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Thermal Expansion and Cracking of Three Confined, Water-Saturated Igneous Rocks to 800° C

By

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Summary

Solutions of engineering problems of very deep drilling, geothermal energy production, and high-level nuclear-waste isolation require adequate understanding of the mechanical and transport properties of rocks at relatively low pressures but high temperatures. Accordingly, the thermal expansions of water-saturated Charcoal Granite, Mt. Hood Andesite, and Cuerbio Basalt have been measured at effective confining pressures (P_e) of 5, 50, and 100 MPa to 800° C. The mean coefficient of linear thermal expansion (α) is a function of lithology, P_e, temperature (T) and initial porosity (ϕ). For example, for the Charcoal Granite, α increases with T at all pressures. The signature of the alpha-beta transition of quartz is more pronounced at the lower pressures; at 100 MPa α nearly mimics that of a crack-free rock for $T < 300^0$ C.

α for the andesite at $P_e = 5$ MPa ranges from 10 to $15 \times 10^{-6}/{}^0$C from 200° to 400° C then decreases gradually to $10.1 \times 10^{-6}/{}^0$C at 800° C. At 50 MPa α ranges from $11.7 \times 10^{-6}/{}^0$C at 100° C to 8.6×10^{-6} at 200° C, then increases at a much lower rate to 11×10^{-6} at 600° C. The basalt, however, has an essentially constant α ($11 \times 10^{-6}/{}^0$C) for $T > 150^0$ C at the lower pressure and shows but a small increase in α from 6 to 9×10^{-6} from 100° to 800° C at 50 MPa.

The difference between measured values of thermal expansion and those calculated from simple mixture-theory relates to new crack porosity generated as a result of differential thermal expansion at the anisotropic grain scale. For the granite, a two to three order of magnitude increase in permeability (k) is predicted from the relation, $k \propto \phi^3$.

Introduction

Dealing with high-temperature environments is a relatively new challenge to engineering rock-mechanics. The need for adequate predictive thermomechanical modeling has grown rapidly along with recent developments in the *in-situ* retorting of coal and oil "shale", tertiary oil-recovery, geothermal-energy exploitation, underground storage of fluids at temperatures well above or below ambient, and isolation of heat-generating, high-level, nuclear waste (Green, 1979). One important material property is thermal expansion.

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In Handbook of Physical Constants, published but 15 years ago, Skinner (1966) lists no data on the thermal expansions of rocks under confining pressures nor at temperatures above 100° C. The term, thermal expansion, does not appear in the index of the latest edition of Jaeger and Cook's (1979) standard textbook. The laboratory-data base relevant to prototype problems is still seriously wanting (Handin and Heard, 1980).

Associated with thermal expansions and contractions is the important phenomenon of thermal cracking. The development of crack porosity can substantially modify mechanical and transport properties such as thermal conductivity, permeability and strength (Richter and Simmons, 1974; Sprunt and Brace, 1974 a, 1974 b; Zoback and Byerlee, 1975; Cooper and Simmons, 1977; Simmons and Cooper, 1978; Bauer and Johnson, 1979; Wong and Brace, 1979; Heard, 1980; Handin and Heard, 1980; Trimmer et al., 1980; Bauer and Handin, 1981; Durham and Abey, 1981; Page and Heard, 1981; Heard and Page, 1982).

It has long been recognized that fracturing is associated with the steep thermoelastic stress-gradients in rapidly heated or cooled materials (thermal shock). Richter and Simmons (1974) show, for example, that the thermal expansion of unconfined granite from 20° to 250° C is fully recoverable for a heating rate of 1° C per minute, but not for 5° C per minute, owing to the permanent strain left after microcracking. Fractures can also develop where heating is slow but non-uniform as, for example, in the inhomogeneous stress field of a thick-walled hollow cylinder heated externally and cooled internally (Johnson and Gangi, 1980).

Also well known is the thermal cracking related to the intergranular stresses developed in polycrystalline aggregates — rocks and other composites like ceramics (Kingery et al., 1976). These stresses arise from the mismatch of thermal expansions across grain boundaries. In a monomineralic rock like quartzite, the mismatch is due to the different crystallographic orientations of anisotropic grains. Most igneous rocks consist of two or more major phases with different thermoelastic properties; differential thermal expansions give rise to tensile and compressive stresses that must balance in a circumscribed volume in static equilibrium (Turner, 1946; Kerner, 1956; Walsh, 1973; Bruner, 1979). There is a threshold temperature below which no thermal fracturing is detectable even by highly sensitive acoustic-emission measurements; it depends on the composition and fabric of the rock and on confining pressure which suppresses cracking (Richter and Simmons, 1974; Johnson et al., 1978; Wong and Brace, 1979).

Few rocks are free of microfractures. Differential thermal expansion can extend and widen pre-existing cracks, create new inter- and intragranular cracks where local tensile stresses exceed bond strengths, and then extend and widen both new and old cracks with a further increase in temperature (Sprunt and Brace, 1974 a; Johnson et al., 1978; Friedman and Johnson, 1978).

As part of our on-going program of research on the mechanical and transport properties of water-saturated igneous rocks, we have measured the thermal expansions and inferred the thermal cracking of Charcoal granodiorite, Mt. Hood andesite, and Cuerbio basalt under effective confining pressure of 5, 50, and 100 MPa to temperature of 800^0 C.

During the procedure for the triaxial-compression testing of strengths and ductilities of these rocks (Friedman et al., 1979; Bauer et al., 1981), thermal expansions can be measured readily and fairly accurately. Inasmuch as the specimens are deviatorically loaded immediately following attainment of pressure-temperature equilibrium in a confined specimen, we cannot measure corresponding contractions during cooling, nor can we examine a deformed specimen for thermal cracking alone. However, previous work on dense crystalline rocks has shown that virtually all thermal cracking occurs during the heating and not the cooling phase and not subsequent cycling to the same maximum temperature (Richter and Simmons, 1974; Bauer and Johnson, 1979). It is also evident that inelastic (permanent) expansions are due almost entirely to the generation of microfractures and/or the extension and widening of them along with any pre-existing cracks. Thus, by subtracting the theoretical elastic expansion of the uncracked rock from our observed expansions, we can infer the increase of crack porosity with temperature at a given confining pressure, even though we can not observe the thermal cracking directly. We can also infer the relative change of permeability, provided that it is all due to the increase of connected crack-porosity.

