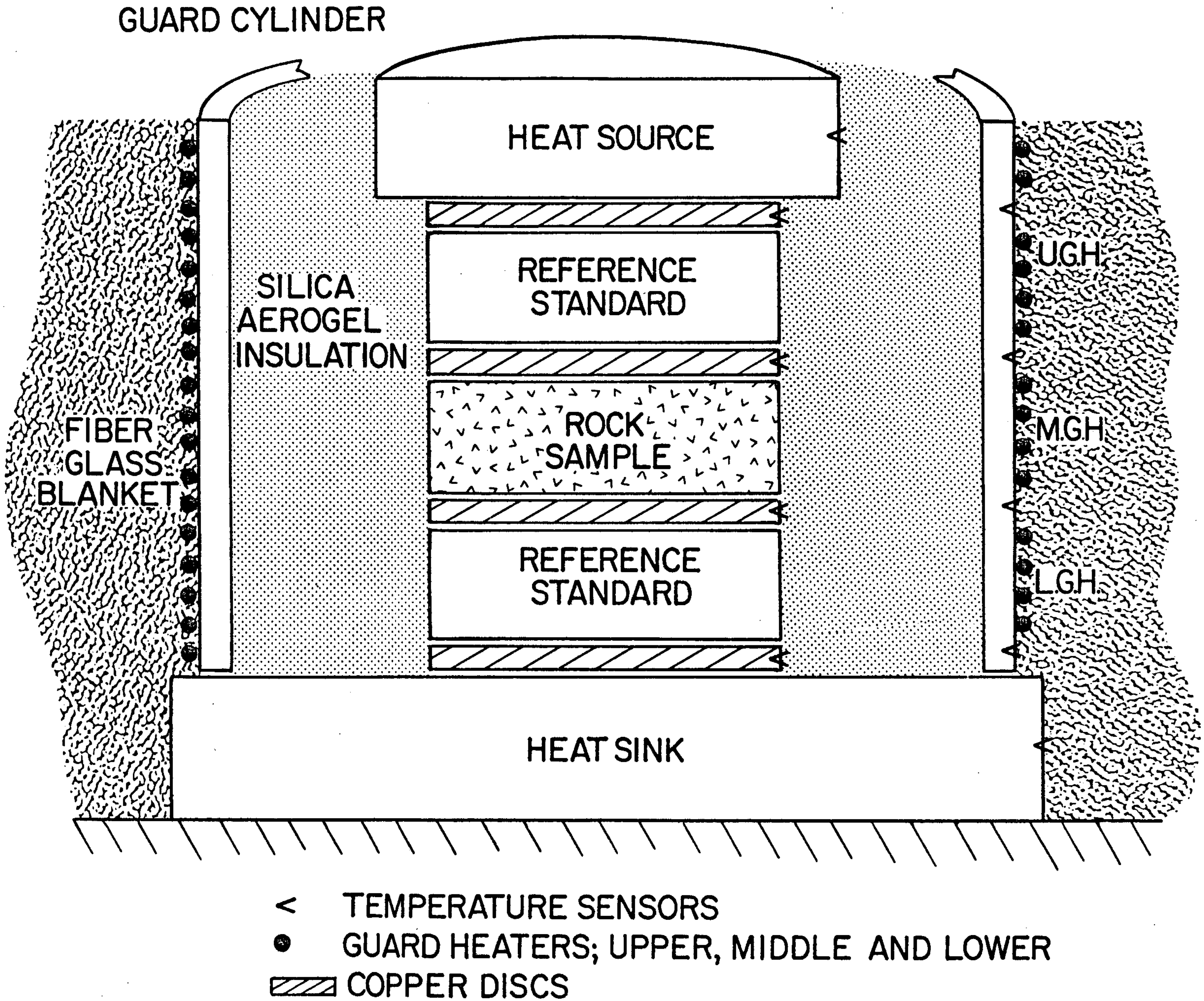
### Steady-State Measurements

Measurements of conductivities parallel to the axis of the cores were made on a guarded, steady-state, divided-bar apparatus (cut-bar thermal-conductivity comparator) [Birch and Clark, 1940]. A schematic assembly is shown in Figure 1. This comparator provides a secondary method for determining $\lambda$: a disc rock specimen of unknown conductivity is placed in series between two standard discs of known conductivity and then the temperature drops are measured with a constant, steady-state heat flux imposed. Two known standards were used to verify that the heat flow is unidirectional. The comparator consists of a sandwich structure with a rock sample fixed in place between two heat meter integral units composed of a comparison standard glued between two copper temperature sensor discs. For dry rock measurements, a high conductivity acryloid-silver composite cement was used to provide minimum contact resistance at all interfaces, as shown in Figure 1. For wet rock measurements, a loaded water-glycerol-silver paste was used. All gluing operations and conductivity measurements were carried out with an axial load maintained on the [heat source-heat meter-rock sample-heat sink] stack assembly. A pressure controlled, total reflux condenser employing different pure organic fluids (ethylene glycol, acetone, ethanol, methanol, methylene chloride) or water was used as the heat source operating at fixed temperatures over a range from 0° to 250°C. An ethylene glycol, controlled temperature bath was used as the heat sink. The space between the guard heaters and the stack of discs is filled with a silica Aerogel insulating powder. The practical working range for this comparator is from 0.1 to 10 W/m K.

