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Experimental Permeability Studies at Elevated Temperature and Pressure of Granitic Rocks

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ABSTRACT

Permeability of quartz monzonite from the Los Alamos hot-dry-rock geothermal well GT-2 was experimentally measured as a function of pressure and temperature. Permeability apparatus slightly modified from that developed by Brace and others (1968) was used in the experiments. Flow rate through the sample was measured using a transient pressure decay method. When combined with the physical characteristics of the sample and several system parameters, this flow rate yielded permeability (k).

Permeability of the GT-2 rocks from depths of 8580 ft and 9522 ft behaves like Westerly granite for changes in effective confining pressure. However permeability of these rocks behaves much differently with increasing temperature. As temperature is increased, the permeability of Westerly granite passes through a slight minimum and then increases exponentially above 100°C. Upon cooling the permeability shows a permanent increase of up to four times its original value. The permeability of GT-2-9522', on the other hand, drops off exponentially with increasing temperature, reaching a minimum near 140°C; above 150°C, permeability rises slowly. The GT-2-8580' sample shows somewhat similar behavior, although the permeability shows a minimum near 130°C, followed by a rather sharp rise in permeability at higher temperature.

These changes in permeability with temperature are postulated to be caused by differential thermal expansion (DTE), a phenomena related to the anisotropic and inhomogeneous coefficients of thermal

expansion of the mineral grains in the rock. Scanning electron photomicrographs of unheated and heated samples of Westerly and GT-2 rocks support the DTE hypothesis. Differences in the behavior of these rocks with temperature are believed to be due to the respective temperature and pressure environments in which they became equilibrated, since both GT-2 rocks had existed at moderately high temperatures and pressures for some time. Temperature disequilibrium of the GT-2 rocks in their present in situ environments is believed to have caused the differences in the behavior between the two samples. This disequilibrium may provide a method for determining the pre-intrusion geothermal gradient of the Jemez area.

Flow channels were observed in GT-2 samples using radioactive tracer techniques. Several radioactive isotopes were tried in these experiments, including ^{22}Na , ^{63}Ni , and ^{35}S . Cores were injected with the radioisotope using the permeability apparatus; they were then thick-sectioned and autoradiographed. ^{63}Ni provided the best spatial resolution, since it emits the lowest energy electron. These channel-tracing efforts indicate that at low pressures, fluid transport through the rock is most commonly along intergranular channels. Intragranular channels were not observed with this technique.

INTRODUCTION

The Los Alamos Scientific Laboratory hot-dry-rock geothermal program was designed to create a man-made geothermal reservoir in hot crystalline basement rock. Knowledge of several physical parameters are necessary to the success of this project: 1) reservoir characteristics of the crystalline rock, including in-situ permeabilities, 2) mineralogy and petrology of reservoir rocks, and 3) rock-water interactions and their effect on the system. The down-hole reservoir is now thought to consist of two nearly parallel hydrofractures separated by a 10 meter thick zone of granodiorite with high impedance to fluid flow (Fig. 1). The development of heat-exchange capacity for this system requires flow of a large volume of water through a thick mass of crystalline rock at higher flow rates than presently obtainable. Knowledge of the permeability of the reservoir rock, changes in permeability with time, and ways to experimentally increase permeability are all relevant to the success of this design.

Many geologic phenomena have been interpreted to be the result of chemically charged fluids permeating through rock. Diagenesis, metasomatism and the formation of many ore deposits all depend on material transport either by diffusion or permeation. Since permeability is directly related to size, location and shape of flow channels in rock, knowledge of the characteristics of these channels is fundamental in understanding permeation and changes in permeability with temperature, pressure and time.

As a first step, this work was undertaken to measure in the laboratory the permeability of the reservoir rock as a function of

Figure 1. Conceptual diagram of the downhole hydrofracture system around the two drillholes, GT-2 and EE-1.

both temperature and pressure. These measurements may be useful in estimating in-situ permeabilities at different depths. A technique for measuring fairly low permeabilities over short time intervals using a transient method was also developed.

A technique for delineating flow channels in rock using a radioactive tracer was developed during this study. This tracer technique was combined with scanning electron microscope (SEM) photography to characterize flow channels with respect to shape, size, location and distribution. Changes in channel morphology due to the effects of temperature, pressure and time were also investigated.

PREVIOUS WORK

Physical properties of rock have been investigated by many authors; however, permeability and related phenomena have only been thoroughly examined for sedimentary rocks, with specific emphasis on petroleum research. However, some work is now being done on sandstones in simulated geothermal systems. Permeabilities of most sedimentary rocks are relatively easy to measure in the laboratory, hence a large volume of permeability data is found in the literature, particularly for room temperature and low confining pressures.

Several workers at Stanford University have been investigating the effects of temperature on the relative and absolute permeability of sandstone. For example, Aruna (1976) measured the permeability to water of nearly pure quartz sandstone and unconsolidated sand.

He observed a factor of two decrease in permeability as temperature was increased to 150°C. He attributed this effect to a silica-water interaction, since no decrease in permeability was observed when other fluids were used.

Extensive work has been done by W. F. Brace, J. B. Walsh, F. J. Birch and others on the relationship of pressure to physical and elastic properties of rock. They have shown that as confining pressure is increased, compressibility (Brace, 1965; Birch, 1942) and permeability (Brace and others, 1968b) are decreased, while resistivity (Brace and others, 1965, 1968a) is increased. Brace and others (1968b) have shown both an empirical and theoretical relationship between resistivity and permeability of the form:

$$k = C m^2 r_s^{-1.5} \quad , \quad (1)$$

where k is the permeability, C is a constant, m is the mean hydraulic radius and r_s is the resistivity of the fluid-filled rock. Compressibility is more closely related to porosity than permeability or resistivity, which depend more on connectivity of channels, but all parameters show a similar relationship with pressure. Brace and others' (1968b) work on permeability forms a basis for the experimental work undertaken in this thesis.

The role of temperature on physical properties is not as well documented. Temperature effects can be manifested in several ways: 1) changes brought about by the pore-fluid chemistry and response of rock to the presence of pore fluid; 2) time-related plastic and

elastic behavior of the rock (creep), and 3) expansion and contraction brought about by temperature changes.

Work in the first area is fairly extensive and deals with phenomena such as mineral dissolution and stress corrosion cracking (Martin, 1972, 1975; Anderson and Grew, 1977) and pressure solution (Sprunt and Nur, 1976) among others. The presence of pore fluid under high pressures and its effect on rock properties has been discussed thoroughly by Brace (1972). Solid state plastic behavior of rock has been rather extensively investigated, primarily from laboratory measurements. For a most complete discussion and bibliography of this subject, the reader is referred to Carter (1976).

Experimental work on thermal expansion and contraction effects in rock is limited. Skinner (1966) listed coefficients of thermal expansion for numerous minerals and rocks. Sprunt and Brace (1974) have observed thermally induced fractures in granites using a scanning electron microscope. Simmons and Cooper (1977) have recently completed work on thermal expansion in several rock types using a differential strain analysis technique. In general, the effects of thermal expansion and contraction on variables such as permeability, porosity, and elastic properties are not well understood.

Adams and Williamson (1943) were the first to recognize the profound effect of cracks and pores on elastic properties of rock. Brace (1972) has discussed the relationship of these microcavities to elastic, material transport, thermal and electrical properties. These cracks and pores are usually so small that their direct observation has presented serious problems.

Several previous attempts have been made to distinguish cracks, pores and microcavities in rock. Initial work was done using traditional thin-section preparation and optical microscopy without fracture or channel decoration (Paulding, 1965; Koide and Hoshino, 1967; Peng and Johnson, 1972; Wawersik and Brace, 1971). Baldrige and Simmons (1971) subsequently used two decoration techniques to observe channels and cracks in quartzite: 1) injected furfuryl alcohol and 2) the yellow color of Canadian Balsam. The furfuryl alcohol technique involved heating samples to 750°C after injection of the polymer; at this temperature the furfuryl alcohol was converted to carbon particles, the presence of which was supposed to identify channels and cracks.

Timur and others (1971) used the SEM to observe channels and pores in both sedimentary and igneous rocks, using standard thin-section surface preparation. Brace and others (1972) and Sprunt and Brace (1974) have described a new technique for observing micro-cavities in rock with the SEM. The method involves polishing samples by standard thin-section techniques, followed by the critical step of removing 3-6 microns of material by ion milling. Ion milling removes traces of damage incurred during polishing and produces a pristine surface which enables a close look at even the smallest of surficial features. This technique has been used by Montgomery and Brace (1975) to observe pores and microstructures in plagioclase crystals and by Richter and others (1976) to observe crack-like features in lunar samples.

EXPERIMENTAL APPARATUS

The permeability apparatus used in these experiments is similar in design to that used by Brace and others (1968b). Major modifications included the ability to measure permeability at high temperature and the ability to control pressure directly in the lower pore-pressure line.

Test samples consisted of core from the Los Alamos Scientific Laboratory geothermal test hole GT-2 at Fenton Hill, New Mexico. These samples were from a well depth of 8580 ft (2615 m) and 9522 ft (2902 m). Petrologic descriptions of these rocks have been compiled by Laughlin and Eddy (1977). The chemical and modal compositions of the test rocks are listed in Table 1. Samples were ground to right cylinders 1.57 cm long with a 4.0 cm cross-sectional area.

The core holder (Fig. 2) consisted of two hardened stainless steel pistons which were separated from the core by two 1-mm thick porous wafers of pressed stainless steel fiber. These wafers acted to distribute stress and pore fluid evenly over the surface of the sample. The entire core-holder assembly was enclosed in a 1.5-mm thick teflon jacket. Teflon was chosen as the jacketing material because it will not break down at 200°C. Several "O-Ring" grooves were cut into the pistons so that the teflon could deform and flow into them, eliminating the possibility of confining pressure leakage around the sleeve into the pore-pressure line.

The core holder assembly was placed in a larger pressure vessel of 4340 stainless steel. The lower pressure line exited the pressure

TABLE 1.
Modal compositions of test samples.

	Westerly granite*	GT-2-8580 ⁺ (2615m)	GT-2-9522 ⁺ (2902m)
Quartz	22.5	29	23
Potassium feldspar	30.7	21	20
Plagioclase feldspar	39.2	34	37
Muscovite	0.7	0.5	0.5
Biotite	5.0	11	13
Chlorite	0.4	0.5	1.0
Opaque	0.9	3.0	4.5
Others	0.6	3.0	4.5
Total	100.0	100.0	100.0
Number of counts	1000	1000	1011

* From Simmons and Cooper (1977)

⁺ From Laughlin and Eddy (1977)

Figure 2. Core holder assembly.

vessel through a Swagelok fitting. Small adjustments of this fitting permitted the application of a hydrostatic stress. A hydrostatic confining pressure of up to 415 bars (6000 PSI) at 200°C could be applied to the core. Pore pressure on both sides of the core could be controlled simultaneously, permitting any number of pore pressure-differential pressure combinations.

Figure 3 shows the general setup of the plumbing and heating fixtures of the apparatus. The large coefficient of thermal expansion of water requires extremely accurate control of system temperature. A Chromel-Alumel thermocouple was placed in the pressure vessel and was connected to a solid-state temperature controller. This controller permitted a $\pm 1^\circ\text{C}$ accuracy over the 200°C temperature range used. The controller was coupled with a resistance heater which encircled the pressure vessel. The entire assembly was placed in a large air bath (Fig. 4). This helped guarantee a nearly isothermal temperature regime. The plumbing was also insulated in styrofoam to help insure temperature stability.