Previous Work

There are few previous data on confined igneous rocks with which to compare ours. The first measurements to be made under confining pressure appear to be those of Wong and Brace (1979) on copper-jacketed specimens to which electrical-resistance strain gages were bonded. Typical average values of α ($10^{-6}/^0$C) over the narrow range of 2^0 to 38^0 C at 100 MPa confining pressure are 7.7 for Westerly granite and 7.8 for Frederick diabase. The critical pressure above which the thermal expansion is reversible is of the order of 100 to 150 MPa, about the same as the crack-closing pressure inferred from compressibility (Brace, 1965) and electrical-resistivity (Brace et al., 1965) measurements on these same rocks. The theoretical expansions of the polycrystalline aggregates were calculated by the method of Walsh (1973) under the assumptions that the anisotropic mineral phases are elastic and randomly oriented, and that the thermoelastic constants of all phases are independent of pressure and temperature. The calculated values of about $5 \times 10^{-6}/^0$C do not agree well with those observed; whereas, the coincidence is remarkably close, 10.69 and 10.6 ± 0.17, for the monomineralic Cheshire quartzite. Supposing that the maximum internal differential stress, $\sigma \sim E \Delta\alpha \Delta T$, where E is average Young's modulus and $\Delta\alpha$ is the thermal-expansion contrast across the grains, Wong and Brace discount the creation of new thermal cracks for the small $\Delta T = 36^0$ C. Hysteretic behavior is ascribed instead to the deformations of pre-existing cracks.

Using a technique essentially similar to ours, Heard (1980) got values of α ($10^{-6}/^0$C) for dry Climax quartz monzonite of 10—15, 7—13, and 7—12 under confining pressures of 0, 13.8, and 27.6 MPa, respectively, over the range of 19⁰ to 300⁰ C. The largest permanent strain is small, about 2×10^{-4}. Nevertheless, by comparing observed expansions with those predicted, as limited by availability of thermoelastic data on quartz and potash- and plagioclase feldspars, he concludes that thermal-crack porosity (ϕ) does develop at low pressure and high temperature, and that permeability (k) should increase fivefold if $k/k_0 \propto (\phi/\phi_0)^3$. Thermal expansions are clearly non-linear functions of both pressure and temperature.

Page and Heard (1981) made further measurements of the elastic moduli and thermal expansion of this quartz monzonite and also of Sudbury gabbro over the range of 19⁰ to 500⁰ C and 7 to 55 MPa confining pressure. Heard and Page (1982) extended these measurements to Westerly granite (19⁰—300⁰ C) and granite from the Stripa mine in Sweden (19⁰—350⁰ C). Permanent strains did not exceed 1.5×10^{-4}. Heating all rocks from 19⁰ to 300⁰ C reduced Young's modulus by some 50%, the largest effects being at the lowest pressures as expected. Thermal expansions were measured at five pressures and 40⁰—50⁰ temperature intervals. Measurements in three orthogonal directions reveal that all these rocks are statistically isotropic. The standard deviations of the values of α are much larger than the uncertainties of their measurements reflecting the heterogeneity from sample to sample. The average mean α for gabbro varies from 6—$11 \times 10^{-6}/^0\mathrm{C}$ over the full range of conditions, being highest at the lowest pressures. For quartz monzonite, α ranges from 8 to 11 at 19⁰ to 60⁰ C and 22 to 25 at 450⁰ to 500⁰ C. For Westerly granite, α increases at all confining pressures from 7—$17 \times 10^{-6}/^0\mathrm{C}$ at low temperatures, to 10—14 near 300⁰ C. However, for Stripa granite at low to intermediate temperatures, values of α of 6 to $8 \times 10^{-6}/^0\mathrm{C}$ are less than that calculated for the crack-free rock, then rising above it to 10—16 at 200⁰ C. At 300⁰ C, α converges to about 13 at all confining pressures.

At 300⁰ C the expansions of all these rocks exceed those calculated for the crack-free polycrystalline aggregates. If this divergence owes to thermal cracking, then the increase in crack porosity, $\Delta\phi \propto \Delta T$, ΔT is multiplied by the quantity enclosed by parentheses (difference between observed and calculated values of α). The initial porosity at atmospheric pressure (ϕ_0) is corrected for the crack-closure pressure. For quartz monzonite at 500⁰ C and the lowest pressure, 7 MPa, ϕ doubles from 6×10^{-3} to 12×10^{-2}. If permeability changes as $k/k_0 \propto (\phi/\phi_0)^3$, it increases by a factor of 8 from 19⁰ to 500⁰ C. Heating to 300⁰ C causes 2-fold and 3-fold increases of k of Stripa and Westerly granite, respectively.

Durham and Abey (1981) made simultaneous measurements of thermal expansion and of thermal conductivity and diffusivity of large hollow cylinders (length = 20 cm, outside diameter = 13.7 cm, inside diameter = 1.9 cm) of this same quartz monzonite (and Avery Island rocksalt) to about 500⁰ C under 50-MPa confining pressure. Their values of α are much higher, 10—15 ($10^{-6}/^0\mathrm{C}$), than Heard's, owing to lack of calibration of their strain-measuring system. Nevertheless, they do demonstrate that thermal cracking lowers thermal conductivity significantly.