There is a finite heat flux through the insulation from the top (adjacent to the heat source) to the bottom (adjacent to the sink) because of the imposed temperature gradient. This energy might be supplied, in part, by either the stack of discs or optimally by the guard heaters. Consequently, the guard heaters are maintained at a somewhat higher temperature than the disc stack at the same axial position. The approximate magnitude of the required temperature mismatch can be established by a numerical analysis for the appropriate useful ranges of thermal and geometrical parameters and/or by a series of experiments with different degrees of mismatch.

Flat disc samples were selected because they were convenient to produce in uniform sizes. The faces of all rock discs were made as nearly parallel and flat as possible. However, because of variations in hardness of the minerals comprising the composite, slight differences in surface perfection resulted. A suitable sample thickness of approximately $\sim$1 to 3 cm was used.

A constant linear temperature gradient in both the stack of discs and in the guard heaters is the ideal situation; thus, known conductivity standards should be selected to match the sample. This is a major problem since the choice of standard materials is very limited. Probably the only materials with any justifiable claim to being standards in the range of conductivities from 1 to 4 W/m K (at temperatures of interest) are Pyroceram 9606 and fused silica. The four thin discs containing the temperature sensors are made of oxygen-free, high-conductivity copper; consequently, they present a negligible resistance to heat flow and because of their thinness do not cause an important deviation from the desired linear temperature


GUARD CYLINDER
HEAT SOURCE
REFERENCE STANDARD
SILICA AEROGEL INSULATION
ROCK SAMPLE
FIBER GLASS BLANKET
REFERENCE STANDARD
U.G.H.
M.G.H.
L.G.H.
HEAT SINK
< TEMPERATURE SENSORS
GUARD HEATERS; UPPER, MIDDLE AND LOWER
COPPER DISCS

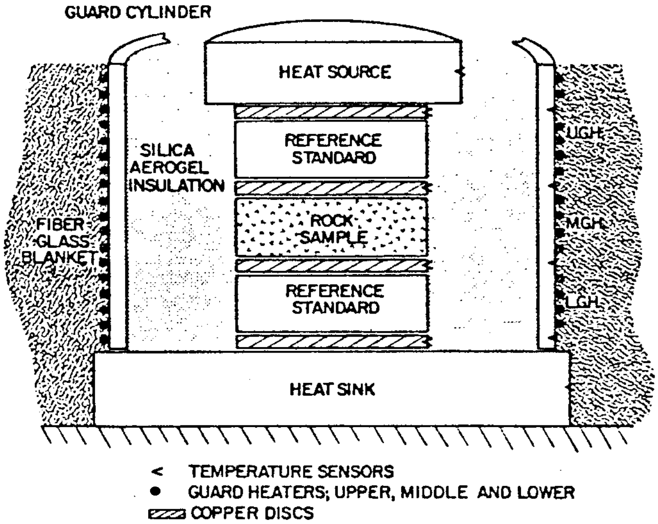



Figure 1. Schematic of cut-bar thermal conductivity comparator.

gradient. Therefore, by varying the cross-sectional areas of the standards, the temperature gradient through them can be matched to the gradient through a given rock sample.

In this modified divided-bar apparatus, two adjacent temperature sensors measure the temperature drop across one sample disc plus two interfaces (copper and rock) plus two copper discs of one-half thickness (see Figure 1). The total temperature difference across the sample and reference standard composite was usually between 10 to 20°C. In each case the measured temperature drop was corrected to obtain the actual temperature drop across the rock disc.

The measured film thickness between two small (∿40-mm-diam) fused quartz disc standards was about $2 \times 10^{-3}$ mm but increased to about $6 \times 10^{-3}$ mm for the large discs (∿60 mm diam). The conductance across such an interface in an air environment, while reasonably high is not reproducible. Consequently, a heat-transfer media was used to decrease the contact resistance and thus obtain a high, reproducible conductance at all interfaces in the stack of discs. A number of semi-liquid and liquid contact materials were tried. Excellent results were obtained with a mixture of Acryloid A-10 (Rohn and Hass) and silver pigment (Silflake 131, Handy and Harmon). This particular cement composite provides a strong bond with uniform high conductance, which can be used in ultrahigh vacuum, is bakeable to 450°C, and yet can easily be removed from the sample when necessary. Cement film thicknesses ($\leq 5 \times 10^{-2}$ mm) were measured for each experimental set-up and then the corresponding temperature drop correction was applied to the data. The corrections were of the order of 0.2 of one percent.

## Transient Measurements

A transient needle conductivity probe was used to measure conductivities perpendicular to the core axis. The needle probe is about 0.91 mm in diam by 36.5 mm in length and required that 1-mm-diam holes be drilled into samples using an expensive sonic technique.

