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Pressure Effects on Thermal Conductivity and Expansion of Geologic Materials

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Sandia Laboratories
energy report

SAND-75-1991

Pressure Effects on Thermal Conductivity
and Expansion of Geologic Materials

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Abstract

The generally held view existing that the thermal conductivity of rock samples were on both a theoretical and experimental basis, materials which are homogeneous and isotropic, and free of interfaces. In the case of heterogeneous materials, only small shifts are expected. Depth of view of these shifts will be indicated and compared to the available temperature variations of the past. In the region of increase, the thermal conductivity of a sample with pressure is decreased, whereas, in the region of decrease, the thermal conductivity is increased. The increasing tendency of expansion, which is observed in the temperature variation of the isobaric thermal conductivity, is a weakly decreasing tendency in the pressure variation of the isobaric thermal conductivity. The present study is based on a theoretical and experimental study of the thermal conductivity of a medium with a constant density increase, as the pressure decreases, with the variation of the temperature being dependent on the thermal conductivity of the medium. A detailed analysis of existing data on the conductivity of geologic materials is presented. In the case of the thermal conductivity of a medium, the weighted average of the data is used. The present study is based on the conductivity of a medium with a constant density increase, as the pressure decreases, with the variation of the temperature being dependent on the thermal conductivity of the medium. A detailed analysis of existing data on the conductivity of geologic materials is presented. In the case of the thermal conductivity of a medium, the weighted average of the data is used.

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*The present work was conducted as part of the Academic Basic Research Project.

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1. Introduction

In order to make an accurate assessment of the thermal response properties of a nuclear waste repository, accurate values of the thermal constants characterizing the geologic formations in and near the repository are required. This paper is devoted to a discussion of the effects of depth or pressure on the thermal expansion and thermal conductivity of geologic materials. This assessment was motivated by a desire to define those laboratory experiments which could most profitably be performed to define the pressure dependence of these properties. Since conductivity and expansion are both functions of temperature and since temperature varies directly with depth, there will be a temperature-induced property variation with depth in addition to any explicit variation caused by the increase in pressure with depth.

A review of the literature on the pressure dependence of conductivity and expansion has been made, and from this review, we have concluded that in the case of homogeneous rock with little or no porosity, the changes in conductivity or expansion at depths up to 3 km is not significant. Throughout this paper the term "homogeneous rock" means rock which is free of porosity, micro-cracks and inclusions on a scale $\geq$ the average grain size and which possesses properties (density, thermal conductivity, etc.) which do not have a spatial variation over distances much larger than a typical linear grain dimension. However, the situation changes in the case of porous rock, and for porosities $\geq 10\%$, significant property variations can occur as pressure increases with depth.

The variation of porosity with depth thus becomes the most significant source of property variation.

In Section II it is shown that the variation of thermal expansion with pressure at constant temperature may be found from the variation of compressibility with temperature at constant pressure. Using tabulated compressibility data, an estimate is made of the expansion coefficient shift at a pressure of 100 MPa, corresponding roughly to a depth of 3.7 km. This variation ranges from $\leq 10\%$ for dense rock such as granite to $\sim 25\%$ for 5% porous dry sandstone. Since it is reasonable to expect that porosity will decrease as depth increases, these predictions are felt to be "worst case" estimates of pressure-induced shifts.

In Section III thermal conductivity data are reviewed and the pressure-induced shifts in dense materials such as rocksalt are found to be in the range $\leq 3\%$. This change is negligible since a temperature shift $\leq 10^\circ\text{C}$ can easily cause a change¹ in conductivity of this magnitude, and temperatures are rarely known to this accuracy. In the case of porous rock such as shale or sandstone, large changes of conductivity can occur as pressure increases as a result of porosity reductions and improved intergranular contact. Since conductivity always increases with pressure, the shift is in a safe direction insofar as a higher conductivity leads to improved heat removal capability of the medium in which a radioactive waste canister is buried. If moisture is present in the rock formation in the form of liquid water, then a conductivity decrease could occur if this water is vaporized from heat given off by a waste canister. Hence moisture content is also an important parameter

in any thermal response study. Theoretical correlation of conductivity data with measured porosity is in a primitive state. An empirical weighted geometric mean formula is recommended based on its success in explaining a variety of data on porous sandstone and tuff conductivities.