Following a procedure similar to Heard's (1980) and ours, van der Molen (1981) measured the linear compressibility at room temperature and thermal expansion of both wet and dry, fine-grained granite (Delegate aplite) up to 900° C under confining pressures of 100, 200, and 300 MPa. We know of no other published thermal-expansion measurements on water-saturated igneous rocks. Unfortunately, however, his tests were undrained. Thus, although the volume of water added to a specimen is known, the pore pressure, and hence the effective confining pressure is not.

The observed decrease in porosity with heating of the dry rock to 200° C, also noted by Page and Heard (1981), is ascribed to the expansions of grains into pre-existing cracks, and the increase in porosity above 200° C to new thermal cracking, owing to the differential thermal expansions of highly thermoelastically anisotropic quartz, orthoclase and plagioclase. This reduction of porosity at low temperature and high pressure is not seen in wet specimens, presumably because pore pressure suppresses internal expansions. Again, confining pressure inhibits thermal cracking; extrapolation suggests that it would cease altogether at about 450 MPa at any temperature.

The expansions of van der Molen's specimens are complicated by the α-β phase transition of quartz which occurs at 573° C at atmospheric pressure and rises about 26° C per 100 MPa in single crystals. The sharp kinks in the thermal-expansion curves that mark this transition reflect an increase of 60° to 70°, several possible explanations for which are proposed. Whatever the reason for it may be, this shift of the transition temperature is not significant to our conclusions.

Between 200° and 840° C, volume increases by about 4.0, 3.6, and 2.6% for 0, 100, 200, and 300-MPa confining pressure, respectively. A calculation of crack porosity is not attempted, but the upper bound, calculated for the expansion of the uncracked polycrystalline aggregate is estimated as 2%. The thermal-expansion curves for dry and wet conditions are essentially parallel, but because little or no contraction occurs at low temperature in wet specimens, their net expansions are somewhat the greater. Although the evidence is not entirely clear, the presence of water does not seem to influence the quartz transition.

Experimental Procedure

The starting materials and the procedure for triaxial-compression testing at high temperature (T) and with independently controlled external confining pressure (P_c) and internal pore pressure (P_p) are fully described elsewhere (Friedman et al., 1979; Bauer et al., 1981). During this procedure it is relatively easy to measure fairly accurately the expansion of a specimen upon heating under an effective confining pressure ($P_e = P_c - P_p$).

The copper-jacketed, 2 by 4-cm cylindrical specimen is sealed to the upper and lower pistons, emplaced in the internally heated triaxial cell, and subjected to a small axial force to seal off the pore-water inlet (Fig. 1). The confining pressure of argon and pore pressure of water, measured by precision Heise gauges, and axial force, measured by a Baldwin CXX gauge, all to ±0.1 percent, are raised together so as to maintain hydrostatic pressure. The specimen is then heated at about 5^0 C per minute while the confining and pore pressures are adjusted manually so as to hold the effective confining pressure constant within about 0.5 MPa. At this relatively fast rate of heating, there are transient thermal gradients within specimens, owing to the relatively low thermal conductivities even of water-saturated rocks, but such gradients are almost surely smaller than the steady lengthwise gradient described below.

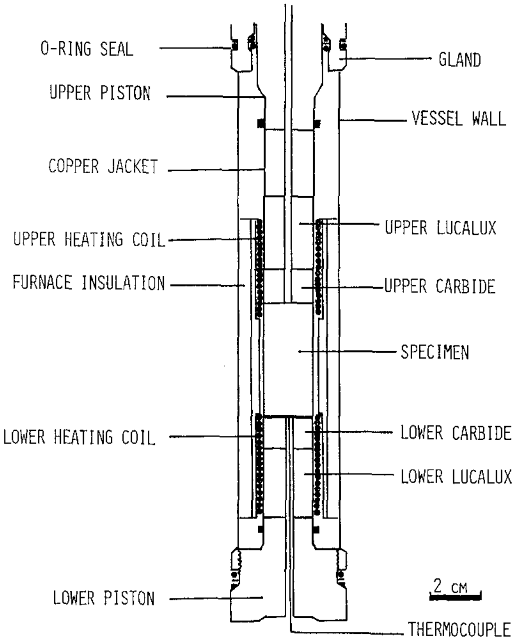

Fig. 1. Loading column

The nominal temperature can be read within ± 1^0 C near the base of the specimens by a thermocouple inserted through the lower piston. Regulation is good to ± 2^0 C. However, the gradients in the specimen are much larger, owing to convection in the confining gas. The temperature distribution along gauge length was measured by moving a thermocouple from top to bottom in the central hole of a dummy specimen. Two separately powered, electrical-resistance, heating coils are used to help smooth the gradient. Furthermore, the space between the walls of the furnace and specimens are packed with boron-nitride powder. Even so, the temperature variation is 2.5 percent. For example, at a mean temperature of 600^0 C at the midpoint of the specimen, the range is about 592^0 to 608^0 C from bottom to top.

As the temperature rises, the entire loading column, including the steel pistons, carbide and lucalox spacers, and specimen itself, expands, and the axial differential stress tends to increase. However, the axial load is continuously adjusted by bleeding oil from the hydraulic ram, thus allowing the loading column to displace and maintaining the load constant to within ±4 MPa. Since Young's modulus of crystalline rocks is of the order of 100 GPa, the useful sensitivities of corresponding axial strain and displacement measurements are 4×10^{-5} and 1.6×10^{-4} cm, respectively. The piston displacements are measured externally by a HP-7-500 DCDT with an accuracy of ±0.5 percent and displayed by a Hewlett-Packard X-Y recorder. After digitization the precision of the displacement data is $\pm 2 \times 10^{-4}$ cm, corresponding to $\pm 5 \times 10^{-5}$ in strain.

What is observed is the total expansion of the entire loading column. The expansion of the specimen alone is determined by subtracting the sum of the expansions of all the extraneous parts, this having been determined in a test under identical conditions on a dummy specimen of steel with a known thermoelastic coefficient (Skinner, 1966).