If a line source of heat is placed in an infinite homogeneous isotropic medium initially at a uniform temperature and the heat is generated by this source at a constant rate q per unit source length then the temperature rise ($\Delta T$) (above the initial temperature) at a distance r from the line source of heat is given as a function of time t by the following equation: [van der Held, 1949]

$$\Delta T = \frac{-q}{4\pi\lambda} \, Ei \left(\frac{-r^2}{4\alpha t}\right) \tag{3}$$

where
Ei = exponential integral
$\alpha$ = thermal diffusivity
For large values of time, this may be approximated by:

$$\Delta T = \frac{q}{4\pi\lambda} \left[ \ell n t + \ell n \frac{4\alpha}{r^2} - \ell n \gamma \right] \tag{4}$$

where
$\ell n \gamma = \ell n$ (Euler's constant) = 0.5772

Therefore, for fixed values of r and $\alpha$, the temperature increases logarithmically with time. A plot of temperature rise versus the logarithm of time should give a straight line whose slope is equal to $q/4\pi\lambda$. This technique is customarily used with the line heat source method to evaluate the thermal conductivity. For probes where the heater and temperature sensor are located within the same sheath, the value of r is indeterminate; however if the probe approximates a line heat source then the value of r is immaterial as long as it is constant for a given series of measurements. This method of calculation gives correct values of conductivity under most conditions achieved in practice. However the probe diameter, length and construction materials must be selected so as to minimize a common deviation from ideal line source behavior - namely, the "initial lag effect" at short experimental times which is caused by the thermal properties of the probe (both intensive and extensive).

To a first approximation, if $t >> r_p^2/\alpha$ ($r_p$ = hole radius) then the logarithmic dependence of eq (4) should be followed. The contact resistance between the probe and sample contribute to this lag effect. An axial heat flow error occurring at long experimental times caused by non-radial heat flow from the probe to the sample also introduces errors [Wechsler, 1966]. These probes were calibrated by a series of measurements on fused quartz and Pyroceram 9606 standard samples using the $\lambda$ values recommended by the Thermophysical Properties Research Center (T.P.R.C.) [1964].

## EXPERIMENTAL RESULTS

The experimental effort cited below concentrates on thermal conductivity measurements for fresh, competent Precambrian crystalline rock core material obtained from two different deep wellbores, GT-1 and GT-2, which are part of the LASL hot dry rock geothermal field demonstration project [Tester and Smith, 1977]. The important feature we emphasize in this paper is that accurate thermal conductivity measurements are reported on samples that have been petrographically characterized in detail with corresponding modal analyses presented for all crystalline rock measurements. [See also Sibbitt, 1976].

In 1972, an exploratory hole located in the Jemez Mountains of north central New Mexico was completed to a depth of 785 m (2572 ft) (Geothermal Test Hole No. 1, GT-1). The bottom 50 m (165 ft) of the hole penetrated Precambrian basement rock and was cored continuously. The Precambrian section was characterized by a wide range of crystalline rock compositions with quartz ($SiO_2$) microcline or potassium feldspar ($KAlSi_3O_8$), plagioclase [albite ($NaAlSi_3O_8$): anorthite ($CaAl_2Si_2O_3$)] and biotite [$\sim K_2(Fe,Mg)_2(OH)_2AlSi_3O_{10}$ (with numerous substitutions)] as major constituent minerals [Perkins, 1973]. The lower 40% of this core section was essentially biotite-amphibolite veined by tonalite-aplite. The upper 60% was essentially granite and granodiorite, partly gneissic in texture. A number of thermal conductivity measurements from 0° to 250°C were made on samples from this core section (see Table 1 and Figure 3). These are averaged values based on approximately 20 determinations on each sample. The samples are designated by the indicated depth below the surface as measured from the drilling platform. The petrography of the rock types in this drill core was studied in some detail [Perkins, 1973] and modal analyses are given in Table 2.

A second hole, GT-2 located at Fenton Hill about 2.5 km (1.5 miles) south of GT-1, was drilled to a depth of 2930 m ($\sim$9600 ft) into Precambrian-age basement granitic rocks. Figure 2 is a lithologic log depicting the formations penetrated and temperature profile in GT-2. The lithology of GT-2 is approximately as follows (starting from a surface ground elevation of 2652 m (8702 ft): First a

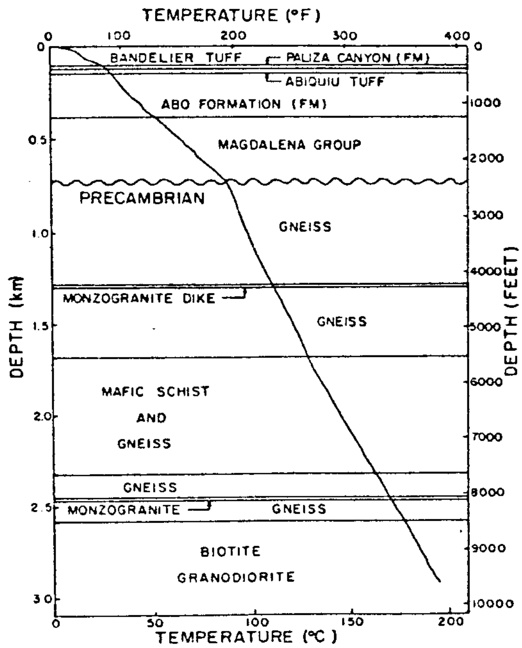



Figure 2. Lithologic log of GT-2 showing temperature versus depth.

