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1. INTRODUCTION

The measurement of the thermal and hydraulic properties of geologic materials has received a lot of attention in the past 20 years as a result of the growing interest in the disposal of nuclear waste, underground storage (compressed natural gas, liquified natural gas (LNG), liquified petroleum gases (LPG), compressed air, oil or water), permafrost engineering, and geothermal energy. Knowledge of rock permeability is also very important when analyzing seepage and uplift below concrete dams and in the prediction of water problems in underground excavations.

2. THERMAL PROPERTIES OF INTACT ROCK

2.1 Thermal expansion

Most engineering materials when unrestrained expand when heated and contract when cooled. The strain associated with a 1 degree temperature change is called the *coefficient of thermal expansion* ". This coefficient has the dimension of 1/°K, 1/°C or 1/°F. Using the engineering mechanics sign convention for strain, the thermal strain can be expressed as follows

$$\epsilon = \alpha (T - T_o)$$
(1)

where T_o is a reference temperature and T is the current temperature T. If $T > T_o$ (heating), the thermal strain is positive and extension takes place. Conversely, if $T < T_o$, (cooling) the strain is negative and contraction occurs. Values for the coefficient of thermal expansion of several rock types are given in Table 1.

For most ideal engineering materials, thermal strains are reversible. Note that if the material is unrestrained, thermal strains are not accompanied by any stress. On the other hand, if the material is constrained, thermal stresses develop in the material when subject to temperature changes. For instance, under uniaxial loading and using the engineering mechanics sign convention for stress, the longitudinal stress is equal to

$$\sigma = -\alpha E (T - T_o)$$
(2)

where E is the Young's modulus of the material assuming isotropic behavior. In this example, the thermal stress develops without any longitudinal strain. The stress is compressive if $T > T_o$ (heating) and tensile if $T < T_o$ (cooling).

For rocks, thermal contraction and extension are not the same in all directions since rocks are aggregates of mineral grains with different degrees of thermal expansion. Because of anisotropic expansion of various mineral grains, stresses develop in the grains or at the grain interfaces.

Rock Type	α ($10^{-6}\,K^{-1}$)
Basalt	5.4
Limestone	2.5-20
Granite	7.5-9
Sandstone	10
Marble	5.4-7
Salt	40

Table 1. Coefficient of thermal expansion for different rock types (after Berest and Vouille, 1988).

Combination of normal and shear stresses will result in separation between the grains or across intergranular cracks, i.e. thermal cracking.

Several devices have been used to measure the coefficient of thermal expansion of rocks (Richter and Simmons, 1974; Van Buskirk et al., 1985; ASTM D5335-92). Experiments conducted by Richter and Simmons (1974) on igneous rocks at temperatures ranging between 25 and 550°C have shown that thermal expansion depends on various factors such as heating rates, previous maximum temperatures, mineralogical composition, the initial porosity of the rock, and crystal orientation. For instance, they found that at low heating rates less than 2°C/min and temperatures less than 350°C, the expansion curves (ϵ, T) were reproducible and showed little hysterisis. They also found that the coefficient of thermal expansion, α, of a rock could be determined from its mineral composition and the coefficients of thermal expansion of the various mineral grains using the following equation

$$\alpha = \frac{\sum_{i=1}^{N} \alpha_i . K_i . V_i}{\sum_{i=1}^{N} K_i . V_i} \tag{3}$$

In equation (3), α_i, K_i and V_i are the coefficient of thermal expansion, bulk modulus and volume fraction of the ith mineral (i=1-N) in the rock. For heating rates larger than 2°C/min and temperatures larger than 350°C, Richter and Simmons (1974) found irrecoverable microstructural damage in the rock and a reduction in the value of its coefficient of thermal expansion. Figure 1 shows the effect of repeated thermal cycling on the thermal coefficient of two specimens of Gabbro. Permanent strain increases with the number of cycles of heating and cooling indicating an increase in thermal cracking in the rock.

Homand-Etienne and Houpert (1988) showed that thermal expansion of a rock depends on its grain size. For instance, they found that a granite with big crystals was more sensitive to temperature than a granite with small crystals. The same was found to apply with the size of the grains of limestone and sandstone. They also found that rocks with high initial porosity were less sensitive to temperature.

2.2 Thermal Conductivity

Most rocks obey *Fourier's law* relating the heat flux vector $\mathbf{q}$ (J/m/s^2) to the thermal gradient **grad**T (or **L**T) as follows

$$\mathbf{q} = -K_t . \nabla T \tag{4}$$

where K_t is the so-called *thermal conductivity* and is expressed in W/m/K or J/m/K/s. It represents the ability of a material to transport thermal energy. Also important for transient heat conduction problems in rock is the so-called *specific heat* C_p (in J/kg/K) which is the capacity of a material to store thermal energy and the *thermal diffusivity* k_t (in m^2/s) which is the ability of a material to level temperature differences. These three thermal properties are related as follows

$$k_t = \frac{K_t}{\rho C_p} \tag{5}$$

where D is the density (kg/m^3). For most rocks, K_t varies between 0.5 and 4.2 W/m/K, and C_p varies between 500 and 1000 J/kg/K (Berest, 1988). Values of K_t and k_t for an "average" rock, air, water and steel (0.5%C) are compared in Table 2.