Measurements are made at increments of 50^0 C during heating alone. At least three specimens of each rock are tested under identical conditions to assess reproducibility. The first reading is taken at 50^0 C. In our experiments the behaviour between 25^0 and 50^0 C is uncertain.

Experimental Results

Figs. 2, 3, and 4 show the so-called "linear" thermal expansions of the three rocks over the interval of 25^0 to 800^0 C under the three different effective confining pressures of 5, 50, and 100 MPa. These are the uniaxial expansions parallel to the cylindrical axis; they are not linear functions of either temperature or pressure. The ranges of measurements on at least three specimens of each rock under identical conditions are indicated by

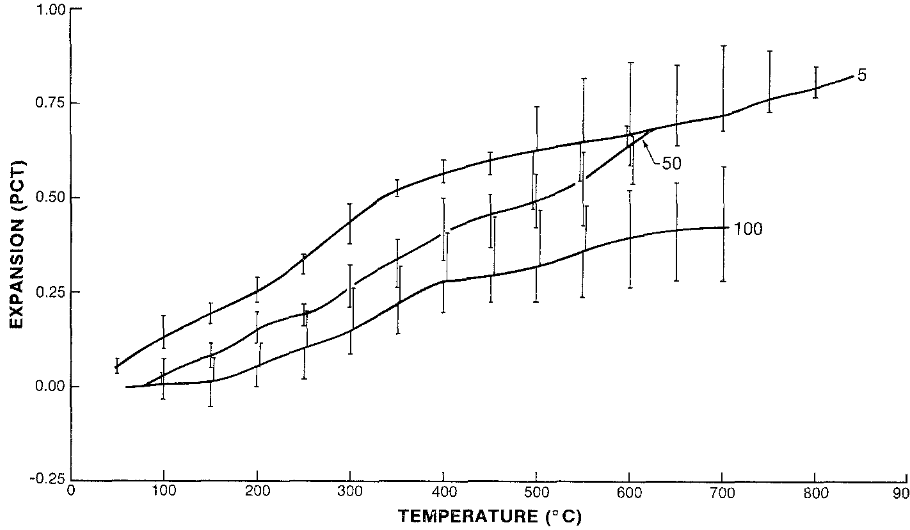

Fig. 2. Linear thermal expansion of Mt. Hood Andesite under different effective confining pressures (MPa)

error bars. The curves are drawn through the average values at each temperature. Evidently the higher is the pressure, the lower is the expansion and the better is the reproducibility at a given temperature in most tests.

Table 1 lists the mean values of the coefficients of linear thermal expansion ($\bar{\alpha}$) in the usual way (e. g., Skinner, 1966, Table 1). The mean value is the slope of the straight line connecting the point of origin (25° C) of the expansion curve to the average point at the particular temperature indicated in increments of 100° C. Also shown are the extreme deviations from the average value. The mean values do not reflect the real non-linearities of the expansion curves. Accordingly, in Table 2 we also list the incremental slopes (α) taken at the midpoint between average values at increments of 50° C.

Andesite

The andesite, collected from a recent flow on the south slope of Mt. Hood, is a porphyritic rock with an aphanitic groundmass. Phenocrysts, mostly of plagioclase, comprise 40 percent of the rocks. They are enveloped by the nearly holocrystalline groundmass. The porosity of about 8 percent is not vesicular, but owes in part to microfractures in the phenocrysts and in part to interstitial voids in the groundmass where the framework of randomly-stacked plagioclase laths has not entirely collapsed.

The resproducibility of the measurements on this rock is poor, owing probably to the heterogeneity of the starting material, especially as regards porosity. At 5-MPa confining pressure the mean coefficient of linear expansion is about $15 \times 10^{-6}/^{0}\mathrm{C}$ to 350° C, above which it declines to 13 at 500° C and 10 at 800° C. At 50 MPa, $\bar{\alpha}$ is about $10 \times 10^{-6}/^{0}\mathrm{C}$ to 600° C (the maximum achieved). The increases at 550° C, where α is 22, are not understood. At 100-MPa the slope decreases in the neighborhood of 600° C. At this pressure $\bar{\alpha}$ ranges from nearly zero at 100° C to about 6 to $7 \times 10^{-6}/^{0}\mathrm{C}$ between 300° and 800° C. We know of no other data on andesite with which to compare ours.

Basalt

The Cuerbio basalt, collected from a Pleistocene flow near La Cienega, New Mexico, consists of a fine-grained, holocrystalline groundmass (85 percent) containing small vesicles (<1 mm) and microphenocrysts (15 percent) mostly of microfractured olivine (see Table 3). The porosity of 5 to 8 percent is largely vesicular.

To date we have measured the thermal expansions of this basalt at 5 and 50 MPa, and we have checked the reproducibility at 5 MPa and 25° to 400°. However, since this reproducibility is good and the effect of pressure is relatively small, we do not expect further tests to affect our conclusions significantly.

At 5-MPa confining pressure, $\bar{\alpha}$ between 25° and 100° C is $5.3 \times 10^{-6}/^{0}\mathrm{C}$, rising to 12.5 from 25° to 200° C, then remaining essentially constant at about 11 to 800° C. The values are somewhat smaller at 50 MPa; the expansion is nearly linear between 400° and 800° C. We know of no other work on confined basalt. Foundation Sciences (1980) reports average values of 5 to $7 \times 10^{-6}/^0C$ from 25^0 to 300^0 C for unconfined samples of plateau basalts from the Hanford Reservation in Washington.