The tubes, valves, fittings and gauges were all of 316 stainless steel. Three heise gauges were used to measure lower and upper pore pressure and confining pressure. Distilled water was used as the confining pressure medium. The pore fluid chosen was distilled, deaerated water. Two small Haskel pumps (0-10,000 psi range) were used to pressurize the confining and pore pressure lines.

The incremental changes in pressure needed for the permeability measurements had to be measured accurately and reliably. A pressure

Figure 3. Schematic diagram of the permeability apparatus.

Figure 4. Photograph of pressure vessel in air bath; note insulation around tubing and valves.

transducer (Precise Sensor #7710) with a sensitivity of 2.3 mv/volt was used. The transducer was coupled with a Hewlett-Packard 3419A digital voltmeter which had a resolution of 1 microvolt. Sensitivity for the entire system was better than 0.01 bar. The digital voltmeter was interfaced with a Hewlett-Packard 9815A calculator which was programmed to automatically gather data and calculate permeabilities.

PROCEDURES

Permeability

Figure 5 shows a schematic representation of the experimental apparatus. Two nearly equal fluid volumes contained in the tubing, valves, gauges and fittings of Figure 3 were present on either side of the sample. The pressure transducer was connected to the upper volume line. During a typical permeability experiment, a particular confining pressure (P_c) was introduced into the vessel. The rock sample was then subjected to a pore pressure and allowed to establish an equilibrium pore pressure (P_p). The average effective confining pressure ($\bar{P}$) that the sample experienced was the difference of the confining pressure and the equilibrium pore pressure ($\bar{P} = P_c - P_p$).

After equilibrium was obtained, the lower pore pressure P_2 was dropped to one-half the desired differential pressure (ΔP), where ΔP = differential pressure across the faces ($= P_1 - P_2$), while the upper pore pressure P_1 was raised an equivalent amount. The pressures change with time and approach a constant value, which is close to the equilibrium pore pressure (P_p).

Figure 5. Schematic representation of experimental apparatus, where

V_1 = volume of tubing and gauges of upper line

V_2 = volume of tubing and gauges of lower line

P_c = confining pressure

P_1 = upper line pressure

P_2 = lower line pressure

P_p = equilibrium pore pressure.

The permeability of the sample was obtained from the decay characteristics of the upper pore pressure with time. The compressibility (β) of the pore fluid is written as

$$\beta = 1/V_S (dV/dP) \quad , \quad (2)$$

where V_S is the volume of the upper line and dV/dP is the change in upper line fluid volume with pressure. The flow rate Q can be written as

$$Q = dV/dt = \beta \cdot V_S \cdot (dP/dt) \quad . \quad (3)$$

Equation (3) is an expression for the flow rate involving the compressibility of the fluid, the system volume and the change in pressure with time, all of which are easily measurable. Over a short time interval, the decay of the upper pressure is nearly linear and dP/dt can be approximated by a linear slope. The flow rate determined from Equation (3) can then be used in Darcy's equation to calculate the permeability k ,

$$k = \frac{QL\mu}{A\Delta P} \quad , \quad (4)$$

where k is the permeability in Darcy's (10^{-8} cm^2), μ is the viscosity in centipoise, L is the length of the sample in cm, A is the cross-sectional area of the face of the sample in cm^2 and ΔP is the differential pressure across the faces in atmospheres (dynes/cm^2).

The differential pressure was chosen so that it was less than a few percent of the confining pressure. Brace and others (1968b) have shown that permeability is markedly dependent on confining pressure, and therefore $\bar{P}$ would not be approximately equal at the different ends of

the sample if the differential pressure were large. In addition, the fluid properties vary with pressure, and if P were large, these properties would not remain constant along the length of the sample.

Transient pressure effects due to porosity and adiabatic changes in temperature were allowed to occur before any measurements were taken. These changes usually took place within 30 seconds to a minute after the start of the experiment.

The usefulness of this experimental technique becomes apparent when the low permeability of these crystalline rocks is taken into account. Using traditional flow measurement methods, low permeabilities are difficult to measure precisely. In addition, traditional techniques and the method outlined by Brace and others (1968b) require 10 minutes or more to make a measurement. This method requires only a short time interval (60-180 seconds) to run an experiment. A sample permeability measurement is shown in Appendix 1, along with all necessary calculations.

Flow-channel Tracing

Two approaches to flow-channel tracing were conceived in the initial stages of this work: 1) flow-channel tracing using radioactive species, which, after being deposited in the channels, could be delineated using autoradiography; and 2) decoration of channels with a metal species, which could be detected using the non-dispersive x-ray unit on the scanning electron microscope. Unfortunately the second method proved unsuccessful because the concentration of metal introduced into the channels as an aqueous species could not be detected with the x-ray analyzer. The first method however has worked with reasonable success.

Samples used in the channel-tracing experiment were GT-2-8580' and 9522' cores that had been previously subjected to permeability experiments. After removal of the core from the permeameter, 1.5 ml of water containing 0.5 millicurie of radioactivity was placed on the upper surface of the core. Following application, the fluid was allowed to dry under a heat lamp for several hours. The core was then jacketed and replaced in the pressure vessel. The sample was subjected to a 100 bar effective confining pressure. Water under pressure was introduced into the upper pore-pressure line; this resulted in a flow gradient along the length of the sample. During this period of flow, the water-soluble radioactive species was flushed through the sample along the various flow channels. Using the estimated porosity and measured flow rate, the amount of fluid flow necessary to transmit two pore volumes through the sample was calculated. After these two pore volumes of fluid had moved through the sample, the upper pore and confining pressures were released. The sample was then removed and allowed to dry.

Cores were then cut using a low-speed diamond wafering saw and mineral oil as the cutting fluid. Samples were roughly polished with 320 grit in mineral oil to obtain a flat surface. Following polishing the samples were autoradiographed for exposure times of approximately two weeks on fine-grain x-ray film. The film was vacuum positioned against the rock slabs for maximum resolution.

After autoradiography the samples were polished with fine silicon carbide and aluminum oxide abrasives (400, 600, 5 μ m, 1 μ m, 0.25 μ m) in successive steps. Samples were then vacuum coated with a thin ($\sim$ 200Å)

thick layer of graphite. Graphite provided a conductive surface layer, which is necessary for viewing in the SEM. SEM photomosaics were made on several of the rock sections that had given good autoradiographs.

PRESSURE EFFECTS

The effect of pressure on the physical characteristics of a rock is important when considering the behavior of the rock at depth. As confining pressure increases (hydrostatic or non-hydrostatic), the pore volume or porosity decreases proportionally. This decrease is accomplished by closing existing fractures, pores and flow channels. Diffusion and permeation, along with elastic behavior of the rock at pressure, are determined by the distribution, shape and size of these cracks and pores (Adams and Williamson, 1923; Birch, 1961; Walsh and Brace, 1966; Walsh and Decker, 1966).

The pressure-permeability relationship is important to the LASL hot-dry-rock geothermal program in that permeability controls loss of the introduced aqueous working fluid from the system into the essentially dry surrounding rock. If the downhole fluid is at a pressure significantly higher than the in-situ pore pressure then, by the law of effective stress, this increased pore pressure would decrease effective stress, resulting in increased permeabilities. The magnitude of water loss into the country rock could seriously effect the performance of the geothermal system.

This section deals with the effect of pressure on the permeability of core from geothermal drill hole GT-2. The first experiments were

run to determine permeability as a function of confining pressure and are similar to the original experiments at room temperature by Brace and others (1968b). The second experiment tests the law of effective stress using permeability measurements.

Data for the first experiment are listed in Table 2. This table shows confining pressure P_c , differential pressure ΔP , upper pressure P_u , and effective confining pressure $\bar{P}$ at initial time t_0 . All pressures are reported in bars. The uncertainty in P_c was less than 0.7 bar, and in P_u and ΔP less than 0.5 bar. The volume of the upper pore-pressure line was measured using the compressibility equation,

$$V_s = \frac{1}{\beta} \frac{dV}{dP} \quad (5)$$

The fluid used was water, as its compressibility is well known; the dV/dP term was measured by weighing the amount of water bled from the upper line over a measured pressure drop. The scatter in the volume measurements was less than 0.5 cc. Compressibility and viscosities of water were taken from the Handbook of Chemistry and Physics (1970 ed.). Permeability is measured in units of Darcy's, which equal approximately 10^{-8} cm^2 . The permeabilities listed in Table 2 are in units of nano-Darcy's, which equal 10^{-9} Darcy (10^{-17} cm^2).

The results of these experiments are shown in Figure 6, along with the values from Brace and others (1968b) on Westerly granite and those of Ohle (1956) on coarse-grained limestone. The small scatter present in the GT-2 measurements is probably related to several effects, such as inability to keep the system completely isothermal during a measurement and possibly a slightly non-hydrostatic application of

TABLE 2.
Permeability at 24° C as a function of confining pressure.

GT-2-8580 sample					GT-2-9522 sample				
Pressure (bars)				Permeability k (nd)	Pressure (bars)				Permeability k (nd)
P _c	P ₁	$\bar{P}$	ΔP		P _c	P ₁	$\bar{P}$	ΔP	
172	165.5	7.7	1.7	1930± 350	21.4	15.5	7.6	1.7	15500± 800
172	158.6	14.6	3.4	704± 82	31.7	15.5	17.9	1.7	3950± 430
172	151.7	21.5	3.4	339± 88	50.0	17.2	36.2	3.4	1430± 250
172	145.0	28.4	3.4	289± 43	67.2	17.2	53.4	3.4	1000± 90
172	138.0	35.3	3.4	238± 11	86.2	20.7	72.4	6.9	990± 50
172	120.6	51.7	3.4	169± 23	120.6	20.6	106.8	6.9	598± 48
172	103.4	68.9	3.4	155± 18	155.1	20.6	141.3	6.9	422± 50
207	68.9	137.8	6.9	63.7± 3.8	224.0	24.1	210.2	10.3	295± 7
275	68.9	207.0	10.3	39.5± 1.3	293.0	31.0	279.2	17.2	136± 7
344	68.9	275.8	13.8	22.6± 3.4	361.9	34.5	348.1	20.7	89± 6
414	68.9	344.7	17.2	12.4± 3.4

Figure 6. Permeability as a function of effective confining pressure.

stress. Another source of scatter might lie in the inability to read and reproduce pressure exactly. The degree of scatter is equal to or less than that obtained by Brace and others (1968b).

The curves from Brace and others (1968b) and Ohle (1951) show a pressure-permeability relationship not unlike that observed for the GT-2 samples. This similarity in slopes of the three sets of curves, particularly for granitic rocks, would suggest analogous flow-channel morphology and similar behavior of these channels to increasing pressure. The absolute values of the permeability, however, show marked differences. These differences may be influenced by many variables, particularly for crystalline rock. These variables include the crystallization history of the rock and subsequent pressure, temperature and fluid conditions. This problem will be discussed in a later section of this report.

In their permeability paper, Brace and others (1968b) suggested that at low pressures, long, low aspect ratio cavities (LARC's) and high aspect ratio cavities (HARC's) are open while at higher pressures only the more equant cavities (HARC's) are open. This theory would seem to fit experimental data from both measurements on GT-2 rocks and Westerly granite. In the very low effective confining pressure range, less than 50 bars, there is a distinct break in slope of the GT-2 measurements. Below this pressure the curve increases rapidly with decreasing pressure. This increase in slope may possibly be due to connected strings of LARC's.