	Rock	Air	Water	Steel
k_t (m^2/s)	$1.5 \ 10^{-6}$	$22 \ 10^{-6}$	$1.45 \ 10^{-6}$	$1.47 \ 10^{-5}$
K_t (W/m/K)	3	0.026	0.6	54

Table 2. Values of K_t and k_t for rock and other media (after Berest, 1988).

The thermal conductivity of geologic materials is usually low requiring sensitive measuring systems with high heat input and temperature gradient (Ashworth and Ashworth, 1990). An overview of the different methods to measure the thermal conductivity of rocks can be found in Ashworth et al. (1985). ASTM standards include D4611-86 (specific heat), D4612-86 (thermal diffusivity), and D5334-92 (thermal conductivity).

Thermal conductivity can be measured using the *transient line heat source method* as proposed by Woodside and Messmer (1961). A linear heat source is placed in a hole drilled along the longitudinal axis of a rock core. Following an increase in temperature, the power dissipated in the test sample is measured. The testing apparatus is shown in Figure 2 (Van Buskirk et al., 1985). In this device, a 0.32 cm-diameter heater was used and the specimen exterior is kept at a constant pressure and temperature. The thermal conductivity K_t is equal to

$$K_t = \frac{P}{4\pi\Delta\theta}\ln\frac{t_2}{t_1}$$
(6)

where P is the power input or dissipated in the test sample per unit length (W/m or J/m/s),)2 is the temperature rise in the line source between times t_1 and t_2. Van Buskirk et al. (1985) have shown experimentally that the thermal conductivity of rocks depends on their composition, anisotropy, grain size, pore size, saturation including the type of saturating fluid, specimen density and porosity, confining and pore fluid pressures conditions, and temperatures. Conductivity was found to decrease with increasing porosity and increasing temperature. Thermal conductivity has also been found to be sensitive to moisture content (Ashworth and Ashworth, 1990).

2.3 Effect of Temperature on Rock Properties

The following trends are illustrated by Figures 3a-3h taken from the paper by Homand-Etienne and Houpert (1988):

1. porosity increases with temperature (Fig. 3a);
2. rock compressibility increases with temperature (Fig. 3b, 3c);
3. sonic velocity decreases as temperature increases (Fig. 3d);
4. rock deformability increases with temperature (Fig. 3e, 3g);
5. rock strength decreases as temperature increases (Fig. 3f, 3g);
6. rock creep increases with temperature (Fig. 3h).

3. HYDRAULIC PROPERTIES OF INTACT ROCK

Permeability can be used to assess the interconnection between the pore phase of an intact rock and its degree of rock fissuring. Most intact rocks obey *Darcy's law* relating the discharge velocity, v, to the hydraulic gradient, i, as follows

$$v = k.i$$
(7)

In equation (7), k is the *coefficient of permeability* and has the dimension of a velocity (L/T). The permeability k can also be represented by the equation

$$k = K\left(\frac{\gamma_w}{\nu}\right) \qquad\qquad (8)$$

where γ_w and ν are the unit weight and dynamic viscosity of water (M/LT), respectively. For water at 20°C, γ_w=62.4 lf/ft^3 (9.80 kN/m^3) and ν=2.098 x 10^{-5} lb.s/ft^2 (1.005 x 10^{-3} N.s/m^2). In equation (8), K is the so-called *absolute* or *specific permeability* and has the dimensions of an area (L^2). Sometimes K is expressed in Darcies where 1 Darcy is equal to 9.87x10^{-9} cm^2. For water at 20°C, 1 Darcy corresponds to a permeability of about 10^{-3} cm/s. In writing equation (8), the coefficient of permeability is divided into two parts, i.e. the absolute permeability K which depends on the properties of the rock only, and (γ_w/ν) which accounts for the properties of the fluid only.

The permeability of intact rocks is usually small compared to the permeability of rock masses where most of the fluid flow takes place along preferred channels in fractures, joints, etc... This is clearly illustrated in Table 3 where values of the permeability coefficients have been reported for intact rocks and rock masses.

Rock Type	k (cm/s) for rock with water (20°C) as a permeant	
	Lab.	Field
Sandstone	3×10^{-3} - 8×10^{-8}	1×10^{-3} - 3×10^{-8}
Shale	10^{-9} - 5×10^{-13}	10^{-8} - 10^{-11}
Limestone, dolomite	10^{-5} - 10^{-13}	10^{-3} - 10^{-7}
Basalt	10^{-12}	10^{-2} - 10^{-7}
Granite	10^{-7} - 10^{-11}	10^{-4} - 10^{-9}
Schist	10^{-8}	2×10^{-7}

Table 3. Permeability of intact rocks and rock masses (adapted from Goodman, 1989).