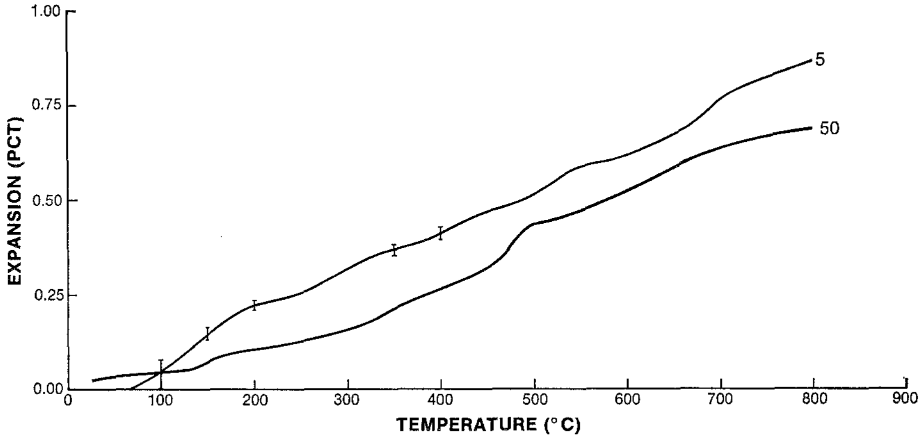

Fig. 3. Linear thermal expansion of Cuerbio Basalt under different effective confining pressures (MPa)

Granodiorite

The Charcoal (properly St. Cloud) granodiorite is a medium- and even-grained Precambrian rock from Minnesota. It is one of the standard suite of rocks selected by the U. S. Bureau of Mines, so that its chemical and physical properties are well known (Krech et al., 1974). Its small porosity of less than 0.1 percent owes largely to unhealed microfractures. Because its quartz content of 21 percent is relatively high (Table 3), the alpha-beta phase transition near 600^0 C is expected to influence thermal expansion substantially.

At 5-MPa confining pressure, $\bar{\alpha}$ is about $15 \times 10^{-6}/^0C$ at 200^0 C whereas at 50 and 100 MPa, it is about 5. From 300^0 to 500^0 C the slope is lower, at 5 MPa, about 13, but higher at 50 and 100 MPa, about 10 and 8. From about 400^0 to 600^0 C the expansion curves are quasi-parallel with an average slope of about 12. At about 600^0 C the curves steepen, $\bar{\alpha}$ at 700^0 C being 17, 14, and 12 at 5, 50, and 100 MPa, respectively. All expansion curves then tend to flatten above 650, owing to the influence of the alpha-beta transition in the quartz phase. Between 525^0 and 625^0 C, the instantaneous values of α jump from 17 to 40, 20 to 23, and 15 to 24 at 5, 50, and 100 MPa, respectively.

We have measured the expansion of a single room-dry specimen under 50-MPa confining pressure. It contracts ($\bar{\alpha}$ is negative) between 25^0 and 100^0 C; it expands less than its wet counterpart between 100^0 and 300^0 C; and the expansions of dry and wet specimens are similar between 300^0 and 800^0 C.

Table 1. *Mean Coefficients of Linear Thermal Expansion* (10^{-6}/°C)

	Temperature Interval (°C)							
P_e (MPa)	25-100	25-200	25-300	25-400	25-500	25-600	25-700	25-800
				Andesite				
5	$17.3^{+7.4}_{-6.0}$	$10.3^{+6.8}_{-0.3}$	$15.6^{+2.0}_{-1.1}$	$14.9^{+0.4}_{-0.8}$	$13.1^{+2.4}_{-1.0}$	$11.7^{+3.1}_{-1.6}$	$10.5^{+2.8}_{-0.7}$	$10.1^{+0.9}_{-0.3}$
50	$11.7^{+1.6}_{-1.7}$	$8.6^{+2.8}_{-1.5}$	$9.6^{+2.0}_{-1.8}$	$10.9^{+2.7}_{-2.1}$	$10.3^{+1.6}_{-1.5}$	$11.1^{+1.1}_{-2.0}$		
100	$0^{+12.0}_{-5.3}$	$3.4^{+6.3}_{-4.0}$	$6.0^{+3.8}_{-3.1}$	$7.3^{+3.8}_{-2.4}$	$6.9^{+3.3}_{-2.2}$	$7.0^{+2.1}_{-2.7}$	$6.4^{+2.5}_{-2.3}$	
				Basalt				
5	$5.3^{+5.3}_{-5.3}$	$12.5^{+0.5}_{-0.5}$	10.0^{+0}_{-0}	$11.0^{+0.3}_{-0.3}$	10.9^{*}	10.8^{*}	11.4^{*}	11.2^{*}
50	6.0	6.0	5.5	7.2	9.0	8.7	9.4	8.8
				Granodiorite				
5	$8.0^{+1.3}_{-1.1}$	$14.9^{+2.5}_{-2.3}$	$12.0^{+1.1}_{-1.2}$	$12.8^{+0.7}_{-1.2}$	$13.5^{+0.7}_{-0.7}$	$15.3^{+0.5}_{-1.0}$	$17.2^{+0.5}_{-0.7}$	$16.0^{+0.6}_{-0.5}$
50	$2.7^{+4.1}_{-3.4}$	$4.6^{+2.8}_{-0.6}$	$8.0^{+0.5}_{-0.4}$	$10.0^{+0.4}_{-0.1}$	$11.6^{+0.8}_{-0.4}$	$13.0^{+0.4}_{-0.2}$	$14.4^{+0.6}_{-0.4}$	$14.2^{+0.2}_{-0.1}$
50 (dry)	-3.3	2.3	6.9	13.3	12.0	13.5	15.6	14.7
100	5.3^{+0}_{-0}	5.7^{+0}_{-0}	$6.5^{+1.9}_{-1.4}$	$8.0^{+1.3}_{-1.3}$	$9.2^{+0.5}_{-0.3}$	$10.6^{+1.2}_{-1.0}$	$12.3^{+0.3}_{-0.9}$	12.0^{+0}_{-0}

Table 2. *Incremental Coefficients of Linear Thermal Expansion* (10^{-6}/°C)