Using differential strain measurements, Simmons and Cooper (1976) discovered that on most of the GT-2 samples measured, a noticeable peak occurred in the strain measurements similar to the break in slope observed in the permeability measurements. They related this peak to large, unoriented fractures with low closing pressures. These fractures, they postulated, resulted from damage incurred during drilling or sample preparation.

As a result of this work on the effect of temperature on the permeability of these rocks, I believe that this low pressure phenomena is a manifestation of thermal contraction effects induced during the nearly 200°C drop in temperature the GT-2 rocks have undergone upon removal of the surface. This effect will be discussed more thoroughly in the next section of this report.

The next experiment was run as a test of the law of effective stress. This experiment consisted of measuring the permeability of a sample at a constant confining pressure over a range of differential pressures. The lower pore pressure was held constant as the upper pore pressure was increased after each measurement. This experiment was run at two confining pressures, 207 and 345 bars.

Results of these measurements are listed in Table 3 and are shown on Figure 7 as permeability against differential pressure. As can be seen from this figure, the values remain nearly constant within the limits of error for the measurements. The permeability of a rock sample subjected to a differential pressure would be determined by the maximum effective stress, which in this case would equal $(P_c - P_1) = \bar{P}$. This is in agreement with the law of effective stress.

TABLE 3.

Permeability of GT-2-9522 as a function of differential pressure.

Pressure (bars)			Permeability k (nd)
P _c	P _u	ΔP	
206.9	41.4	6.9	3373 $\pm$ 105
206.9	48.3	13.8	3070 $\pm$ 85
206.9	68.9	34.5	3040 $\pm$ 116
206.9	103.4	68.9	2956 $\pm$ 80
206.9	137.9	103.4	2935 $\pm$ 105
206.9	172.5	137.9	3180 $\pm$ 149
344	41.4	6.9	509 $\pm$ 23
344	48.3	13.8	516 $\pm$ 23
344	68.9	34.5	457 $\pm$ 7
344	103.4	68.9	473 $\pm$ 18
344	137.9	103.4	527 $\pm$ 16
344	172.4	137.9	625 $\pm$ 13

Figure 7. Permeability as a function of differential pressure at constant confining pressure (P_c).

The absolute permeability of adjacent cores from any one depth may show measurable differences. The variation in permeability may be as high as two or three times when measured at room temperature. These differences are believed to be either due to; 1) small scale anisotropy in the sample cores, such as foliation and sealed or unsealed fractures or, 2) orientation of sample core relative to the drill core. No apparent foliation is observable in either core, however the GT-2-9522' cores often show well defined calcite filled fractures. These filled fractures may or may not have some effect on permeability of the cores.

Initially, permeability was measured on both parallel and perpendicular cut cores from the 8580' depth. These showed within the limits of accuracy at that time essentially no change in permeability. These experiments were not redone on GT-2-9522' cores which is unfortunate since the 9522' permeabilities show the greatest latitude. The reason for this is that during sample coring the orientation of the cores were not recorded.

TEMPERATURE EFFECTS

In order to obtain a realistic set of laboratory permeability measurements on GT-2 samples, these measurements were run at temperatures and pressures approximating in-situ conditions. Temperature logging near the bottom of the GT-2 hole indicates that the highest temperature is near 200°C. It would be desirable to have permeabilities at temperatures up to 300°C; however, in this temperature region the sealing and material-breakdown problems were considered too difficult to solve in this preliminary work.

Experiments testing the effect of temperature on permeability were of three types: 1) permeability was measured as a function of temperature at a constant confining pressure; 2) permeability against pressure was measured at several temperatures; and 3) time-dependent behavior was measured by a long term flow experiment. These experiments were run on GT-2-9522' and -8580' cores and, in some cases, Westerly granite.

The first experiments investigated the effect of increasing temperature on permeability of samples from GT-2-9522', -8580', and Westerly granite. A core was placed in the permeability apparatus and subjected to a constant confining and pore pressure. The vessel was then heated in the controlled furnace at a rate of approximately 30°C per hour. This heating rate was used to avoid fracturing due to uneven thermal gradients in the sample (Todd, 1973). Permeabilities were measured along the heating curve at approximately 10-15°C intervals. The vessel was then allowed to cool after the maximum

temperature had been reached, and permeabilities were also measured along the cooling portion of the curve for GT-2-9522' and Westerly samples.

The results of these experiments on GT-2 and Westerly samples are shown in Tables 4, 5 and 6. Viscosities of water at temperature have been taken from the Handbook of Chemistry and Physics (1970 ed.). Since these measurements took place while the sample was heating or cooling, a temperature-produced adiabatic pressure change had to be accounted for during the measurement. This was accomplished by setting the pore pressure at a constant value, then measuring the change in pore pressure due to thermal expansion of the pore fluid over a small time interval. A flow rate was then measured by introducing a pressure gradient along the sample. The temperature-induced pressure change was then either subtracted or added, depending on whether the core was heating or cooling respectively. This value gives the flow rate through the sample, and permeability could be calculated from this data. Data from Tables 4 and 5 are shown plotted on Figures 8, 9 and 10 as permeability against temperature. Confining pressure is lower for the GT-2-8580 run, but this variation is not considered to have significantly altered the behavior of permeability with temperature for 8580'.

As temperature is increased, the permeability of the GT-2-9522' sample drops nearly an order of magnitude (Fig. 8), reaching a minimum near 140°C. Upon cooling (Fig. 9), the permeabilities follow very nearly the same curve. The effect that causes the decrease is not permanent and is reversible in this case.

Results for GT-2-8580' are similar to that seen in the 9522' measurements, except that the curve drops off more rapidly with

TABLE 4.
Permeability as a function of temperature for sample GT-2-9522

$P_c=207$, $P_1=106$, $\Delta P=6.9$, and $\bar{P}=110$ Bars

Heating			Cooling		
Temperature (°C)	Viscosity (cp)	Permeability (nd)	Temperature (°C)	Viscosity (cp)	Permeability (nd)
31	.773	201	26	.876	236
37	.684	178	29	.808	230
48	.564	106	33	.753	203
59	.469	81	38	.678	160
64	.435	70	45	.590	134
73	.385	54	46	.584	150
88	.319	46	51	.536	142
95	.295	58	60	.468	119
131	.208	27	66	.428	87
147	.184	30	70	.403	80
153	.175	32	77	.367	74
162	.164	33	83	.341	58
168	.157	34	91	.309	39
175	.150	29	102	.277	39
180	.144	38	110	.253	41
192	.133	41	120	.232	30
195	.131	33	126	.218	31
200	.127	40	138	.197	20
			148	.183	26
			155	.173	24
			161	.165	24
			168	.157	24
			173	.152	40
			200	.127	40

Figure 8. Permeability as a function of temperature for GT-2-9522 (heating).

Figure 9. Permeability as a function of temperature for GT-2-9522 (cooling).

temperature, reaching a minimum near 130°C; with increasing temperature it goes through a level region and then starts to increase near 150°C (Fig. 10).

The results of the measurements on Westerly granite show strikingly different permeability behavior as a function of temperature (Fig. 11). As the temperature is increased the permeability shows a slight decrease up to 110°C, after which permeability increases exponentially with increasing temperature. Upon cooling the curve decreases slightly from the highest value, but permeability is permanently increased nearly four times the original value.

Both the decrease in the 9522' permeability values and the increase and subsequent hysteresis in the Westerly permeabilities can be explained by one mechanism, differential thermal expansion. This mechanism will be discussed in a later section of this report.

Permeability as a function of temperature over a range of pressures was next investigated. Permeabilities at several temperatures over a range of confining pressures were measured on a GT-2-9522 sample. The results of this experiment are shown on Table 7. These measurements were run with a slightly different procedure than the preceeding experiments. After the permeability was measured over a range of pressures at constant temperature, the temperature was increased and allowed to equilibrate. Permeabilities were then remeasured over the range of pressures and so forth. Data from Table 7 are plotted on Figure 12 as permeability against temperature.

TABLE 5.

Permeability as a function of temperature for sample GT-2-8580

 $P_C=140$, $P_1=106$, $\Delta P=6.9$, and $\bar{P}=35$ Bars

Temperature (°C)	Heating	
	Viscosity (cp)	Permeability (nd)
25	.892	534
30	.804	563
35	.717	512
39	.657	478
45	.595	345
50	.542	219
57	.483	109
62	.450	76
73	.388	56
82	.344	45
96	.294	44
104	.269	36
112	.247	33
127	.217	33
139	.196	44
147	.182	40
159	.167	40
168	.157	33
172	.153	36
177	.147	57
183	.141	74
188	.137	100
191	.134	109
198	.128	135
203	.124	143
214	.117	139
216	.115	172

Figure 10. Permeability as a function of temperature for GT-2-8580 (heating).

TABLE 6.

Permeability as a function of temperature for Westerly granite

 $P_c=207$, $P_1=117$, $\Delta P=9.6$, and $\bar{P}=110$ Bars

Heating			Cooling		
Temperature (°C)	Viscosity (cp)	Permeability (nd)	Temperature (°C)	Viscosity (cp)	Permeability (nd)
26	.89	45	25	.89	162
60	.466	39	54	.512	181
103	.28	42	73	.388	192
134	.199	66	83	.342	219
142	.190	73	95	.297	202
147	.183	87	104	.270	208
152	.171	94	124	.220	280
158	.168	119	127	.216	313
166	.159	164	142	.189	418
172	.152	163	154	.174	417
178	.146	236	166	.159	399
184	.140	248	178	.146	419
190	.135	264	192	.134	402
195	.131	328	202	.126	432
198	.128	326

Figure 11. Permeability as a function of temperature for Westerly granite.

The permeability of the 9522' samples at 103 bars (Fig. 12) behaves similarly to that of the constant confining pressure, heating experiment (Fig. 8) except that the values at higher temperatures are slightly increased in this experiment. This may be due to a different pressure-permeability relationship. There is also a noticeable rise in the values at the 200°C temperature for all pressures except at 345 bars. In the very low pressure region (35 bars), the permeability drops drastically with an initial increase in temperature (20-80°C). This is the low pressure region discussed earlier in this report. The large increase in permeability in the low pressure region may be due to the existence of thermal expansion channels and is probably not a result of damage due to drilling or sample preparation. At higher pressures this initial drop is usually not as large, since most of the temperature-induced fractures have low closing pressures, as was shown by Simmons and Cooper (1976).

The final temperature-related experiment involved long-term fluid flow through a sample and the effect of this flow on the permeability. This experiment involved measuring the permeability at 200°C and then subjecting the core to a constant pressure gradient for 120 hours, after which the permeability was again remeasured. Data from this experiment are tabulated in Table 8 and shown on Figure 13.

Permeability has been shown to be dependent on both pressure and temperature and may also be considered dependent on the presence of pore fluids and their movement through rock. Figure 13 shows

TABLE 7.

Permeability as a function of temperature over a range of pressures for GT-2-9522

Pressure (bars)				Permeability k (nd)			
P_c	P_1	ΔP	$\bar{P}$	at 24°C	at 83°C	at 140°C	at 200°C
85.4	51.7	3.5	34.5	3160± 127	225± 10	117± 8.6	124± 6.1
155	55.1	6.9	103.5	246± 35	101± 8.2	42± 5.2	50.6± 6.4
258	55.1	6.9	206.9	134± 36	25.5± 3.0	21.1± 2.2	28.8± 6.4
403	68.9	20.6	344.8	49.6± 3.8	13.0± 2.0	6.3± .75	5.6± 1.0

Figure 12. Permeability at four confining pressures as a function of temperature for GT-2-9522.