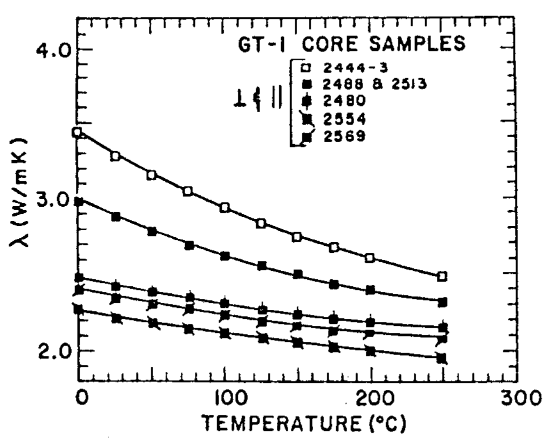



Figure 3. GT-1 core sample thermal conductivity. See Table 2 for modal compositions.

Table 1

THERMAL CONDUCTIVITY OF ROCKS FROM GT-1

($\lambda$ in $\frac{W}{m\,K}$ parallel to the axis of the core)*

| Sample No. | Temperature °C 0 | 25 | 50 | 75 | 100 | 125 | 150 | 175 | 200 | 250 |
|---|---|---|---|---|---|---|---|---|---|---|
| 2444-3 | 3.435 | 3.287 | 3.165 | 3.052 | 2.945 | 2.842 | 2.763 | 2.686 | 2.615 | 2.484 |
| 2488 | 2.978 | 2.865 | 2.777 | 2.697 | 2.627 | 2.564 | 2.503 | 2.447 | 2.397 | 2.313 |
| 2513 | 2.985 | 2.872 | 2.784 | 2.702 | 2.630 | 2.559 | 2.494 | 2.433 | 2.376 | 2.277 |
| 2524 | 2.480 | 2.423 | 2.379 | 2.338 | 2.303 | 2.267 | 2.235 | 2.207 | 2.181 | 2.140 |
| 2554 | 2.272 | 2.213 | 2.178 | 2.140 | 2.108 | 2.076 | 2.046 | 2.018 | 1.993 | 1.945 |
| 2569 | 2.407 | 2.354 | 2.310 | 2.272 | 2.238 | 2.205 | 2.176 | 2.149 | 2.124 | 2.078 |

*Measurements taken with needle probe ($\perp$ to core axis) yield $\lambda$ values within the precision of data shown for // measurements taken with cut-bar comparator.

Table 2

PROXIMATE ANALYSIS OF GT-1 ROCKS [PERKINS, 1973]

(volume percentages of minerals)

| Sample No. | K-feldspar (microcline) | Plagioclase | % An | Plagioclase altered | Quartz | Biotite | Myrmekite | Chlorite | Epidote | Sphene | Hornblende | Opaques | Rock Type | Density Ratio |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 2444-3 | 35 | 21 | 25 | 13 | 20 | 6 | | | | | | 1 | Granite, adamellite | 2.69 |
| 2488 | 25 | 28 | 30 | 5 | 32 | 8 | | | | | | 1 | (Gneissic granite, adamellite) | 2.70 |
| 2513 | 11 | 36 | 29 | 10 | 25 | 14 | 1 | 1 | 1 | | | | (Gneissic biotite-granodiorite) | 2.70 |
| 2524 | | 17 | 29 | 19 | 16 | 15 | | | | | 32 | | Biotite-amphibolite | 2.96 |
| 2554 | | 23 | 36 | 11 | 34 | 20 | | | | | 12 | | Biotite-amphibolite | 2.97 |
| 2569 | | 23 | 36 | 13 | 13 | 19 | | 1 | | 1 | 29 | | Biotite-amphibolite | 2.97 |

Pleistocene layer of Bandelier tuff followed by thin layers of Paliza Canyon and Abiquiu tuffs which terminate at a depth of about 140 m (460 ft); next a Permian layer, the Abo Formation (red beds; shale, sandstone with limestone stringers) which terminates at a depth of about 380 m (1250 ft); next the Magdalena group comprised of a Pennsylvanian layer of Madera Limestone (limestone with clay and shale layers) which terminates at a depth of 660 m (2165 ft) and a Mississippian layer of Sandia Formation (limestone with shale and sandstone layers) which terminates at a depth of 734 m (2404 ft); finally, the Precambrian granitic rocks. The Precambrian section although showing a wide range of lithologic composition can generally be characterized as a competent crystalline section of low permeability (0.01 to 1.0 μ darcy) containing natural fractures with a frequency 1-5 per cm of core which have been completely sealed by calcite and/or silica [see Laughlin and Eddy, 1977 for details].

Table 3 and Figure 4 summarize the conductivity values from 0° to 250°C as measured on selected GT-2 samples which are described in Table 4. The sample designation corresponds to the indicated depth (in feet) along the wellbore below the surface as measured from the Kelly Bushing on the drilling rig platform.

Sample 9608 was described as a foliated gneiss; therefore it was selected for conductivity determination in two directions; parallel to the axis of the core and perpendicular to the axis of core - the foliation was parallel to the axis of the core. The conductivity values and their temperature coefficients for orthogonal directions were nearly equal for a sample thickness of 3.75 cm. The conductivity parallel to the foliations increased as the sample thickness was decreased: for example, for a sample thickness of 0.8 cm the conductivity in the parallel direction was more than 20% greater than the conductivity in the perpendicular direction. Thus as we indicated earlier, measurements made on foliated rocks can be very misleading until a sufficiently thick sample is used to provide a statistically representative path for heat flow. For this core, 3.75 cm was adequate. For all other granitic samples tested, there was no observable difference between parallel (steady state cut-bar comparator) and perpendicular (transient needle probe) measurements of $\lambda$ on the same sample for thicknesses of 1 cm or more.