P_e (MPa)	Temperature Interval (°C)														
	25-100	100-150	150-200	200-250	250-300	300-350	350-400	400-450	450-500	500-550	550-600	600-650	650-700	700-750	750-800
								Andesite							
5	17.3	10.0	14.0	15.0	21.0	18.0	8.0	8.0	4.0	4.0	6.0	4.0	6.0	8.0	7.0
50	4.0	10.0	14.0	8.0	15.0	14.0	15.0	10.0	6.0	10.0	20.0	12.0			
100	1.3	1.0	9.0	8.0	13.0	12.0	10.0	5.0	6.0	6.0	8.0	4.0	2.0		
								Basalt							
5	5.3	21.5	14.5	6.0	5.0	17.6	9.9	11.5	10.0	12.0	8.0	10.0	20.0	10.0	10.0
50	6.0	6.0	6.0	8.2	5.2	12.6	11.0	0.09	23.0	0.07	0.09	12.0	11.0	7.0	2.0
								Granodiorite							
5	8.0	25.0	15.0	8.0	6.0	14.0	16.0	18.0	14.0	17.0	31.0	40.0	16.0	8.0	8.0
50	2.7	2.0	10.0	12.0	16.0	14.0	17.0	17.0	18.0	20.0	20.0	23.0	22.0	9.0	16.0
50 (dry)	-3.3	0	13.0	14.0	14.0	16.0	19.0	19.0	21.0	16.0	25.0	31.0	24.0	10.0	8.0
100	5.3	6.2	5.8	4.0	12.0	14.0	10.0	12.0	15.0	15.0	20.0	24.0	20.0	16.0	8.0

Let us compare our data with those of Heard (1980) and Page and Heard (1981) on a similar rock, quartz monzonite, over the interval of 100^0 to 300^0 C. Their values of expansion under 14-MPa confining pressure are bracketed by ours taken under 5 and 50 MPa (Fig. 4). The data of Heard and Page (1982) on the Westerly and Stripa granites are also consistent with ours at comparable temperatures and pressures. Richter and Simmons (1974) and Cooper and Simmons (1977) measured the expansions of six unconfined granites from 20^0 to 400^0 C. The average mean values for the six rocks range from 7.5 to $24.2 \times 10^{-6}/^0$C, as compared with 8.0 to 12.8 for our granodiorite under the lowest confining pressure of 5 MPa.

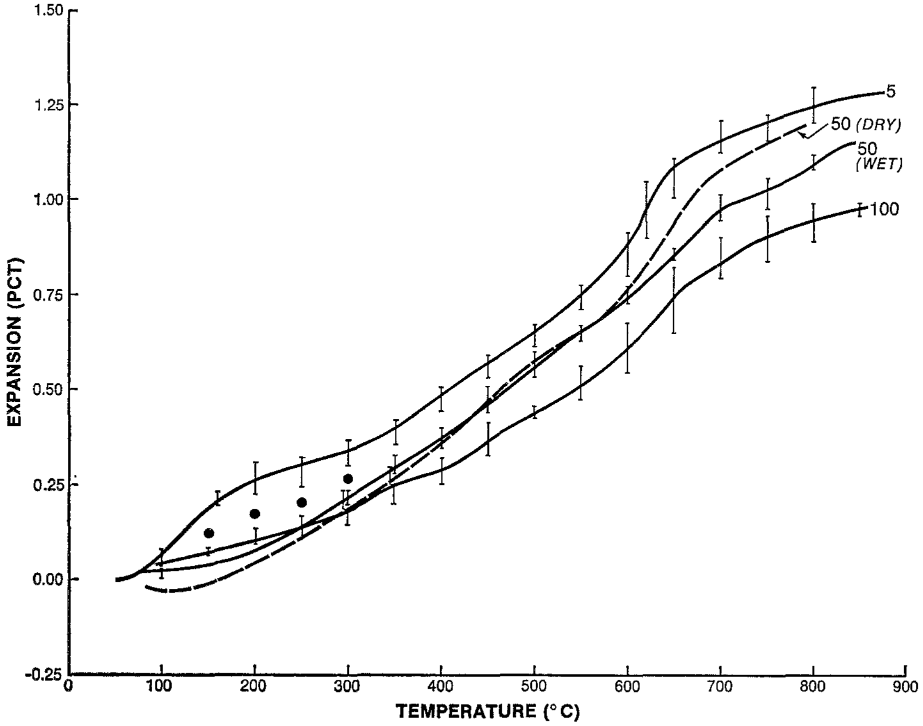

Fig. 4. Linear thermal expansion of Charcal Granodiorite under different effective confining pressures (MPa). The dashed curve is for a single dry specimen. Solid circles are Heard's (1980) data points

Discussion

Although we have not observed thermal cracking directly, we do know from previous petrofabric measurements on the unconfined granodiorite that microcrack density increases about tenfold upon heating from room temperature to 800^0 C (Friedman et al., 1979, Fig. 8; Bauer et al., 1979, Fig. 4). Here we infer crack porosity by subtracting the thermal expansions calculated for the crack-free rocks (Fig. 5) from those observed (Figs. 2, 3, 4).

Calculations

We have fitted the mean volume-expansion data of Skinner (1966, Table 6-1) on the constituent minerals of our specimens to third-order polynominals and taken the derivatives, the thermal-expansion coefficients, at intervals of 50° C. Following Cooper and Simmons (1977), Page and Heard (1981), and Heard and Page (1982), we have used the method of Turner (1946) to calculate the incremental coefficients of volume expansion of the rocks (see Kingery et al., 1976, p. 603—606). The assumptions are that (1) the several mineral phases are randomly oriented and distributed; (2) all volume expansions are isotropic; (3) the rocks are initially crack-free; (4) their structures are not disrupted by heating, that is the expansion of every grain is the same as that of the rock in bulk; (5) thermal-expansion coefficients are independent of pressure (Sweet, 1979), and compressibilities, as taken from Birch (1966, Table 7-10) are independent of temperature; (6) intergranular stresses, arising from mismatches of expansions across grain boundaries, must sum to zero in a circumscribed volume in static equilibrium, and (7) linear expansions are one-third of volume expansions.