As a result of this study we recommend that more effort should be expended in acquiring porosity versus depth data. In addition, variations of porosity over a horizontal repository plane will also need to be determined if accurate thermal modeling is to be performed. Laboratory experiments under pressure may provide erroneous or inconclusive results unless the actual variation of porosity with pressure can be accurately replicated in the lab. The problem of how included gas or liquid might escape as the pressure increases needs to be addressed prior to the development of these experiments.

II. Effect of Pressure on Thermal Expansion of Geologic Materials

In general we can show that the coefficient of volumetric thermal expansion β of homogeneous materials decreases with increasing pressure. For typical geologic rock materials, the change in β with pressure is not large for pressures corresponding to the maximum depths proposed for waste repositories. We shall review the available data in order to quantify the effect and estimate the maximum possible decrease in β from elevated pressure.

The volume thermal expansion coefficient is defined by

$$\beta = \frac{1}{V} \left(\frac{\partial V}{\partial T} \right)_P = \left[\frac{\partial (\ln V)}{\partial T} \right]_P, \quad (1)$$

where V = volume, T = temperature and p = pressure. The compressibility κ is defined by

$$\kappa = -V \left(\frac{\partial V}{\partial p} \right)_T = - \left[\frac{\partial (\ln V)}{\partial p} \right]_T \quad (2)$$

κ is related to the bulk modulus B by

$$B = 1/\kappa \quad (3)$$

Using rules for calculating with mixed partial derivatives,³ it can be shown that,

$$B = \kappa \left(\frac{\partial p}{\partial T} \right)_V \quad (4)$$

Using Eqs. (1-4) it can be shown that the partial derivative of β with respect to pressure at constant temperature is

$$\left(\frac{\partial \beta}{\partial p} \right)_T = - \left(\frac{\partial \kappa}{\partial T} \right)_p \quad (5)$$

$$= \frac{1}{B^2} \left(\frac{\partial B}{\partial T} \right)_p \quad (6)$$

Since $(\partial B / \partial T)_p < 0$ for rock³ as for metals,⁴ we predict from Eq. (6) that the thermal expansion coefficient will decrease with increasing pressure.

The range of pressures to be encountered can be estimated from the standard pressure equation,

$$p = \rho gh \quad (7)$$

where ρ = earth density, g = gravitational acceleration and h = depth in the ground. The density of typical types of rock is $\leq 2.8 \text{ g/cm}^3 = 2.8 \times 10^3 \text{ kg/m}^3$ ⁵ and with $g = 9.8 \text{ m/s}^2$,

the rate of change of pressure with depth is $dp/dh \leq 2.7 \times 10^4$ Pa/m = 27 MPa/km. If we assume a maximum burial depth of ~ 3 km, the maximum pressure to be encountered is $p_{\max} \leq 81 \text{ MPa} = 1.18 \times 10^4 \text{ psi} = 810 \text{ bar}$. The maximum pressure used in the following calculations and estimates will be set at 100 MPa for convenience. Using data in Clark³ we can calculate the expected change in β at a pressure of 100 MPa. From Table 7-17 of Ref. 3, we find data on the compressibility (called β in Ref. 3) as a function of T and p for ten types of rock. This table is reproduced here for convenience as Table I.

As an example we use the data for Caplen Dome dry sandstone (entry 3) and Eq. (5) to find $(\partial\beta/\partial p)_T$ with the compressibility data from the last column, $(\partial\kappa/\partial T)_p \approx 10^{-5} (4.83-3.50)/200-25) = 7.6 \times 10^{-8} (\text{MPa-K})^{-1}$ at $p = 50 \text{ MPa}$. Thus, estimating $\Delta\beta$ from Eq. (5), $\Delta\beta \approx (\partial\beta/\partial p)_T \Delta p = -(\partial\kappa/\partial T)_p \Delta p \approx -7.6 \times 10^{-8} (\text{MPa-K})^{-1} \times 10^2 \text{ MPa} = -7.6 \times 10^{-6} \text{ K}^{-1}$. From Table 6-10 of Ref. 2 we find that the linear coefficient of expansion for sandstone is $\alpha = 10 \times 10^{-6} \text{ K}^{-1}$ and thus, assuming the sandstone is homogeneous, $\beta = 3\alpha = 30 \times 10^{-6} \text{ K}^{-1}$. Thus $\Delta\beta/\beta \sim -0.25$. In Table II the results of this type of calculation are presented for the various types of rock in Table I. As expected, the most porous rock (sandstone) exhibits the largest change in expansion coefficient while the granites typically show very small changes.