TABLE 8.

Permeability measured prior to and after long term flow experiment

Pressure (bars)				Permeability k (nd)	
P _c	P ₁	ΔP	P	before flow	after flow
42.0	36.2	1.7	6.9	699± 43	824± 110
52.4	36.2	1.7	17.2	279± 37	759± 20
70.7	37.9	3.5	36.2	242± 34	524± 29
87.9	37.9	3.5	53.4	185± 24	409± 14
105.1	37.9	3.5	70.7	154± 15	254± 8
141.3	41.4	6.9	106.9	108± 20	158± 16
175.8	41.4	6.9	141.3	69.5± 2.5	97.6± 7.4
246.5	44.8	10.3	212.0	31.8± 1.8	48± 3.5

Figure 13. Permeability as a function of effective confining pressure for GT-2-9522 sample prior to and after long-term flow experiment.

that after a significant amount of fluid flow, permeability is increased significantly over the entire pressure range. The dominant reaction between rock and water to increase permeability is thought to be the dissolution of quartz. Silica solubility in water at 200°C is near 230 ppm, while the distilled water entering the system has a concentration of much less than 1 ppm. Therefore if the fluid has enough time to come to equilibrium with the rock, then channels, particularly those in or around quartz grains, will probably be etched by the fluid. The fluid that had moved through the rock in this experiment was caught, but was lost after analyzing for uranium, so no analysis was carried out for major element concentrations. Uranium concentration in the fluid was found to be less than background levels (<0.2 ppb).

FLOW-CHANNEL TRACING RESULTS

The choice of radioactive isotopes used in the channel tracing experiments was determined by: 1) the penetration range of emitted particles in granite, 2) ease of use, and 3) half-life and availability. The three isotopes used in these experiments are listed in Table 9 along with the particle emitted and penetration distance of this particle into a 1 cm-thick section of material with the density of granite (2.67 g/cc). The penetration distance can be found from

$$d = \frac{R}{\rho g} \quad , \quad (6)$$

where d is the penetration distance in cm, R is the areal range of the emitted particle in mg/cm^2 , and ρg is the density of granitic material in mg/cm^3 .

TABLE 9.

Penetration distances of radioisotopes used in the flow channel experiments.

Isotope	Particle	Energy (MEV)	Range of particle (mg/cm ²)	Range in granite (μm)	half life
²² Na	β ⁺	.545	170	636	2.6 years
⁶³ Ni	β ⁻	.0659	6.6	24.7	100 years
³⁵ S	β ⁻	.167	30	112	87 days

The initial radioactive species used in the channel tracing experiments was ^{22}Na . ^{22}Na emits a fairly high energy positron with a penetration range of over 500 microns. Figure 14 shows an autoradiograph taken on a slab of 8580' injected with ^{22}Na solution cut parallel to the direction of flow. Also shown in Figure 14 is an optical photograph made of the same sample. The dark areas on this and all autoradiographs in this report represent regions of high radioactivity. The large fracture in this sample was formed following cutting when the slab was accidentally dropped.

Even though this autoradiograph exhibits very poor resolution, a distinction can be made between light and dark areas, and a definite tendency of these areas to follow traceable patterns can be ascertained. There also appears to be a clear relationship between the most radioactive regions and the dark areas on the optical photograph, mainly biotite and magnetite.

Following this initial experiment it was decided that another radioactive isotope would be tried, since ^{22}Na showed such poor resolution. ^{63}Ni was chosen for this next experiment because of the very low energy beta-particle ^{63}Ni emits during radioactive decay (66 Kev end point energy). This particle has a penetration range of about 25 microns in granite. Figure 15 shown an autoradiograph of a section of 8580' that was injected with ^{63}Ni . This section was cut perpendicular to the direction of fluid flow. Also shown on Figure 15 is an SEM photomosaic taken from the same section of rock.

Figure 14. Autoradiograph (A) and optical photograph (B) of a slab of GT-2-8580' injected with ^{22}Na .

Fig. 23a

0 ————— 1 cm

Figure 15. Autoradiograph (A) and SEM photograph (B) of a section of GT-2-8580' injected with ^{63}Ni .

Excellent spatial resolution was obtained using ^{63}Ni . Distinct channel outlines are apparent when compared to the autoradiograph from ^{22}Na . Note the large biotite grain in the center of the photomosaic and the same respective area on the autoradiograph. This biotite grain (or grains) is transected by numerous dark, channel-like features, as shown in the autoradiograph.

To determine whether deposition of the radioactive species was physically rather than chemically controlled, a species was then chosen that could not easily substitute into any of the mineral structures. Both nickel and especially sodium may be preferentially deposited in regions of the granite where minor amounts of cation exchange are possible, although this is highly unlikely at room temperatures. ^{35}S was chosen for this experiment since, in the form used, sulfur ion would be in the sulfate ion and because ^{35}S emits a low energy beta-particle during decay. The autoradiograph and photomosaic of a section of 9522' injected with ^{35}S are shown on Figure 16. This sample was also cut perpendicular to the direction of flow.

Channel distribution between the ^{35}S and ^{63}Ni are very similar in both autoradiographs; however the ^{63}Ni autoradiograph tends to emphasize the channels in the biotite with respect to channels in or around the other grains. The ^{35}S autoradiograph is slightly fuzzier because of the higher energy particle. The large white spot on the right side of the autoradiograph is a region where the sample was ion-milled for several hours to test whether the radioactivity could be smeared over the surface during cutting of specimen. Only a small amount of smeared radioactivity was observed.

Fig. 24a

Figure 16. Autoradiograph (A) and SEM photograph (B) of a section of GT-2-9522' injected with ^{35}S .

DISCUSSION

Relating laboratory-measured permeabilities to in-situ measurements presents numerous problems including: 1) inability to determine the surface area of the downhole hydrofracture, 2) presence of porous zones and fractures not found in small experimental cores, and 3) large reservoir compressibility due to fractures with respect to fluid and rock compressibility. Downhole permeability measurements are most commonly given in terms of $(A \sqrt{k})$, where A is the area of the hydrofracture, and k is the permeability. One major problem in trying to relate the two is that the triaxial stress conditions present downhole are nearly unreproducible in the laboratory.

However large these problems, the laboratory measurements still have some practical application with respect to in-situ values. Fisher (1977) has developed a model using both laboratory and in-situ measurements which explains some of the observed discrepancies. Since fractures or porous zones are unlikely to be found in samples of laboratory size, the laboratory measured permeabilities are taken as minimum values. Fisher's model postulates extremely competent rock (similar to laboratory samples) with several widely spaced fractures or fracture zones that would give higher permeability values.

Originally the downhole permeability was thought to be very dependent on hydrostatic pressure; however, after careful reexamination of the pressure decay data, the flow rate appears to be only very slightly pressure dependent. Fisher's model would account for this observation by assuming a much higher intermediate stress (σ_2) than

was originally thought to occur. This stress field would then allow channel expansion only in a direction perpendicular to the minimum stress (σ_3). This is shown schematically in Figure 17. Fluid movement would be restricted to those channels which are parallel to the hydrofracture. Increasing the pore pressure above σ_3 would only increase the size of the hydrofracture. According to this model then, fluid flow would be almost solely determined by the existence, size and location of porous zones and fractures.

Results of permeability-temperature experiments may have important implications for the behavior with temperature of the downhole geothermal reservoir. The original assumption was that as heat is removed from the reservoir rock, gradual decreases in temperature would result in differential thermal contraction, yielding increased reservoir permeability. Increased reservoir permeability would have two contrasting effects, loss of circulating fluid into rock and increased surface area. Loss of fluid may or may not have a negative effect on the system, since a large portion of this fluid is recoverable. The second effect would be positive, since increased surface area would mean better heat extraction rates and would give a more efficient geothermal system.

The data from Figure 8 seems to indicate that as heat is removed and the temperature of the reservoir rock at a depth of 9522' decreases, permeability would not start to increase until the temperature was lower than 140°C. In fact decreasing the temperature from 200-150°C would result in a decreased reservoir rock permeability. This effect is limited however, to the competent reservoir rock and does not take into account effects in porous or fracture zones.

Figure 17. Schematic representation of downhole GT-2 hydrofracture and associated principal stress orientations.

Explanation of the behavior of permeability as a function of temperature requires a mechanism that acts both in a reversible manner, as in the 9522' values, and can cause permanent changes, as observed in the Westerly measurements (Fig. 11). Several mechanisms explain some of the phenomena observed but fail to explain others. Originally some type of clay formation was thought to cause the initial decrease in permeability of the 9522' and 8580' samples. Summers and others (1975) had observed permeability decreases with time on granite samples; these changes were thought to be the result of clogging due to clay formation along flow channels. Their experiments were run at high temperatures (up to 400°C) and high axial loads (~4KB). Balagna and Charles (1975) observed changes in permeability with time on granite samples subjected to alternating periods of fluid flow and rest. During periods of static pore pressure, permeability is observed to decrease slightly. They relate these decreases to clay formation along channels.

Pressure solution was also suggested as a mechanism which could produce the initial permeability decreases observed. This phenomena is caused by increased mineral solubility at high stress points and acts to reduce stress by redepositing material from high stress to lower stress regions. This effect has been observed by Sprunt and Nur (1976) to decrease porosity in axially loaded sandstones.

Clay formation would not, however, provide a good mechanism for the phenomena observed. Since the kinetics of the clay formation would be relatively slow, reversability, such as seen in the 9522' measurements, would be improbable. Pressure solution effects are also

slow and require fairly large inhomogeneous stress conditions.

Differential stresses need not be large if intergranular contact is limited, such as in loosely packed sandstones; however the grains in these rocks fit closely together.

Mineral dissolution, particularly around quartz grains, might explain permeability increase; however this is a slow process, as is shown by the results of Figure 13. This process also requires a substantial amount of fluid flow through the sample. Only a small amount of fluid (less than $.01 \text{ mm}^3$) was passed through the sample during an experiment.

Differential thermal expansion (DTE) would best explain the phenomena observed. Although DTE is believed to have a dominant influence, the other processes such as mineral dissolution may play a smaller role.

Differential thermal expansion is a phenomenon caused by the anisotropy and inhomogeneity of the coefficients of thermal expansion of different mineral grains. Figure 18 shows data on volumetric thermal expansion plotted as a function of temperature for quartz, plagioclase (Ab_{56}) and orthoclase. In addition, quartz has a relatively large thermal expansion, especially near the β transition at 573°C , although this point is not shown on Figure 18. However, in the $0\text{-}200^\circ\text{C}$ region used in these permeability experiments, all expansion-temperature slopes are nearly linear. Quartz also shows a very distinct anisotropy in expansion depending on crystallographic orientation. Figure 19 compares the volumetric expansion of quartz grains with thermal expansion in directions perpendicular and parallel to the crystallographic c-axis.

Figure 18. Volumetric thermal expansion as a function of temperature for several different minerals (after Skinner,1966).

Figure 19. Thermal expansion of quartz as a function of temperature
(after Skinner,1966).