Table 3. *Parameters in Equation (1)*

Mineral	Volume Fraction	K (10^{-6})	Coefficients of volume thermal expansion, $\alpha(10^{-6}/°C)$				
			20°	200°	400°	600°	800°
			Andesite				
Plagioclase	79.0	1.45	11.5	14.7	17.5	19.4	20.5
Olivine	2.6	0.79	28.7	30.6	32.8	35.2	37.8
Augite	13.2	1.02	18.5	21.7	25.3	28.9	32.5
Hornblende	2.6	1.15	21.8	26.1	29.5	31.3	31.7
Magnetite	2.6	0.55	24.9	35.5	43.9	48.9	50.4
Whole rock	---	----	12.8	16.1	19.1	21.2	22.5
			Basalt				
Plagioclase	69.4	1.45	11.5	14.7	17.5	19.4	20.5
Olivine	15.6	0.79	28.7	30.6	32.8	35.2	37.8
Augite	6.5	1.02	18.5	21.7	25.3	28.9	32.5
Magnetite	8.5	0.55	24.9	35.5	43.9	48.9	50.4
Whole rock	---	----	14.1	17.4	20.4	22.6	24.0
			Granodiorite				
α quartz	21.7	2.71	35.8	45.5	68.8	165.2	----
β quartz	21.7	1.78	----	----	----	----	-7.6
Orthoclase	26.8	1.15	0.4	14.6	26.7	34.0	36.3
Plagioclase	39.0	1.45	11.5	14.7	17.5	19.4	20.5
Hornblende	12.5	1.92	21.8	26.1	29.5	31.3	31.7
Whole rock	----	----	17.1	25.6	37.7	71.8	19.8

Turner's equation is

$$\alpha_v = \frac{\Sigma \alpha_i K_i V_i}{\Sigma K_i V_i} \tag{1}$$

where α_v is the coefficient of bulk volume-expansion, and α_i, K_i, and V_i are the coefficients of volume expansion, compressibility, and volume fraction of the i-th phase. These parameters for the major phases are listed in Table 3.

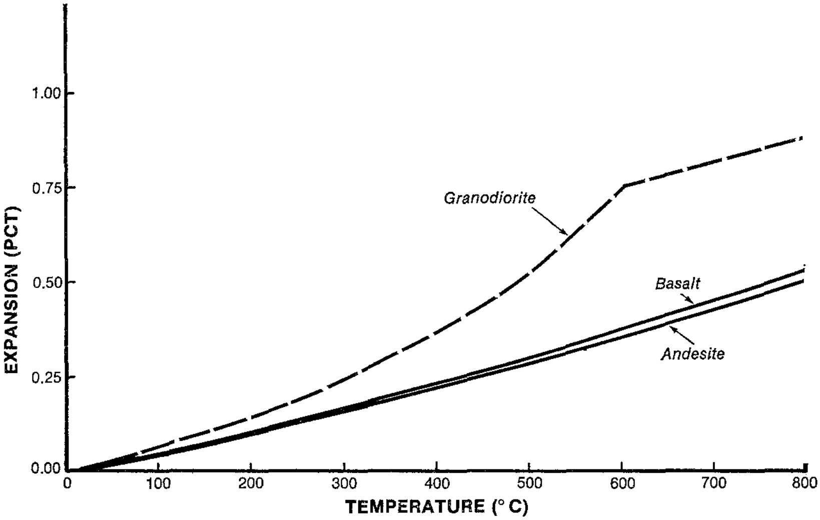

Fig. 5. Calculated linear thermal expansions of the crack-free rocks at atmospheric pressure

The minor phases, including micas, have been distributed proportionally amongst the major ones such that their fractions sum to 100 percent for the whole rock.

The calculated linear expansion is largest for the granodiorite as expected. It contains quartz, the thermal expansion of which is much larger than those of the other phases below the alpha-beta transition, assumed to occur at about 600° C in our confined specimens. The pronounced kink in the granodiorite curve reflects this transition. In single crystals of quartz the inversion is raised 26° C per 100 MPa of confining pressure (Coe and Paterson, 1969), but appears to be displaced 40° to 50° C upward in polycrystalline aggregates like granite (van der Molen, 1981). Specifying this transition temperature exactly is not essential to our conclusions.

Inferred Permeability-Changes

By assuming that the inelastic expansions, the differences between the calculated and observed values, are due solely to thermal cracking, we can infer the corresponding increase of connected-crack permeability, $k = e^2/12$, where e is the aperture (Brace et al., 1968; Gangi, 1978; Bauer and

Johnson, 1979). For a parallel array of planar cracks, $k = e^3/12\,b$, where b is a uniform spacing between cracks. The aperture, $e \propto \phi$, the linear crack porosity, the increase of which we have inferred. Accordingly,

$$\left(\frac{\phi_T}{\phi_0}\right)^3 \propto \frac{k_T}{k_0} \tag{2}$$

ϕ_0 and k_0 being measured at room temperature, and ϕ_T and k_T being the values at a particular temperature, T.

The largest changes occur in the granodiorite because matrix porosity is very low, and the intensity of thermal cracking is relatively high. At atmospheric pressure and room temperature, $\phi_0 = 3 \times 10^{-3}$. We tried to correct this value for the effect of confining pressure, but our data on compressibility are too scattered for meaningful calculations. Thus the relative changes of linear crack porosity between 25^0 and 800^0 C (Fig. 6), and our inferences about corresponding relative changes of connected-crack permeability (k) are minimum values because ϕ_0 is certainly somewhat lower than supposed.