A number of determinations were made on sedimentary rock samples from the Abo Formation. These rocks have not yet been completely characterized. One red shale sample of compacted minute particles which immediately disintegrated in water (dry density ratio of 2.475) had a dry conductivity of 2.056 W/m K at 0°C and 2.109 W/m K at 125°C. A weakly compacted red sandstone (dry density ratio 2.355) had a dry conductivity of 2.160 W/m K at 0°C and 1.810 W/m K at 125°C. A strongly compacted red sandstone (dry density ratio of 2.407) had a dry conductivity of 3.11 W/m K at 45°C. Based on the assumption of

Table 3

THERMAL CONDUCTIVITY OF ROCKS FROM GT-2

($\lambda$ in $\frac{W}{m\ K}$ parallel to the axis of the core)*

| Sample No. | Temperature °C 0°C | 25 | 50 | 75 | 100 | 125 | 150 | 175 | 200 | 250 |
|---|---|---|---|---|---|---|---|---|---|---|
| 3-2580-43 | 3.785 | 3.615 | 3.475 | 3.330 | 3.206 | 3.098 | 2.998 | 2.903 | 2.820 | 2.680 |
| 12-4918 | 3.800 | 3.622 | 3.475 | 3.336 | 3.209 | 3.091 | 2.981 | 2.882 | 2.797 | 2.646 |
| 5964-2A | 2.900 | 2.796 | 2.714 | 2.635 | 2.565 | 2.498 | 2.440 | 2.387 | 2.341 | 2.260 |
| 6153-3A | 3.475 | 3.343 | 3.222 | 3.103 | 2.992 | 2.908 | 2.836 | 2.770 | 2.713 | 2.608 |
| 6153-3B | 3.413 | 3.264 | 3.143 | 3.026 | 2.921 | 2.882 | 2.735 | 2.653 | 2.584 | 2.466 |
| 17-6156-1 | 2.908 | 2.866 | 2.777 | 2.693 | 2.625 | 2.560 | 2.503 | 2.446 | 2.393 | 2.292 |
| ($\lambda$ in $\frac{W}{m\ K}$ perpendicular to the axis of the core)* | | | | | | | | | | |
| 8579-1 | 3.125 | 3.062 | 2.990 | 2.916 | 2.852 | | 2.750 | | 2.660 | 2.595 |
| 9608-1 | 3.115 | 3.005 | 2.906 | 2.813 | 2.728 | | 2.587 | | 2.473 | 2.376 |

*$\perp$ (needle probe) and // (cut-bar comparator) measurements yield results within the precision of data shown.

Table 4
PROXIMATE ANALYSIS OF GT-2 ROCKS [LAUGHLIN AND EDDY, 1977]
(volume percentages of minerals)

| Sample No. | K-feldspar | Plagioclase | % An | Quartz | Biotite | Chlorite | Muscovite | Myrmekite | Eipdote | Amphiobole | Opaques | Sphene | Rock Type/Location | Density Ratio |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 3-2580-43 | 36±2 | 32±2 | 34 | 25±2 | 1 | 3±1 | 1 | | | | 1 | | Leucocratic Monzogranite gneiss | 2.661 |
| 12-4918 | 29±3 | 29±3 | 30 | 36±3 | 3±1 | 1 | | 1 | | | | | Leucocratic Monzogranite gneiss | 2.648 |
| 5964-2A*** | 1 | 40±3 | 38 | 6 | 2 | | | | | 47±3 | 1 | | Amphibolite | 2.882 |
| 6153-3A | 22±2 | 42±3 | 33 | 28±3 | 5±1 | | 1 | | | | | | Leucocratic granodiorite gneiss | 2.635 |
| 6153-3B | 22±2 | 42±3 | 33 | 28±3 | 5±1 | | 1 | | | | | | Leucocratic granodiorite gneiss | 2.630 |
| 17-6156-1 | 9±2 | 46±3 | 37 | 35±3 | 6±1 | | 1 | | | | 2 | | Granodiorite gneiss | 2.727 |
| 8579-1 | 21±3 | 34±3 | 31 | 29±3 | 11±2 | | | 1 | 1 | | 1 | 2 | Biotite granodiorite | 2.723 |
| 9608-1 | 12±1 | 43±2 | 36 | 31±2 | 10±1 | | | | 1 | | 2 | | Biotite granodiorite gneiss | 2.715 |
| Tonalite* | - | 50 | 45 | 28 | 15 | | | | | 7 | | | Val Verde, CA | 2.735 |
| Westerly* | 33 | 40 | 100 | 19 | 6 | | | | | | | | Granite-Westerly, RI | 2.643 |
| Rockport* | 64** | - | - | 28 | | | | | | 6 | | | Granite-Rockport, MA | 2.610 |
| Quartz Monzonite* | 27 | 33 | 27 | 34 | 4.5 | | | | | 0.2 | | | Porterville, CA | 2.637 |

*From Birch and Clark [1940].
**Microperthite.
***Modal analysis actually from amphibolite 5983-3B.