Another way to examine the effect of pressure on expansion coefficient is through the Murnaghan equation of state.^{6,7} Anderson⁷ showed that the expansion coefficient at high pressures

Table 1. Rock Elastic Constants as a Function of Pressure and Temperature. From
S. P. Clark, Jr., Ref. 3.

Rock Type/ Density (g/cm^3)	T ($^{\circ}\text{C}$)	p (MPa)	E (10^5 MPa)	G (10^5 MPa)	ν	κ ($1/10^5$ MPa)
Limestone, Solenhofen, Bavaria, $\rho = 2.565$	25	50.0	.633	.237	.336	1.56
		500.0	.655	.245	.337	1.49
	200	50.0	.605	.228	.327	1.72
		500.0	.623	.234	.332	1.62
Marble, Danby, VT/ $\rho = 2.698$	25	50.0	.705	.263	.341	1.35
		500.0	.762	.283	.346	1.21
	200	50.0	.586	.221	.325	1.79
		500.0	.705	.261	.350	1.28
Sandstone, Capler Dome, dry/ $\rho = 2.543$ porosity = 5.1%	25	50.0	.457	.185	.234	3.50
		500.0	.573	.224	.278	2.33
	200	50.0	.404	.172	.175	4.83
		500.0	.570	.224	.272	2.40
Granite, biotite, Woodbury, VT/ $\rho = 2.644$	25	50.0	.792	.315	.257	1.84
		500.0	.887	.357	.242	1.75
	200	50.0	.774	.307	.260	1.86
		500.0	.868	.351	.237	1.82
Granite (gray), Llano County, TX/ $\rho = 2.609$	25	50.0	.804	.328	.225	2.05
		500.0	.852	.340	.253	1.74
	200	50.0	.751	.304	.236	2.11
		500.0	.836	.337	.240	1.87
Granite (pink), Llano County, TX/ $\rho = 2.636$	25	50.0	.752	.288	.306	1.55
		500.0	.795	.302	.317	1.38
	200	50.0	.670	.258	.300	1.79
		500.0	.781	.298	.312	1.45
Basalt, hornblende, Chaffee County, CO/ $\rho = 2.586$	25	50.0	.675	.271	.246	2.26
		500.0	.708	.279	.269	1.96
	300	50.0	.637	.257	.240	2.45
		500.0	.697	.273	.276	1.93
Gabbro, "San Marcos," Escondido, CA/ $\rho = 2.993$	25	50.0	.960	.363	.322	1.11
		500.0	.998	.375	.331	1.02
	300	50.0	.883	.334	.322	1.21
		500.0	.983	.370	.329	1.04
Gabbro, bytownite, Duluth, MN/ $\rho = 2.885$	25	50.0	.901	.343	.314	1.24
		500.0	.953	.362	.317	1.15
	300	50.0	.781	.303	.289	1.62
		500.0	.901	.344	.309	1.27
Gabbro, hornblende/ $\rho = 2.933$	25	50.0	.981	.378	.297	1.24
		500.0	1.054	.407	.295	1.17
	300	50.0	.878	.345	.272	1.56
		500.0	1.037	.403	.287	1.23