One apparatus used to measure intact rock permeability was proposed by Bernaix (1966) for the testing of the gneiss at the site of the Malpasset dam. The apparatus called the *radial permeability test apparatus* is shown in Figure 4. The test is conducted on a hollow cylindrical rock specimen.

Let R_1 and R_2 be the inner and outer radii of the test specimen and p_1 (or head h_1) and p_2 (or head h_2) be the pressures applied on its inner and outer surfaces. Depending on the values of p_1 and p_2, convergent or divergent flow can be conducted. Assuming Darcy's law for the radial flow in the test specimen, the coefficient of permeability, k, can be expressed as follows

$$k = \frac{Q}{2\pi L \Delta h} \cdot \ln \frac{R_2}{R_1} \tag{9}$$

where Q is the flow rate passing through the rock sample and Δh =h_2-h_1 for convergent flow and h_1-h_2 for divergent flow. In equation (9), Δh can also be replaced by the difference in pressure Δp with Δp=(γ_w Δ)h. Convergent flow creates compressive-type body forces that tend to close rock microfissures whereas divergent flow creates tensile-type body forces that tend to open the microfissures. In both cases, the seepage force per unit volume is equal to

$$\frac{dp}{dr} = \frac{(P_2 - P_1)}{\ln(R_2/R_1)} \cdot \frac{1}{r} \tag{10}$$

Comparison between the values of k measured during convergent and divergent flows gives an indication of the stress dependency of the rock material and therefore of its degree of fissuring. Bernaix (1966) and Habib and Bernaix (1966) proposed an *index of fissuring*, S, equal to

$$S = \frac{k(-0.1\ MPa)}{k(5\ MPa)} \tag{11}$$

where k(-0.1 MPa) is the radial permeability measured under radial flow with a pressure difference p_2 - p_1 = 0 - 0.1 MPa = -0.1 MPa and k (5 MPa) is the radial permeability measured under radial flow with a pressure difference p_2 - p_1= 5 MPa - 0 = 5 MPa. The following classification was proposed:

- slightly fissured rock   S = 3 - 10;
- rock of average quality   S = 10 - 20;
- extremely fissured rock   S > 20.

It was found that the gneiss at the Malpasset site had values of s in excess of 100 with some measurements well over 1000, and as large as 50,000 (Habib, 1987). Note that in deriving equation (9), the permeability coefficient is also assumed to be constant and independent of the stress level.

Another permeability apparatus was developed at the University of Colorado at Boulder for the testing of radial flow in intact rocks under axisymmetric loading (Sewell, 1979). The apparatus shown in Figure 5 allows the measurement of the permeability under a combination of axial and radial stresses.

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Problems (Due March 9, 2001)

1) Consider the geometry of the radial permeability test. Let p_1 and p_2 be the applied pressures on the inner and outer surfaces of the test specimen with inner radius R_1 and outer radius R_2. Let $\Delta p = p_2 - p_1$.

a) What is the expression for the pressure p at any distance r from the center of the test specimen. Express $(p-p_1)/\Delta p$ in terms of r/R_1 and R_2/R_1.
b) What is the expression of the seepage body force per unit volume?
c) Write the equations of equilibrium in terms of total and effective stresses.

2) Consider a rock mass cut by a single joint set with average spacing, S, and average aperture b. The intact rock permeability is denoted as K_m. The joint permeability is equal to $K_j = gb^2/12\nu$ where g is the acceleration due to gravity (9.81 m/s^2 or 32.2 ft/s^2), b is the joint aperture, and ν is the kinematic viscosity of the fluid (for water it is equal to 1.3 x 10^{-6} m^2/s or 14 x 10^{-6} ft^2/s at 20° C).

Show that the rock mass permeability is anisotropic with $K_z = K_m$ and $K_2 \cong K_m + K_j b/S \cong K_j b/S$ for small values of S. Numerical example: b = 1 mm (0.04 in), $K_m = 10^{-5}$ cm/s and S varies between 0.1 m and infinity (intact rock).

Figure 1. Effect of repeated thermal loading on two specimens of gabbro. Crosses are heating data and circles are cooling data. Numbers indicate the number of the thermal cycle (after Richter and Simmons, 1974).

Figure 2. Thermal conductivity test apparatus (after Van Buskirk et al. ,1985).

Figure 3. Effect of temperature on various rock properties (after Homand-Etienne and Houpert, 1988).

(a) porosity vs. temperature
(b) pressure vs. volumetric strain for granite
(c) pressure vs. volumetric strain for sandstone
(d) sonic velocity vs. temperature
(e) Young's modulus vs. temperature for granitic rocks
(f) uniaxial compressive strength vs. temperature for granitic rocks
(g) stress-strain curves vs. temperature for granite
(h) creep curves for granite at room temperature (top) and at 600°C (bottom).

Figure 4. Radial permeability apparatus of Bernaix (after Habib, 1987).

Figure 5. Radial permeability apparatus of Sewell (1979).