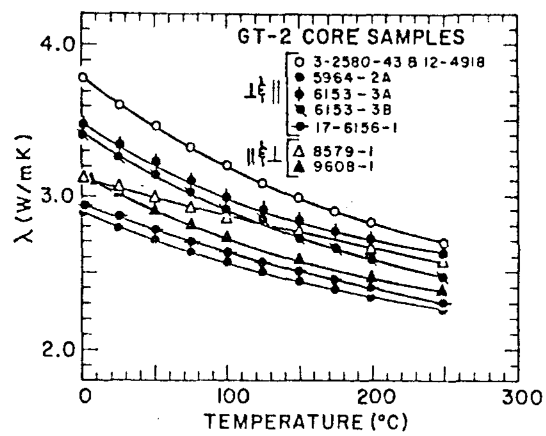



Figure 4. GT-2 core specimen thermal conductivity. See Table 4 for modal compositions.

uniform heat flux between the Precambrian and sedimentary section of GT-2, the higher average temperature gradient in GT-2 from 0 to 0.8 km of ∿100°C/km versus ∿50-60°C/km for the basement section from 0.8 to 3 km indicates that the in situ conductivity of the sedimentary section should be lower than the measured experimental values of 1.8 to 3.1 W/m K. The much higher gradient in the volcanic tuff portion is consistent with normally lower (<1.0 W/m K) conductivities associated with these low density, highly porous rocks.

## DISCUSSION

The conductivity values are given to four places to facilitate interpolation and numerical analysis; they are not an indication of the accuracy of the measurements. The cut-bar thermal-conductivity comparator and the transient needle conductivity probe are secondary instruments. Therefore the accuracy is limited by the accuracy of the conductivity values assigned to the standard comparison materials. The values recommended by T.P.R.C. for fused quartz and Pyroceram 9606 [TPRC, Purdue University, 1964] were used since they were mutually consistent in a series of comparison experiments in both instruments. The recommended values are given in Table 5. A direct comparison was made of our measurements with those made by the U.S. Geological Survey at Menlo Park, CA (see Table 6). Both LASL and USGS measurements were in agreement and were internally precise (a dispersion of the order of $\pm$ 1%) but their absolute accuracy was limited to $\pm$ 4-5% by the standards used.

The use of the term "wet" rock in Table 6 is a misnomer. Both USGS and LASL allowed the rocks to imbibe water at atmospheric pressure thus they were never completely saturated with water. Complete saturation of low-porosity rocks such as granite is very difficult. Hirschwald [1912] recommended the following procedure to obtain optimum saturation of the pores: clean the rock sample; remove the air by warming in a hard vacuum environment; let the sample imbibe water vapor for at least 3 hours in a partial vacuum equal to the vapor pressure of water at room temperature; cover with water and then apply a pressure of 50 to 150 bars. This technique probably saturates most of the open pore-volume; pores which intercommunicate and are connected to the surface including the "dead-ended" pores. However, we presently know of no method of proving that the rock is ever completely saturated.

The conductivities of these so-called "wet" rocks were of the order of 1-4% higher than the conductivities of these same dry rocks. The dry rocks were prepared by warming at 70°C in a hard vacuum for 3 hours. The temperature effects on dry rocks were reversible to at least 250°C. This is consistent with results cited by Birch and Clark [1940]. Thermal conductivity values for GT-1 and GT-2 rocks up to 250°C are plotted in Figures 3 and 4. A comparison between GT-2 $\lambda$ values in the biotite granodiorite section and data presented by Birch and Clark [1940] for a number of granitic rocks is shown in Figure 5. The gradual decrease of conductivity with increasing temperatures reflects the classical dependence expected for dense crystalline materials in the anharmonic phonon coupling region. As Birch and Clark [1940] point out, plots of thermal resistivity ($1/\lambda$) versus temperature are linear for materials of this type. The crosshatched region of Figure 5 indicates the expected variation of $\lambda$ in the biotite granodiorite section, where fluid circulation experiments are underway at LASL [Tester and Smith (1977)]. The temperature effect on thermal conductivity is similar for the selected Birch and Clark [1940] data for tonalite, Rockport and Westerly granite, and quartz monzonite in comparison to the GT-2 cores at 8580 and 9608 ft. The agreement between GT-2 9608 and quartz monzonite is fortuitous in the sense that the modal compositions are different as shown in Table 4. The Birch and Clark [1940] thermal conductivity data represent the most comprehensive study available in the high temperature region where complete modal analyses are provided. Heating above 400°C sometimes resulted in permanent changes with a slight decrease in the conductivity.

Murphy and Lawton [1977] were able to estimate <u>in situ</u> thermal conductivities of rock contained around GT-2 and a second deep wellbore, EE-1, drilled nearby. Basically, they extended the transient

Table 5
THERMAL CONDUCTIVITIES OF STANDARD MATERIALS

| Temperature (°C) | Fused Quartz TPRC[b] (W/m K) | Fused Quartz Ratcliffe (W/m K) | Fused Quartz Birch & Clark, 1940 (W/m K) |
|---|---|---|---|
| -23 | 1.28 | 1.275 | -- |
| 0 | 1.33 | 1.323 | 1.36 |
| 27 | 1.38 | 1.374 | 1.40 |
| 77 | 1.45 | 1.431[a] | 1.46 |
| 127 | 1.51 | -- | 1.51 |
| Error | ±4% | ±2% | ∿±1% |

| Temperature (°C) | Pyroceram[b] 9606 (W/m K) |
|---|---|
| 0 | 4.13 |
| 27 | 3.99 |
| 77 | 3.79 |
| 127 | 3.65 |
| 227 | 3.45 |
| 327 | 3.31 |
| Error | ±5% |

[a]Extrapolated.