E = Young's modulus

G = Shear modulus

ν = Poisson's ratio

κ = compressibility

can be approximated by

$$\beta(p,T) = \beta(0,T) \left\{ 1 + \frac{p}{B(0,T)} \left[\frac{\partial B(0,T)}{\partial p} \right] \right\}^{-1}, \quad (8)$$

where $B(0,T)$ is the bulk modulus at zero pressure. Using $\kappa = 1/B$, Eq. (8) can also be written

$$\beta(p,T) = \beta(0,T) \left\{ 1 - p \left[\frac{\partial (\ln \kappa)}{\partial p} \right] \right\}^{-1} \quad (9)$$

For the case of sandstone in Table I, $\partial(\ln \kappa)/\partial p = -1.55 \times 10^{-3} \text{ MPa}^{-1}$ at $T = 200^\circ\text{C}$. At $p = 100 \text{ MPa}$, Eq. (9) predicts $\beta(100,T)/\beta(0,T) = 0.87$, a 13% reduction as compared to the 25% reduction predicted in Table II. In view of the approximations implicit in Eq. (8),⁷ this is considered to be reasonable agreement. Equation (9) predicts that the change in β will be larger at higher temperatures. At 25°C , Eq. (9) yields only an 8% reduction in β at 100 MPa.

The results for rock presented above can be compared to that predicted for NaCl (halite). From Table 7-11 of Ref. 3 an estimate of $[\partial(\ln \kappa)/\partial p]_T \approx 10^{-4} \text{ MPa}^{-1}$ can be made. Then Eq. (9) predicts about a 1% reduction in β at $p = 100 \text{ MPa}$. Thus for halite the effect of pressure on expansion is predicted to be negligible.

We conclude this section on expansion by observing that for most types of rock, as well as halite, the predicted effect of a 100 MPa pressure on thermal expansion is to cause a decrease of 10% or less in this quantity. For porous rock such as sandstone, the effect will be larger, with a 25% reduction

Table II. Absolute and Relative Decrease in Volume Thermal Expansion Coefficient, $\Delta\beta$, from Atmospheric Pressure to 100 MPa for Various Types of Rock

Rock Type	$\beta(10^{-6}\text{K}^{-1})$	$-(\partial\kappa/\partial T)$	$-\Delta\beta(10^{-6}\text{K}^{-1})$	$-\Delta\beta/\beta$ (%)
		$(10^{-6}\text{K}^{-1}\text{MPa}^{-1})$ $p = 50 \text{ MPa}$	$\Delta p = 100 \text{ MPa}$	
1. Limestone, Solenhofen, Bavaria	24	9.14×10^{-3}	0.91	3.8
2. Marble, Danby, VT.	21	2.51×10^{-2}	2.51	10.5
3. Sandstone, Caplen Dome, dry (porosity = 5.1%)	30	7.60×10^{-2}	7.60	25.3
4. Granite, biotite, Woodbury, VT	24	1.14×10^{-3}	0.11	0.4
5. Granite (gray), Llano County, TX	24	3.43×10^{-3}	0.34	1.4
6. Granite (pink), Llano County, TX	24	1.37×10^{-2}	1.37	5.7
7. Basalt, hornblende Chaffee County, CO	16	1.09×10^{-2}	1.09	6.8
8. Gabbro, "San Marcos," Escondido, CA	16	5.71×10^{-3}	0.57	3.5
9. Gabbro, bytownite, Duluth, MN	16	2.17×10^{-2}	2.17	13.6
10. Gabbro, hornblende	16	1.83×10^{-2}	1.83	11.4

in expansion being calculated for the 5% porosity sandstone referenced in Table II. It would appear that the pressure dependence of thermal expansion is dominated by porosity effects, and thus a measurement of porosity for rock samples would probably be as useful as a measurement of the expansion coefficient under pressure.

III. Effect of Pressure on Thermal Conductivity of Geologic Material

A. Introduction

In an insulating solid, heat is transported by lattice vibrations (phonons). Resistance to heat transport arises from various types of phonon scattering in the solid. The principal phonon scattering interactions are:⁸

1. Scattering of phonons by static imperfections (point defects, dislocations, etc.) and by boundaries (grain boundaries, external boundaries).
2. Interaction and scattering of phonons with each other. Phonon interaction is a fundamental consequence of the anharmonic nature of the atomic interaction potential in the solid.