A rock containing numerous grains, all of which are unoriented with respect to one another, will undergo a certain amount of expansion with increasing temperature. Minerals such as quartz, due to their large coefficients of expansion, will increase volumetrically more rapidly than other grains. These nonlinear and disproportionate increases in volume will close channels and fractures between grains. As temperature is further increased, stresses will tend to build up in and around these grains until brittle fracture takes place. The direction of maximum stress within a grain will coincide with the direction of maximum thermal extension; hence, for example, the largest principal stress will built up in a direction perpendicular to the c-axis for quartz grains.

The behavior of permeability of both Westerly and GT-2 samples as a function of temperature (Figs. 8-11) can be explained rather simply by differential thermal expansion. Both GT-2 and Westerly rocks contain rather large amounts of quartz (Table 1), hence the thermal expansion effects could be quite noticeable.

Permeability as a function of temperature for the Westerly sample (Fig. 11) is postulated to show the effects of differential thermal expansion. From 20°C to nearly 100°C, the curve shows only a slight decrease in permeability. This effect may be related to nearly complete closure of low aspect ratio cavities (LARC's) that separate quartz from other grains. Since quartz increases in volume much more rapidly with temperature than the other minerals in the rock, it would be expected that the LARC's separating grains such as plagioclase-orthoclase would tend to close much more slowly with increasing

temperature. The presence of these unclosed channels may provide an explanation for why the permeability does not drop off to a much lower value than is recorded. High aspect ratio cavities (HARC's) will probably remain relatively unaffected by the thermal expansion since most of the HARC's present in the rock are intragranular; also HARC's require a much higher closure pressure (Brace and others, 1968; Walsh, 1966).

Continued heating from 100-200°C causes an exponential increase in permeability of the Westerly sample. This temperature region may coincide with brittle fracture within grains or around grain boundaries. These fractures are probably well connected, since permeability is increased dramatically. Upon cooling, permeability decreases from nearly 400 nDarcys at 200°C to 200 nDarcys at 100°C. This decrease can be explained by assuming that most fracturing is intragranular and that with subsequent thermal contraction, these channels within grains are partially closed. A small decrease in permeability is observed from 100°C to 20°C; this decrease can be attributed to incomplete closure of both preexisting intergranular channels and temperature-induced intragranular channels. A permanent change in the permeability of the 21°C value of nearly four times occurred after heating.

If DTE is a valid explanation for these permeability phenomena, then one ought to be able to observe fractures within quartz grains. The Westerly sample which had been used in the permeability experiments was polished and examined with the SEM. Figure 20 shows a photograph taken with the SEM of an area of the Westerly and an enlargement of

TABLE 10.

Symbols used on SEM photographs

Q ---	Quartz
P ---	Plagioclase
K ---	Potassium feldspar
B ---	Biotite
M ---	Magnetite
S ---	Sphene
A ---	Apatite
C ---	Calcite
Py --	Pyrite

A

B

Figure 20. SEM photograph of grain boundary relationships in heated Westerly granite (A); B is a closeup of a grain boundary indicated by the box in A.

one of the quartz grains. Fractures are almost completely restricted to the quartz grain and grain boundaries. Figure 21 shows a similar surface area of an unheated Westerly sample. Note the complete lack of fractures and identifiable grain boundaries in this sample. The symbols used in these photos are listed in Table 10. Some of the fractures observed in Figure 20 show a crude alignment and may be roughly oriented with respect to the c-axis of quartz. Fabric relations were not determined in this study.

Since most of the fracture channels are present within quartz grains, and water was used in these experiments, quartz solubility and its effects on permeability may have some effect. The high magnification photograph of the fractures in the quartz grain show effects of silica dissolution. Note the large size and rounded appearance of the channels (Fig. 20). This dissolution probably helped to increase permeability.

Explanation of the permeability behavior observed in the GT-2-8580' and 9522' samples is more complicated. Both samples undergo permeability decreases, while only the 8580' measurements show a significant later increase. Assuming that the rocks are at their lowest permeability under in-situ conditions, then removal to the surface will cause thermal contraction within the rock. This thermal contraction will be dependent on the temperature difference between downhole and surface conditions and the composition and expansion coefficients of the various minerals.

The effect of differential expansion due to unloading must also be taken into account in this analysis. This expansion will accompany

A

B

Figure 21. SEM photograph of quartz and other grains in unheated Westerly granite sample (A); B is a closeup of the grain boundary indicated by the box in A.

removal of the sample from depth to the surface. This effect would work in an opposite sense to the thermal contraction. Since mineral compressibility commonly varies with crystallographic orientation, this effect would cause higher stress concentration in certain regions. The magnitude of unloading expansion can be calculated using the mineral compressibilities listed in Birch (1966). Volumetric expansion of quartz due to unloading from a certain depth was compared to volumetric thermal contraction of quartz from the temperature measured at that depth in the GT-2 hole. The pressure and temperature difference were calculated from the 9522' depth of the GT-2 hole, where $T = 190^{\circ}\text{C}$ and $P = 775$ bars. P was taken as the lithostatic pressure.

The amount of volumetric decrease in a quartz grain for a temperature drop of 170°C is 0.70%, while the volumetric increase over a pressure drop of 775 bars is close to 0.18%. The thermal contraction in the GT-2 rocks is thus by far the dominant effect. However, in areas where the geothermal gradient is not as high, the pressure-induced expansion could be more significant.

Upon heating under pressure in the laboratory, differential thermal expansion will again tend to close those fractures that had been opened during contraction. The permeability will decrease proportionately with expansion. However, the permeability will never be as low as the in-situ values, since permanent damage has been induced during contraction.

If the rock had come to equilibrium with respect to permeability at a particular temperature, then permeability would be lowest at this temperature. Permeability against temperature curves for the

GT-2 rocks should then show minimums at temperatures to which the rocks have equilibrated. This theory was first suggested by Dr. R. C. Fletcher (personal communication, 1976). Curves for 8580' and 9522' samples show the initial decrease predicted by this theory.

A permeability minimum is observed in the 8580' measurements at approximately 130°C (Fig. 10). A higher temperature increase in permeability, similar to that observed in the Westerly measurements, occurs at about 130°C. This effect may be caused by the same expansion-fracturing seen in the Westerly sample.

Permeability of the 9522' sample drops to a minimum near 140-150°C, and a very slight rise is observed from 150-200°C (Fig. 8). Upon cooling, no hysteresis is observed. It is probable that any damage incurred during heating past the minimum may be so slight as to have negligible effect at lower temperature.

Thermally induced fractures and channels in the GT-2 rocks, whether by contraction or expansion, should be easily observable under the SEM. These have been identified by the flow channel technique and are thoroughly characterized in a later section of this discussion.

If the flow channels at the permeability minimum occupy very nearly their in-situ shape and dimensions, then a simple calculation of post-expansion channel size can be made from the data on Figures 8 and 10. Original contraction effects can be calculated using coefficients of thermal expansion, average grain size and the difference between the permeability minimum temperature and surface temperature. The maximum contraction would occur between quartz-quartz grain boundaries.

However, these are probably less common than plagioclase-quartz and plagioclase-orthoclase boundaries, so the quartz-plagioclase and plagioclase-orthoclase boundaries were simulated.

Thermal expansions are listed in Skinner (1966) as percentages, so lengths of crystals had to be used. Two and one-half mm was chosen as an average crystal length. The differential temperature (ΔT) was the difference in temperatures where the permeability was a minimum (Fig. 8) and surface temperature. ΔT varied for each rock and is shown on Figure 22 as the temperature between points A and B for the 8580' sample and points C and D for the 9522' sample. The change in length of a crystal (Δ) in cm, can be found using

$$\frac{d}{100} \times L \times \Delta T = \Delta \quad , \quad (10)$$

where d is the thermal expansion in percent per $^{\circ}\text{C}$, L is half the length of an average crystal in cm, and ΔT is the differential temperature in $^{\circ}\text{C}$. Delta can be found for both quartz and plagioclase crystals, and the sum of the two Δ 's is equal to the channel width at 21°C , minus the original width of the channel at in-situ temperatures. The quartz-plagioclase and plagioclase-orthoclase channel dimension values at 21°C for both 8580' and 9522' samples have been calculated and are listed in Table 11.

INTERGRANULAR FLOW

The autoradiographs and photographs of Figures 14 and 16 seem to indicate that most of the fluid flow through the GT-2 rocks is

Figure 22. Permeability-temperature plots showing location of points used for channel size calculations.

TABLE 11.

Calculated channel expansions for several grain boundaries.

Mineral pair	slope expansion (%/°C)	GT-2-8580'	GT-2-9522'
		T=130-21°C (μ m)	T=150-21°C (μ m)
Orthoclase	.00091	3.45	4.07
Quartz	.0016		
Orthoclase	.00091	1.66	1.95
Plagioclase	.0003		
Plagioclase	.0003	2.61	3.08
Quartz	.0016		
Quartz	.0016	4.4	5.2
Quartz	.0016		

along intergranular channels. Several of the corresponding areas in the autoradiographs and SEM photographs were examined in greater detail. A higher magnification photograph of the 8580' slab along with a blowup of the ^{63}Ni autoradiograph are shown in Figure 23.

Differences in appearance between the photomosaic of Figure 15 and the enlargement of Figure 23 are due to differences in surface preparation. When the photomosaic was originally photographed, the surface had been polished with 900 grit. The slab was subsequently polished with 0.25 μm compound. The different mineral grains present in the photo have been identified using the x-ray analyzer on the SEM and are shown on the transparent overlay, along with the scale.

The area of the SEM photo has been located on the overlay of the autoradiograph enlargement of Figure 23. Note how the outlines of the two large quartz grains on the right side of the photo stand out in the autoradiograph. The smaller plagioclase crystal is clearly discernable between the two large quartz grains. Along the boundary between the biotite crystal and the quartz grain, the autoradiograph shows a large distinctive dark band. These dark areas, which in most cases correspond to intergranular channels, are much wider than most of the intergranular lines on the autoradiograph, although their appearance does not seem to be restricted to any particular grain. The origin of these bands is unknown, although they may be related to the thin-section cutting process.

The large biotite grain in the center of the SEM photograph has been outlined on the autoradiograph transparent overlay. Numerous

A

B

Figure 23. ^{63}Ni autoradiograph enlargement (A) and SEM enlargement (B) of a slab of GT-2-8580'.

channels can be seen on the radiograph that seem to bisect this large grain. After close inspection of the SEM photograph, one can observe these channels in the biotite grain as well. These channels outline distinctive breaks in the morphology of the biotite grain and probably represent individual grains in one large biotite mass. Magnetite grains are commonly in close proximity to the biotite grains; an example of this can be seen in the lower left portion of the SEM photograph.

The well-developed fractures between the orthoclase, plagioclase and quartz grains in the lower right portion of the photograph are not well documented on the autoradiograph. This phenomena of lack of radioactivity in regions away from biotite grains is seen throughout the sample. Perhaps the sheet structure of biotite provides surfaces for increased retention of the aqueous radioactive species. Whatever the reason, biotite almost always shows darker channels with better resolution.

An area on the 9522' sample injected with ^{35}S was investigated in a manner similar to that for ^{63}Ni . Figure 24 shows an enlargement of the ^{35}S autoradiograph and corresponding SEM photograph. The large quartz grain present in the center of the photograph has been located on the autoradiograph.