[b]TPRC, Purdue University, 1964.

[c]Error given as average deviation from mean.

Table 6
COMPARISON OF CONDUCTIVITY VALUES FOR SAMPLE GT-1, 2425 FT
(Density, wet, 2.66 g/cm$^3$)

| | |
|---|---|
| USGS, wet at ∿25°C | 3.78 W/m K |
| LASL, wet, 1st run at 25°C | 3.797 ± 0.040 W/m K |
| LASL, wet, 2nd run at 25°C | 3.789 ∓ 0.035 W/m K |

NOTE: Between Runs 1 and 2, the stack was disassembled; the sample was cleaned, dried, and resaturated with water; and then the stack was reassembled.

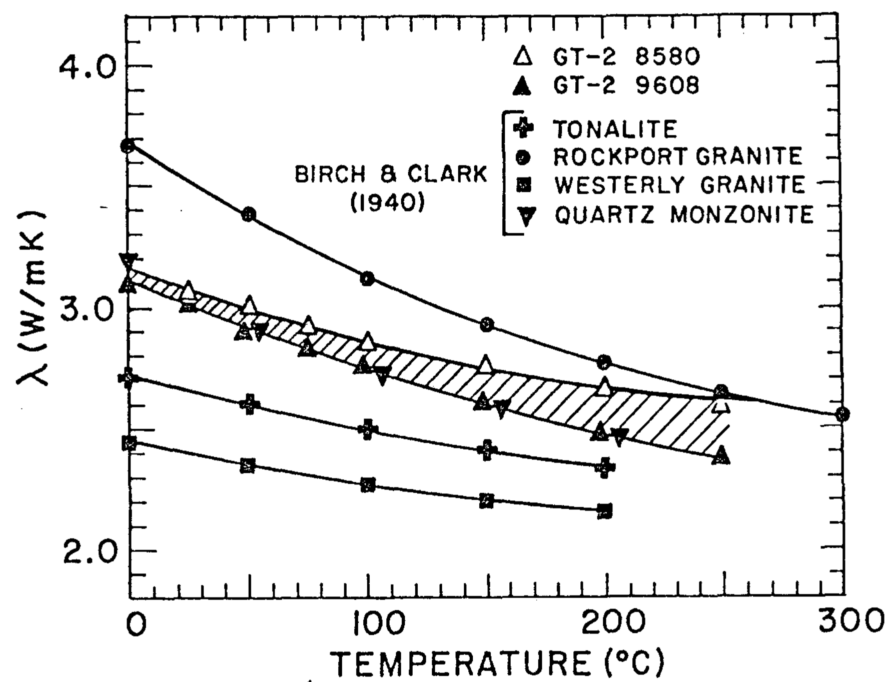



Figure 5. Comparison of GT-2 biotite granodiorite thermal conductivities with those for selected igneous rocks. See Table 4 for modal compositions.

line source method described earlier to include effects caused by flowing fluid in the wellbores. By comparing the conductive heat flux from the rock to the convective heat transported by the wellbore fluid, Murphy and Lawton [1977] showed that temperature measurements made between ∿0.25 and 100 hours provide meaningful and sufficient data for independently estimating a mean conductivity $\lambda$ and diffusivity $\alpha$ of the formation. Numerical solutions in the form of dimensionless temperature-time type curves were used to analyze experimental flowing temperature logs of the wellbores. $\lambda$ was estimated at 2.9 W/m K and $\alpha = 1.0 \times 10^{-6} m^2/sec$ [Murphy and Lawton, 1977]. By using mean values for $\rho_r = 2700$ kg/m$^2$ and $C_r = 1050$ J/kgK, a $\lambda$ of 2.8 W/m K was calculated from the estimate of $\alpha$. This internal consistency for in situ measurements coupled with the excellent agreement with the laboratory measurements of $\lambda$ supports the hypothesis that core material in terms of its thermal conduction properties is representative of the actual conditions that exist in the reservoir.

A number of simplistic schemes can be used to estimate the thermal conductivity of massive dense igneous rocks which are macroscopically isotropic and homogeneous and do not contain large volume fractions of either quartz or potassium feldspar. We applied the estimating method developed originally by Maxwell for composite materials and later modified by Birch and Clark [1940], Powers [1961], and Mitoff [1968]. Although other techniques such as the statistical approach of Hashin and Shtrikman [1962] have been applied to igneous rock systems [Horai and Baldridge, 1972] with reasonable success, we felt that variations in modal analysis for thin sections taken from the same core sample (See Tables 2 and 4) were sufficiently large to limit the accuracy of any prediction method and consequently we adopted a dispersion analysis modification of the Maxwell approach using thermal conductivity values for the pure minerallic phases as provided by Birch and Clark [1940] and others [Diment, 1967; Haskin and Shtrikman, 1962; Horai and Simmons, 1969; and Horai and Baldridge, 1972].