A large change occurs at the lowest effective confining pressure (5 MPa) between 25^0 and about 150^0 C. Not much happens until the temperature reaches about 550^0 C where the relative crack porosity rises sharply. Above 600^0 C, k_T would exceed k_0 by nearly three orders of magnitude. At 50 MPa, there are small increases between about 400^0 and 600^0 C, followed by a dip, probably associated with the alpha-beta quartz inversion, and then a sharp rise. At 700^0 C, the inferred permeability is 100 times greater than that at room temperature. At 100 MPa, the change is still smaller, but heating to 700^0 C does increase the inferred permeability by a factor of more than ten.

These changes are somewhat larger than those reported by Heard (1980) and Page and Heard (1981) for quartz monzonite and Heard and Page (1982) for two granites in the range of 25^0 to 400^0 C under comparable confining pressures, but they are consistent with the direct permeability measurements of Summers et al. (1978).

Changes of permeability along with other mechanical and transport properties of the andesite and basalt would be much smaller because the intensity of thermal cracking is relatively low in these fine-grained rocks. Ceramicists have long recognized this grain-size effect (Kingery et al., 1976), and since intergranular stresses should be independent of grain size, Kuszyk and Bradt (1973) suggest that microcracks can develop only when sufficient strain energy is released to form new surface area. In a single grain, strain energy is a function of volume (L^3) whereas fracture surface energy is a function of area (L^2).

Effects of Pore Pressure

In a porous rock heated under confining pressure, processes of compaction compete with those of expansion. (High temperature favors plastic flow by lowering the strength and enhancing the ductility of constituent minerals. However, except in the micas, there is no petrographic evidence of plastic deformation in these rocks even at 800^0 C in specimens deformed to failure (Bauer et al., 1981)). We agree with van der Molen (1981) that the contraction in the low-temperature range of 25^0 to 100^0 C of the dry granodiorite is due to expansions of grains into available pore space, but that they are suppressed by the pore pressure in the wet rock. Once the pore space is filled, the dry rock begins to expand, and at temperatures above 150^0 C the thermal-expansion curve parallels that of the wet rock.

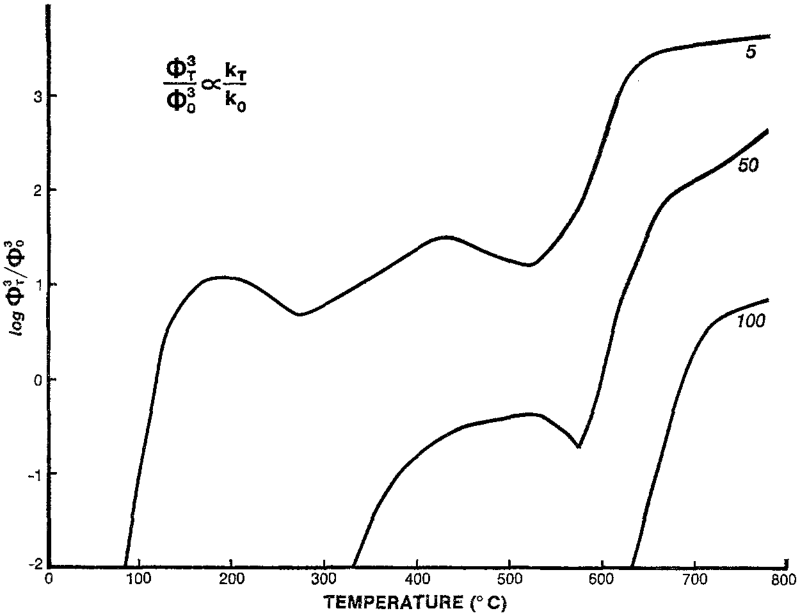

Fig. 6. Relative changes of linear crack density (ϕ) and permeability (k) with temperature of the Charcoal Granodiorite under different effective confining pressures (MPa)

Because the behaviors of the dry and wet granodiorite are essentially similar, and because our results agree well with those of other workers on comparable dry rocks at least over the interval from 25^0 to 400^0 C, we regard the effect of pore pressure as essentially mechanical and not chemical. It acts to reduce effective confining pressure and hence favor thermal cracking in water-saturated rocks over their dry counterparts under otherwise similar conditions of temperature and total confining pressure. There is no evidence, at least from our short-term tests, of time-dependent chemical effects like stress corrosion that would enhance thermal cracking.

Conclusions

The heating of anisotropic polycrystalline aggregates to temperatures much above 100^0 C under effective confining pressures associated with crustal depths of less than about 10 km may develop tensile stresses that exceed local cohesive strengths and hence cause thermal cracking. The creation of crack porosity, especially in a rock like granite with very low matrix porosity, can greatly modify many important mechanical and tranpsort properties, including strength (Bauer and Johnson, 1979; Friedman et al., 1979), elastic moduli (Brace, 1965; Walsh, 1965; Page and Heard, 1981; Heard and Page, 1982), sound velocity (Hadley, 1976; Kern, 1978; Johnson et al., 1978; Bauer and Johnson, 1979), thermal conductivity (Walsh and Decker, 1966; Durham and Abey, 1981), electrical conductivity (Brace and Orange, 1965), and especially permeability (Brace et al., 1968; Bauer and Johnson, 1979; Heard, 1980; Page and Heard, 1981; Heard and Page, 1982; Summers et al., 1978; Trimmer et al., 1980) as we have emphasized here.

In a fractured, water-saturated rock mass heated to about 200° C under effective confining pressure ≤ 100 MPa by, say, the emplacement of high-level radiogenic waste, the permeability of a convective system will depend strongly on the opening or closing of pre-existing fractures and the creation of new thermal cracks. The thermoelastic deformations of the rock mass are closely coupled to the flow of groundwater. Thermal cracking could lower conductivity and so affect the temperature distribution and predicted thermoelastic stresses which in turn deform the cracks or create new ones, and so influence the rate of flow of groundwater which in turn affects the local temperatures in the convection system.

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