Some correlation of the features observed in the SEM photograph and autoradiograph can be made; however, the resolution of the ^{35}S autoradiograph is not good. The large biotite grain present on the left side of the SEM photograph shows no apparent channels, yet the

A

B

Figure 24. ^{35}S autoradiograph enlargement (A) and SEM photograph (B) of a slab of GT-2-9522' .

autoradiograph indicates several areas of activity. These radioactive regions may possibly have been smeared during sectioning and caught in holes which are often present in the biotite following cutting, or there may be channels here which are too small to observe at this magnification. Several of the channels on the left side of the SEM photograph between orthoclase and quartz can be easily discerned in the autoradiograph.

The channel tracing technique has proved most valuable in delineating the spatial relationships between channels and grain boundaries. Most of the flow through the GT-2 rocks appears from the autoradiograph to be primarily along well-developed intergranular channels. Some of the larger biotite grains were at first thought to contain many intragranular channels. Upon further investigation these channels appear to separate distinct biotite grains within larger clumps of biotite. No evidence was seen for appreciable amounts of flow through the numerous HARC's in many of the grains. These HARC's may possibly represent unconnected pores and fluid inclusions, which would not receive any radioactivity.

Resolution is the largest disadvantage of this technique. Several things might be done to increase resolution, particularly use of a finer grain x-ray film. This would mean that exposure times and possibly the amount of radioactivity introduced into the rock would have to be increased significantly. Finer polishing would also increase resolution. Future uses of this technique might include the determination of fracture connectivity in experimentally fractured rock and paths of flow through various types of rocks in conjunction

with some of the work being done on radioactive waste disposal in crystalline and sedimentary rocks.

THERMALLY INDUCED FEATURES

Fracture and channel-like features were observed in both unheated and heated GT-2 and Westerly samples. An attempt was made to delineate specific thermally induced features. To this end intergranular channels between coexisting quartz and orthoclase and in some cases plagioclase grains were studied in more detail.

Intergranular channels between grains in unheated Westerly granite were examined in detail and compared to the results of Sprunt and Brace (1974). Figure 21 shows a typical plagioclase-quartz boundary in unheated Westerly granite. Compare this photo to Figure 20 showing a quartz-orthoclase boundary from a Westerly sample heated to 200°C. Damage involving the breakage of the regions separating the individual LARC's has occurred after heating. Upon heating, intergranular cracks become much longer and show distinct brittle-fracture characteristics, such as sharp ends and straight fracture surfaces. Sprunt and Brace (1974) heated a sample of Westerly granite and noticed similar changes in morphology. Prior to heating, the intergranular channels in Westerly granite consisted of unconnected strings of LARC's showing delicate bridge-like features and commonly having rounded corners (Sprunt and Brace, 1974; Brace and others, 1972).

The nature of channels between grains in the GT-2-8580' and -9522' samples were observed in detail with the SEM. The boundary in an unheated 8580' sample between quartz and orthoclase grains,

along with a higher magnification photograph of the channel, are shown on Figure 25. Note the similarity of this channel to those observed in the unheated Westerly sample (Figure 21). Channels in this rock are typically strings of partially connected or unconnected LARC's, which in certain areas, especially around quartz grains, show some traces of brittle fracture.

Following heating, the cracks and channels in the 8580' sample take on a much different geometry. Figure 26 shows a region in the heated 8580' sample between several different grains. Notice that the intergranular channels now consist of both well-defined fracture zones and, in some cases, large single fractures. The right side of this photograph shows one of these zones between the biotite-quartz-plagioclase region. Another zone can be seen on the left side of the photograph between coexisting quartz grains. An enlargement of this region is shown in Figure 26B.

Figure 26B shows some of the details of these fracture zones. The surfaces in the fracture are rough, possibly indicating the effects of mineral dissolution and redeposition. The small individual fractures have widths on the order of $1-4\mu\text{m}$ while the entire fracture zone has a width of approximately $40\mu\text{m}$. Figure 26A shows an interesting feature: the singular fracture on the lower left seems to have a smaller en-echelon fracture directly below it. This fracture may be preliminary to a fracture zone forming along this channel. Changes in the fracture and channel morphology of the 8580' sample after heating apparently include complete destruction of any bridge-like features

A

B

Figure 25. SEM photograph of grain boundary relationships in unheated GT-2-8580' sample (A); B is an enlargement of the region indicated by the box in A.

A

B

Figure 26. SEM photograph of grain boundary relationships in heated GT-2-8580' sample (A); B is an enlargement of the region indicated by the box in A.

between LARC's and the development of large fractures and fracture zones between grains.

Channels and fractures in the 9522' sample show a characteristically similar morphology both prior to and after heating. These features are different than those of the 8580' sample. Prior to heating, channels in the 9522' are well-defined, long LARC's often suggesting features related to a brittle origin. Figure 27 shows the boundaries between quartz and orthoclase grains. A well-defined fracture is present between the orthoclase and quartz grain running up the center of the upper photograph (Figure 27A). A higher magnification photograph of the fracture is also shown (Figure 27B). This fracture appears to bifurcate in the upper portion of the enlargement. These fractures appear to have long straight fracture surfaces. The width of the small crack-like feature is less than $2\mu\text{m}$.

Figure 28 shows a similar area from the 9522' sample heated to 200°C . Morphology of the channels after heating are essentially the same. The effect of heating on the 9522' rock is apparently so slight that it is impossible to determine visually.

Visual observation seems to provide support for the DTE theory. Fracturing and changes in channel morphology due to heating can be seen in both Westerly granite and 8580' samples, although they are absent from the 9522' samples. The type of fracturing in the Westerly granite and 8580' samples may be dependent on grain shape, particularly in quartz, and the amount of heating past the permeability minimum temperature. Since fracturing did not take place in the

A

B

Figure 27. SEM photograph of grain boundary relationships in unheated GT-2-9522' sample (A); B is an enlargement of the region indicated by the box in A.

A

B

Figure 28. SEM photograph of grain boundary relationships in heated GT-2-9522' sample (A); B is an enlargement of the region indicated by the box in A.

Westerly sample until about 100°C and not until approximately 150°C in the 8580' sample, then the degree of overstepping of the permeability minimum temperature may have played a significant role. Grain shape would effect the type of fracturing taking place, since rounded or equidimensional grains could not break off edges of corners easily. In this case, stresses would tend to built up within the grain. Grains with an interlocking fabric or protrusions might break more easily during thermal expansion, thus reducing stress within the grain. Westerly granite grains seem to have more of a rounded appearance compared to the 8580' and tend to show more intragranular fracturing.

Since the permeability minimum temperature for the 9522' sample is slightly higher than the 8580', a smaller increase in permeability due to channel damage occurs from heating past the minimum temperature. If the 9522' sample was heated to around 240°C, then thermal expansion damage might be very similar to the 8580', which had been heated to 200°C.

The behavior of permeability with temperature for GT-2 crystalline rocks is due to several factors. However, the depth of the rock and the in-situ temperature at 9522 feet in the GT-2 hole is about 190°C, while at 8580 feet the temperature is 180°C. The temperature indicated by the permeability minimum for the 9522' sample is near 140°C, while the permeability minimum for the 8580' sample is nearer to 130°C. This paper has shown that the lowest permeability coincides with the greatest depth and highest temperature at equilibrium with their respective environments, since the lowest permeabilities coincide with temperatures 50-60°C lower than the in-situ temperatures. Since the kinetics governing the development of in-situ permeabilities are

probably very slow at these temperatures, it is probable that equilibrium in the superimposed temperature and pressure regime has not occurred. Therefore samples 8580' and 9522' may contain a remnant permeability reflecting a pre-existing pressure-temperature environment.

The changes in the geologic environment must be considered to understand completely what changes might have occurred in the temperature and pressure history of the GT-2 samples and the Westerly granite. For example, in crystalline rocks such as the Westerly granite, which have been near or at the earth's surface for fairly long periods of time, crystallization history, fluid content and depth of crystallization will all play a role in determining permeability (Brace, 1972).

Brace and others (1972) and Sprunt and Brace (1974) have observed cracks and microcavities in Westerly granite and conclude that most of these features are associated with the early history of the rock. They believe that none of the cavities are related to brittle processes such as thermal stress or relief of pressure, although they indicate that many of the fine intergranular cracks show bridging and rounded corners believed to be formed at a later time than the original crack.

These results have led the authors to postulate that laboratory measurements on crystalline rocks such as Westerly granite are representative of values at higher temperatures and pressures. This point may be valid for elevated pressure; however, temperature effects could introduce serious errors into the measured physical parameters, since even at low temperatures fracturing within and around mineral grains would have an important effect on bulk rock properties that are dependent on fractures, pores and stress within the rock.

Another problem is that anisotropic stresses developed within mineral grains might seriously affect stress-related measurements. Even rocks depleted in quartz, such as diabase, show strong thermal expansion effects, as was shown by Simmons and Cooper (1977).

The recent geologic history of the Nacimiento-Jemez region was extremely important in determining the in-situ permeability of the 8580' and 9522' samples. The uplift of a large block of Precambrian and overlying Paleozoic and Mesozoic sediments was initiated in late Cretaceous time during the Laramide orogeny. Uplifting was initiated along the Nacimiento fault, a high-angle reverse fault which may have undergone up to twenty thousand feet of vertical separation (Woodward and others, 1972). If the whole block was crudely tilted eastward, then the area around the Fenton Hill region may have also undergone a rather substantial amount of uplift. Since most of the uplift occurred about 50 to 60 million years ago, the pre-uplift permeability would have probably been significantly altered by the new physical environment.

In the period from 50 m.y. to about 10 m.y.b.p., these rocks probably equilibrated with the geothermal gradient of the area at that time. This gradient may have been close to the maximum estimated for the Rio Grande rift, approximately $30^{\circ}\text{C}/\text{km}$ (Bridwell, 1976). At this gradient the 8580' and 9522' rocks would have been at approximately 95°C and 105°C respectively.

An increased geothermal gradient associated with the onset of volcanic activity in the Jemez area probably increased the in-situ temperatures of the 8580' and 9522' rocks quite rapidly. Most of this heat was probably added within the last 2 m.y., since most of the volcanic

activity in the region was in the span 1.1 - 1.4 million years (Doell and others, 1968). Increased temperature would cause thermal expansion of the mineral grains in the crystalline rocks, leading ultimately to differential stresses between grains. The magnitude of stress would be dependent on the difference in the pre- and syn-intrusion temperatures. Stresses could conceivably be large enough to fracture grains.

In order to establish an equilibrium stress field in the rock, effects such as solid-state plastic flow (creep) and diffusion would tend to redistribute material from high differential stress points; this would require a concomitant porosity reduction. It is unlikely that intragranular porosity (micropores in plagioclase crystals, fluid inclusions and so forth) would be affected as much as intergranular porosity (low and high aspect-ratio channels and microcavities). The result would be the driving off of fluid from the rock, or increasing the pore pressure in the rock if fluid is present during the temperature increase.

The length of time in which these porosity changes take place is a critical part of this analysis. Creep and related phenomena act very slowly, particularly at relatively low temperatures (Carter, 1976). The effect of fluid on creep and plastic behavior of rock is two-fold. Water tends to enhance ductility by increasing the number of slip dislocations and decreasing the temperature of recovery and recrystallization (Griggs and Blacic, 1965; Griggs, 1967; Blacic, 1971, 1975; Griggs, 1974). The chemistry of these solutions may also effect processes such as pressure solution (Sprunt and Nur, 1976) and ductility (Griggs and Handin, 1960).