The discrepancies between the measured and estimated $\lambda$ values using a series combination of resistances were usually less than 9%. This type of discrepancy is acceptable since a small degree of anisotropy and inhomogeneity is inherent in solid rock samples and the modal compositions vary appreciably at different cross sections in the specimens although the density may appear to be invariant. However a single value of thermal conductivity cannot be assigned to a modal mineral when it is defined to exist over a range of chemical compositions: for example the conductivity of the plagioclase feldspar series has a minimum value at an intermediate composition of anorthite and albite [Horai and Simmons, 1969]. Variations based on chemical impurities and structural defects may also be necessary to refine estimates of $\lambda$.

Since these simplistic schemes all indicate that the conductivity of the rock is essentially determined by the volume fractions of constituent rock-forming minerals; then the crystal boundaries apparently present a low, constant resistance to the flow of heat. This generalization apparently applies to all of the accessory minerals, and the major minerals, biotite, the plagioclase feldspars, and quartz, but not to the potassium feldspar and quartz composite whose interface must present a lower heat flow resistance. The conductivity of the potassium feldspar is about 25 percent greater than that of the plagioclase feldspar series but this difference is not sufficient to explain the very high conductivity of rocks which contain large volume fractions of both quartz and potassium feldspar. Figure 6 shows a rough empirical correlation between $\lambda$ and the volume fraction of potassium feldspar and the temperature. The four different lines correlate different compositional regions for the GT-2, GT-1 and Birch and Clark [1940] data plotted in Figures 3, 4, and 5. For a given granite of fixed composition, $1/\lambda$, the resistivity, is linearly proportional to temperature; and at a fixed temperature for a particular type of granite, $\lambda$ increased linearly with the volume fraction of potassium feldspar present (x(K-feldspar)) as determined by modal analysis of thin sections. Consequently, for empirical reasons we selected x(K-feldspar)/T as a correlating parameter for $\lambda$. At very low K-feldspar concentrations of $\leq 1$ volume percent it was difficult to plot all the data because x(K-feldspar)/T varied over a small range 1 to 4 x $10^{-5}K^{-1}$. Therefore only some of the points for the low K-feldspar granites are shown in Figure 6.

If a thin specimen of a rock with a coarse texture is used, then these simplistic schemes are not appropriate: just the modal analysis alone is not sufficient to predict $\lambda$ for this specimen. The continuous phases must be identified and then the appropriate relationship can be selected to estimate the effective conductivity of the specimen.

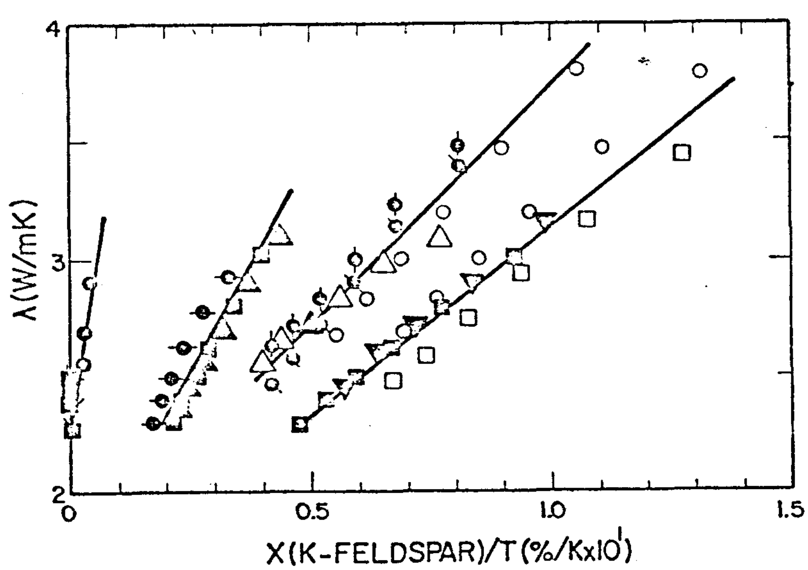



Figure 6. Empirical representation of the dependence of thermal conductivity of GT-2 and GT-1 rocks on K-feldspar volume fraction and temperature. Symbols correspond to those used in Figures 3 and 4. In addition data for quartz monzonite (▼, Birch and Clark [1940]) are presented.

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## NOMENCLATURE

$A$ = cross-sectional area perpendicular to heat flow direction, $m^2$
$C_r$ = heat capacity of granite, ∿1050 J/kgK
$C_w$ = heat capacity of water, ∿4200 J/kgK
$\dot{m}_w$ = water flow rate through fracture, kg/sec
$q$ = radial heat flux per unit length, W/m
$q_z$ = heat flow along z axis, J/sec or W
$r$ = radial distance from line source, m
$R$ = fracture radius, m
$t$ = time, sec
$T_i$ = mean initial rock temperature, °C
$T$ = temperature, K or °C
$T_{min}$ = fluid temperature as it enters the reservoir, °C
$Z$ = heat flow direction, m
$\alpha$ = thermal diffusivity, $m^2$/sec
$\lambda$ = thermal conductivity, W/m K
$\lambda_r$ = thermal conductivity of granite, W/m K
$\rho_r$ = density of granite, kg/$m^3$

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