As previously mentioned, both the 8580' and 9522' measurements are representative of rocks in some stage of disequilibrium with the present in-situ temperature. If we plot the temperature of the permeability minimum against the depth of the sample for both rocks (Fig. 29) and compare these points to geothermal gradients for the Rio Grande rift ($\sim 30^{\circ}\text{C}/\text{km}$, Bridwell, 1976) and the present geothermal gradient in the Fenton Hill area ($55\text{-}65^{\circ}\text{C}/\text{km}$, West and others, 1975), we note that the points do not lie on a single curve intermediate to the two gradient curves. It is perhaps reasonable to assume that the two samples represent various degrees of equilibration with respect to the present geothermal gradient. Since the phenomena which act to equilibrate stresses are exponentially dependent on temperature, it is logical to conclude that the function relating the degree or rate of equilibrium to temperature is exponentially dependent on depth or the proximity to a heat source.

At lower in-situ temperatures, the equilibration reaction is less likely to have run to completion, hence the P-T curve suggested by points of permeability minimum would grade asymptotically into the pre-intrusion geothermal gradient. This is indicated on Figure 29 by the dashed line drawn between the 8580' and 9522' points. Since only two points have been determined for this curve, it is impossible to make any predictions about the pre-intrusion gradient. The P-T curve has been drawn to approximate a $30^{\circ}\text{C}/\text{km}$ gradient, but could also have been drawn to a 25° or $35^{\circ}\text{C}/\text{km}$ gradient.

Figure 29. Temperature-depth diagram showing gradients associated with the Rio Grande Rift (A), Jemez volcanic area (B), and curve drawn between the GT-2 permeability minimums (C).

Several problems are involved in this analysis, including the inability to determine the exact temperature at which the permeability minimum occurs and the lack of an unambiguous determination of the temperature at which rocks have equilibrated. The permeability minimum is not really a minimum but rather the temperature at which a distinctive break in the negative $k - T$ slope occurs. If measurements were taken at much closer temperature intervals (2-5°C), especially near the slope break, then location of the inflection point would be much more precise.

Pre- and post-intrusion vertical movements by faulting are difficult to determine. For simplicity it was assumed that the vertical position of these rocks has remained nearly constant since uplift in Laramide time. Bandelier tuff and associated volcanoclastic rocks have been deposited in the area following intrusion; these rocks have an accumulated thickness of around 140 m as measured in the GT-2 hole (West and others, 1975). The addition of these rocks would tend to shift the minimum points upward on the depth axis of Figure 29 by 0.14 km, since these rocks were deposited following intrusion and subsequent higher heat flow values. This shift will change the overall picture only very slightly, since a small thickness of rock is involved.

Application of the analysis of permeability minimums to several other geologically interesting problems might be possible. One might determine in areas of anomalously high heat flow whether the heat flow is a relatively steady-state phenomena or a transient or recent effect, possibly associated with igneous intrusion. This could be determined by establishing several permeability minimum points and then comparing

them to the measured geothermal gradient in the region. If the minimum points coincide with the gradient, then the rocks have been at the higher temperature for a considerable length of time. If the permeability minimum lies below the gradient, then equilibration has not yet occurred, as was seen in the GT-2 samples.

Another application might be as a geothermometer for rocks that have equilibrated at higher temperatures and pressures and are then rapidly brought to the surface. These types of rocks would include lherzolite inclusions in basalts and mafic inclusions in granitic rocks. The temperature range required for observation of a minimum would be much higher, on the order of 1000°C or more, and also the recovery processes such as creep and diffusion would be much more rapid at these temperatures.

Finally, in conjunction with creep and strain measurements made on crystalline rocks, this technique might be used to approximately date intrusions, where the degree of equilibration at a specific temperature is a function of time.

CONCLUSIONS

Permeability of the GT-2 rocks was measured as a function of pressure, temperature and changes due to geochemical effects associated with the movement of pore solutions. The permeability curves as a function of effective confining pressure for the GT-2 rocks were determined to have a similar shape as those for the Westerly granite. Variations in absolute permeability of the three samples were not extremely large at 21°C, while the measurements of minimum permeability at moderate temperatures were much closer in absolute value. These similarities may reflect comparable channel morphology and size, especially when the rocks are at conditions approximating the in-situ environment.

In accord with the law of effective stress, permeability-differential pressure measurements have shown that the permeability of the rock is determined by the maximum stress to which the sample is subjected. In these experiments, the maximum effective stress would be the difference of the confining pressure and the pressure of the lower pressure line.

Permeability as a function of temperature was found to be distinctive for each sample. Permeability of Westerly granite was observed to reach a slight minimum below 100°C and then to rise exponentially as temperature was increased. The permeability of sample GT-2-8580' was observed to drop from the initial value at 21°C to a minimum at approximately 130°C, after which permeability increased significantly with temperature. The permeability of sample GT-2-9522' was observed to drop sharply to a minimum near 140°C, after which the permeability

increased only slightly with temperature. These changes were generally observed to be less noticeable at higher confining pressures.

Differential thermal expansion (DTE) is postulated to have caused the phenomena observed in the temperature-permeability plots of both GT-2 and Westerly samples. DTE is due to the anisotropy and inhomogeneity of the thermal expansion of the various minerals in the rock. Quartz has a large coefficient of thermal expansion and is present in all of the samples tested. Quartz is believed to be the focus of many of the thermal expansion-related effects.

The slight drop and subsequent rise in the Westerly permeabilities can be explained by assuming that as the grains expand with increasing temperature, intergranular channels between grains will initially close, followed by grain collision which causes stresses to build up within grains (particularly quartz) until brittle fracture takes place. This apparently occurs at a temperature of around 100°C in the Westerly samples. Permanent hysteresis in permeability of the Westerly took place upon cooling, presumably due to the presence of newly formed fractures.

Using the SEM, visual observations of quartz grains in the Westerly have provided support for the DTE theory. Quartz grains in the heated sample are noticeably fractured, while quartz in the unheated sample is relatively unfractured.

Permeability behavior as a function of temperature in the GT-2 samples is postulated to have been caused by DTE, although the permeability-temperature behavior is quite different for the two rock types. Both GT-2 permeabilities drop rather sharply to a minimum

with increasing temperature. This effect can be explained by assuming that the rocks have equilibrated with respect to permeability for a particular set of temperature and pressure conditions. At these temperatures and pressures the rocks would presumably show a permeability minimum. Upon removal to the surface and lower temperatures, thermal contraction within the rock causes the opening of large intergranular channels. Upon re-heating, these channels again close, causing the permeability to drop, reaching its minimum at the equilibration temperature.

The GT-2-8580' sample equilibrated at a temperature approximately 10-20°C lower than that of GT-2-9522'. Heating to temperatures above the permeability minimum causes brittle fracturing within and around grains on both samples. This fracturing was more noticeable in the 8580' measurements since the equilibration temperature was lower.

Visual observations of GT-2-8580' and GT-2-9522' using the SEM support the DTE mechanism. Unheated 8580' shows channels somewhat similar to those in the Westerly, although they have a slightly more fractured appearance. After heating, channels within the 8580' sample are distinctively larger and show some brittle fracture characteristics. Intergranular channels in the unheated 9522' are characteristically much better defined and only occasionally appear similar to those in the Westerly. Channels in heated and unheated 9522' are very similar, probably because only the contraction-induced cavities are being closed, and very little stress damage is occurring. This is in contrast to GT-2-8580', where significant channel damage took place with heating.

Contraction-induced channels and fractures are thought to be identical to the isotropically oriented fractures observed by Simmons and Cooper (1976) in similar GT-2 samples. They have stated that these features were induced during drilling, sample preparation or removal to the surface. They have also postulated that since these fractures are present within the sample, laboratory permeability, compressibility and related measurements on these samples would be from several times to orders of magnitude higher than the in-situ values. However, the experiments of Simmons and Cooper (1976) were run at room temperature and did not take into account the effects of thermal expansion on crack and channel distribution and morphology. I contend that laboratory measurements on variables such as permeability and compressibility on the GT-2 samples can rather closely approximate in-situ values if both temperature and pressure are taken into consideration.

It is also postulated that rocks such as Westerly granite, which have equilibrated in a much lower temperature regime, might show realistic behavior at room temperatures but would deviate from expected behavior with increasing temperature. Certainly the stress distribution within the sample would no longer be known, since high stresses within grains such as quartz have been shown to occur with increasing temperature.

Permeability was observed to change as pore fluid was passed through the sample for an extended period of time. An increase in permeability of two to three times was observed following flow of distilled water through the sample for 120 hours. Dissolution of

minerals, particularly quartz, along flow channels is thought to be the mechanism for this change in permeability. The formation of clays and alteration products along channels was not observed in this study, although this has been reported elsewhere in somewhat similar flow experiments (Summers and others, 1975; Balagna and Charles, 1975).

Flow-channel tracing using a radioactive tracer and autoradiography has proven valuable in delineating the spatial relationships of channels with respect to grains and grain boundaries. Initial experiments using ^{22}Na yielded low resolution autoradiographs due to the high energy of the emitted beta particle. Subsequently ^{63}Ni was used and gave excellent resolution due to the low energy of the emitted beta particle. The ^{63}Ni autoradiographs indicate that the majority of flow through the samples is along low intergranular aspect ratio channels (LARC's). Autoradiography also suggested the presence of numerous intragranular channels in the larger biotite grains; however upon further examination, it was discovered that these channels separate crystallographically distinct biotite grains in larger multigranular clumps of biotite.

^{35}S tracing experiments were run to determine whether deposition of the radioactive species was physically or chemically controlled. The ^{35}S autoradiograph was similar to that of the ^{63}Ni experiment in that the intergranular channels in the rock are well documented, and deposition appears to be primarily physically controlled. Resolution of the ^{35}S isotope is slightly poorer owing to the higher energy of the emitted beta particle. Future work and improvements in the

radiograph tracing technique might include: 1) use of finer grain x-ray film, 2) use of higher concentrations of radioactivity, and 3) use of an alpha emitter as the isotope species, since the alpha particle has a much lower penetration range in granite and thus would yield higher resolution.

SUMMARY

- 1) The permeability of quartz monzonite from the Los Alamos Scientific Laboratory hot-dry-rock geothermal well GT-2 and Westerly granite was measured experimentally as a function of temperature and pressure. Flow rate through the sample was measured using a transient pressure decay method. When combined with certain physical parameters of the sample and permeameter, this flow rate yielded the permeability (k).
- 2) Permeability of both GT-2 and Westerly samples behave similarly as a function of confining pressure, indicating similar flow channel morphology.
- 3) The behavior of the permeability of the GT-2 and Westerly rocks is markedly different as a function of increasing temperature. Both GT-2-8580' and GT-2-9522' go through distinct permeability minimums. The 8580' minimum is located near 130°C, while the 9522' minimum is closer to 150°C. Permeability of the Westerly granite, however, shows only a very shallow minimum from 80 to 100°C followed by a sharp rise at higher temperatures. The permeability-temperature phenomena observed in both GT-2 and Westerly measurements are postulated to be the result of differential thermal expansion (DTE). DTE is due to anisotropic and inhomogeneous coefficients of thermal expansion of the grains within the rock. Visual observations on unheated and heated GT-2 and Westerly samples support the DTE hypothesis.

- 4) Differences in the behavior of the Westerly, GT-2-8580' and GT-2-9522' permeabilities with temperature are believed to be due to the respective pressure-temperature environments in which they equilibrated. GT-2 rocks have recently been at higher pressures and temperatures than the Westerly granite. Temperature disequilibrium due to increases in the geothermal gradient accompanying intrusion of the Jemez pluton may have affected the GT-2 samples. Analysis of this disequilibrium may allow determination of the pre-intrusive geothermal gradient of the area.
- 5) A technique was developed for the delineation of flow channels in the GT-2 rocks using radioactive tracers. Three radioactive isotopes were used in these experiments: ^{22}Na , ^{63}Ni , and ^{35}S . Cores were injected with the radioisotope using the permeability apparatus; they were then thick-sectioned and autoradiographed.
- 6) ^{63}Ni provides the best resolution of the three isotopes, since it emits the lowest energy beta particle. The channel tracing work indicates that at low pressures the majority of fluid transport through the rock is along intergranular channels. Intragranular channels were not observed with this technique.

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REFERENCES CITED

- Adams, L. H., and Williamson, E. D., 1923, On the compressibility of minerals and rocks at high pressures: J. Franklin Inst., v. 195, p. 475-529.
- Anderson, O. L., and Grew, P. C., 1977, Stress corrosion theory of crack propagation with application to geophysics: Rev. Geophys. and Space Physics., v. 15, p. 77-104.
- Aruna, M., 1976, The effects of temperature and pressure on the absolute permeability of sandstones: Ph.D. dissertation, Palo Alto, Stanford University, 310 p.
- Balagna, J. P., and Charles, R. W., 1975, Permeability of a monzogranite at elevated temperatures (abs.): EOS, Trans. Am. Geophys. Union, v. 52, p. 342.
- Baldrige, S., and Simmons, G., 1975, Progress in microcrack decoration (abs.): EOS, Trans. Am. Geophys. Union, v. 52, p. 342.
- Birch, F. J., 1961, The velocity of compressional waves in rocks to 10 Kilobars; Part 2: J. Geophys. Res., v. 66, p. 2199-2224.
- Birch, F. J., 1966, Compressibility: in "Handbook of Physical Constants", S. P. Clark ed., Geol. Soc. Amer. Mem. 97, p. 97-173.
- Birch, F. J., Schairer, J. F., and Spicer, H. C., 1942, Handbook of physical constants: Geol. Soc. Amer. Spec. Paper 36, p. 39-62.
- Blacic, J. D., 1971, Hydrolitic weakening of quartz and olivine: Ph.D. dissertation, Los Angeles, Univ. of Calif., 216 p.
- Blacic, J. D., 1975, Plastic deformation mechanisms in quartz: The effect of water: Tectonophysics, v. 27, p. 271-294.
- Brace, W. F., 1965, Some new measurements on the linear compressibility of rocks: J. Geophys. Res., v. 70, p. 391-398.
- Brace, W. F., 1972, Pore pressure in geophysics: Am. Geophys. Union, Geophysical Monograph 16, p. 265-274.
- Brace, W. F., Orange, A. S., and Madden, T. M., 1965, The effect of pressure on the electrical resistivity of water saturated crystalline rocks: J. Geophys. Res., v. 70, p. 5669-5678.
- Brace, W. F., and Orange, A. S., 1968, Further studies of the effects of pressure on the electrical resistivity of rocks: J. Geophys. Res., v. 73, p. 5407-5420.

- Brace, W. F., Walsh, J. B., and Frangos, W. T., 1968, Permeability of granite under high pressure: *J. Geophys. Res.*, v. 73, p. 2225-2236.
- Brace, W. F., Silver, E., Hadley, K., and Goetze, C., 1972, Cracks and pores: A closer look: *Science*, v. 178, p. 162-163.
- Bridwell, R. J., 1976, Lithospheric thinning and the late Cenozoic thermal and tectonic regime of the northern Rio Grande Rift: *New Mexico Geol. Soc. Guidebook*, 27th Field Conf., p. 283-292.
- Carter, N. L., 1976, Steady state flow of rocks: *Reviews of Geophys. and Space Physics*, v. 14, p. 301-359.
- Doell, R. R., Dalrymple, G. B., Smith, R. L., and Bailey, R. A., 1968, Paleomagnetism, potassium-argon ages, and geology of rhyolites and associated rocks of the Valles Caldera, New Mexico: *Geol. Soc. Amer. Mem.* 116, p. 211-248.
- Fisher, H., 1977, In-situ permeability determinations in LASL drill hole GT-2, LASL internal report, in press.
- Griggs, D. T., 1967, Hydrolytic weakening of quartz and other silicates: *Geophys. Jour.*, v. 14, p. 19-31.
- Griggs, D. T., 1974, A model for hydrolytic weakening of quartz: *J. Geophys. Res.*, v. 79, p. 1653-1661.
- Griggs, D. T., and Handlin, J. W., 1960, Observations on fracture and a hypothesis of earthquakes: *Geol. Soc. Amer. Mem.* 79, p. 347-364.
- Griggs, D. T., and Balci, J. D., 1965, Quartz: Anomalous weakness of synthetic crystals: *Science*, v. 147, p. 292-295.
- Handbook of Chemistry and Physics, 1970, R. C., Weast, ed., Chemical Rubber Co., Cleveland, Ohio, p. F-36.
- Koide, H., and Hoshino, K., 1967, Development of microfractures in experimentally deformed rocks (preliminary report): *Jishin*, v. 20, p. 85-97.
- Laughlin, A. W., and Eddy, A., 1977, Petrography and mineralogy of crystalline rocks from the LASL drill hole GT-2, LASL internal report, in press.
- Martin, R. J., III, 1972, Time-dependent crack growth in quartz and its application to the creep of rocks: *J. Geophys. Res.*, v. 77, p. 1406-1419.

- Martin, R. J., III, and Durham, W. B., 1975, Mechanism of crack growth in quartz: *J. Geophys. Res.*, v. 80, p. 4837-4844.
- Montgomery, C. W., and Brace, W. F., 1975, Micropores in plagioclase: *Contrib. Miner. Petrol.*, v. 52, p. 17-28.
- Ohle, E. L., 1951, The influence of permeability on ore distribution in limestone and dolomite: *Econ. Geol.*, v. 46, p. 667-706.
- Paulding, B. W., Jr., 1965, Crack growth during brittle fracture in compression: Ph.D. dissertation, Cambridge, Mass., Mass. Inst. of Technology, 174 p.
- Peng, S., and Johnson, A. M., 1972, Crack growth and faulting in cylindrical specimens of Chelmsford granite: *Int. J. Rock Mech. Min. Sci.*, v. 9, p. 37-86.
- Richter, D., Simmons, G., and Siegfried, R., 1976, Microcracks, micropores, and their petrologic interpretation for Lunar Samples 72415 and 15418: *Proceed. Seventh Lunar Conf.*, v. 2, p. 1901-1923.
- Simmons, G., and Cooper, H., 1976, DSA of the microcracks in GT-2 core: Interpretation and implications: Report to Los Alamos Scientific Lab. on contract X66-54922-1, p. 30-37.
- Simmons, G., and Cooper, H., 1977, Thermal cycling cracks in crystalline rocks, *Earth and Planet. Sci. Letters*, in press.
- Skinner, B. J., 1966, Thermal expansion: in "Handbook of Physical Constants", S. P. Clark, ed., *Geol. Soc. Amer. Mem.* 97, p. 75-96.
- Sprunt, E. S., and Brace, W. F., 1974, Direct observation of microcavities in crystalline rocks: *Int. J. Rock Mech Min. Sci.*, v. 11, p. 139-150.
- Sprunt, E. S., and Nur, A., 1976, Strain and solution rate: An empirical relationship (abs.): *EOS, Trans. Am. Geophys. Union*, v. 56, p. 1060.
- Summers, R., Winnkler, K., and Byerlee, J., 1975, Permeability changes during fluid flow through hot granite (abs.): *EOS, Trans. Am. Geophys. Union*, v. 56, p. 1060.
- Timur, A., Hemphkins, W. B., and Weinbrandt, R. M., 1971, Scanning electron microscope study of pore systems in rocks: *J. Geophys. Res.*, v. 76, p. 4932-4948.

- Tood, T., 1973, Effects of cracks on elastic properties of low porosity rocks: Ph.D. dissertation, Cambridge, Mass., Mass. Inst. of Technology, 201 p.
- Walsh, J. B., 1965, The effect of cracks on the uniaxial compression of rocks: J. Geophys. Res., v. 70, p. 399-411.
- Walsh, J. B., and Brace, W. F., 1966, Elasticity of rock: A review of recent theoretical studies: Rock Mech. Eng. Geol., v. 4, p. 283-297.
- Walsh, J. B., and Decker, E. R., 1966, Effect of pressure and saturating fluid on the thermal conductivity of compact rock: J. Geophys. Res., v. 71, p. 3053-3061.
- Wawersik, W. R., and Brace, W. F., 1971, Post failure behavior of granite and diabase: Rock Mech., v. 3, p. 61-85.
- West, F. G., Kintzinger, P. R., and Laughlin, A. W., 1975, Geophysical logging in the Los Alamos Scientific Laboratory Geothermal test holt No. 2: Los Alamos Report LA-UR-75-1518, p. 1-6.
- Woodward, L. A., Kaufman, W. H., and Anderson, J. B., 1974, Nacimiento fault and related structures, northwestern New Mexico: Geol. Soc. Amer. Bull., v. 83, p. 2382-2396.

APPENDIX 1

Sample permeability calculation from actual 21°C measurement.

Pressure conditions $P_c = 171.2$ Bars
 $P_1 = 137.3$ Bars
 $P_2 = 133.9$ Bars
 $\Delta P = 3.42$ Bars
 $\bar{P} = 137.3$ Bars

Measured change in pressure with time

Time (seconds)	Pressure (Bars)
0	137.318
10	137.284
20	137.270
30	137.246
40	137.233
50	137.216
60	137.205

To calculate the flow rate through the sample, the pressure drop over a specified time interval is chosen. This interval is chosen so that the slope of the pressure time plot is approximately linear. For instance the pressure time slope is approximately linear through the whole region of this measurement so an arbitrary value at 30 seconds is chosen to make the measurement.

Change in time = 30 seconds. Change in pressure = .072 Bar.

To calculate the flow rate through the sample equation (3) is used

$$Q = \beta \cdot V_s \cdot (\Delta P / \Delta t) \quad , \quad (3)$$

where

$$\beta = 4.2 \times 10^{-11} \text{ cm}^2 / \text{dyne}$$

$$V_s = 36 \text{ cm}^3$$

$$(\Delta P / \Delta t) = .072 \times 10^6 \text{ dynes} / \text{cm}^2 / 30 \text{ sec}, = 2.4 \times 10^3 \text{ dynes} / \text{cm}^2 / \text{sec}$$

$$Q = 3.629 \times 10^{-6} \text{ cm}^3 / \text{sec}.$$

The permeability can then be calculated from equation (4),

$$k = \frac{Q \cdot \mu \cdot L}{A \cdot \Delta P} \quad (4)$$

where

$$\mu = 1.0 \times 10^{-2} \text{ dyne} \cdot \text{sec} / \text{cm}^2$$

$$L = 1.58 \text{ cm}$$

$$A = 5.06 \text{ cm}^2$$

$$\Delta P = 3.42 \times 10^6 \text{ dynes} / \text{cm}^2$$

$$k = 3.31 \times 10^{-15} \text{ cm}^2 \quad , \text{ one darcy} = 10^{-8} \text{ cm}^2 \text{ so}$$

$$k = 3.31 \times 10^{-7} \text{ darcy or } 331 \text{ nanodarcy}.$$