We expect that both of these mechanisms will be important in a geologic material. The first will be of special importance in a highly inhomogeneous material with small grains or a large number of inclusions, pores, or microcracks. The second mechanism depends on basic properties of the solid and is not influenced by the degree of homogeneity.

In inhomogeneous materials, the relative concentration of the phases present and their distributions will have a marked effect on the effective thermal conductivity of the material. In particular, for rock, the major inhomogeneity present is porosity with either water or a gas filling the void spaces.⁹ Parkhomenko lists the following ranges of porosities for various types of rocks.

Table III. Porosity of Various Rock Types.
From Parkhomenko, Ref. 9.

Rock Type	Porosity (%)
Shale	20-50
Sand	20-35
Sandstone	5-30
Limestone	1.5-15
Dolomite	3-20
Quartzite	~ 0-0.8
Granite	0-1.2
Diabase	0-1.0
Biotite Granite	0.4-5.2
Nepheline Syenite	0.7-5.0
Diorite	1.3-5.1
Gabbro	0.3-3.5
Norite Gabbro	1.3-2.0

Porosity tends to decrease for rocks under pressure, and at a pressure ~ 100 MPa, all gas-filled porosity is believed to disappear.⁹ Since gases tend to have low thermal conductivities, porous rock with gas-filled voids will have a conductivity which is significantly lower than its homogeneous or water-filled counterpart.

In the next section the effects of pressure on the conductivity of homogeneous, porosity-free rock will be considered. This effect will be seen to be small for all pressures of interest. In the following section we shall review some data for inhomogeneous porous rock and discuss several conductivity models which have been developed to explain these data. The amount of porosity and the way in which this porosity decreases with increasing pressure will be seen to have a dominant effect on the variation of conductivity with pressure.

B. Variation of Conductivity with Pressure for Homogeneous Rock

We would expect that the pressure dependence of the conductivity of relatively homogeneous samples would be small because very large pressures are required to change the interatomic spacing sufficiently to change the form of the atomic potential. Some data obtained by Bridgman support this conjecture.¹⁰ Bridgman fit his conductivity versus pressure data to a linear equation of the form

$$\lambda(p) = \lambda_0(1 + \gamma p) \quad , \quad (10)$$

where λ_0 = thermal conductivity at atmospheric pressure and γ is the coefficient of the linear pressure term. The data obtained by Bridgman as presented by Clark¹¹ are shown in Table IV, with the percent change in λ calculated from Eq. (10) shown in the last column.

Table IV. Effect of Hydrostatic Pressure on Thermal Conductivity of Various Nonmetallic Solids

	$\lambda(0)$ (W/m-K)	$\gamma(10^{-6}\text{MPa}^{-1})$	$\Delta\lambda/\lambda_0(\%)$ at 100 MPa ($\Delta\lambda/\lambda_0 = \gamma p$ at 100 MPa)
Pyrex Glass, 30° & 75°C	10.91	38.8	0.39
Basalt, 30°C	~1.4	50.0	0.50
Basalt, 75°C		22.4	0.22
Limestone, 30°C	2.19	10.2	0.10
(Solenhofen), 75°C	1.89	68.4	0.68
Rocksalt, 30° & 75°C	~5-6	367.3	3.67
Pipestone, 30°C	1.8	306.1	3.06

From the results shown in Table IV it is evident that the effect of pressure on the homogeneous thermal conductivity is small at pressures below 100 MPa.

Measurements of the pressure dependence of the thermal diffusivity for teflon, halite (NaCl), quartz, and silica have recently been reported by Kieffer, Gettings and Kennedy.¹² The thermal diffusivity κ is defined by the relation $\kappa = \lambda/\rho c_p$, where ρ = mass density and c_p = specific heat at constant pressure. The diffusivity is frequently the required parameter in thermal calculations, and many experimental determinations of λ are found from a measurement of κ . The κ versus p data can be fit to a linear relation of the form

$$\kappa(p) = \kappa_0(1 + \epsilon p) \quad , \quad (11)$$

similar to Eq. (10) for $\lambda(p)$. The values for κ_0 (cm^2/s) and ϵ (MPa^{-1}) are shown in Table V.

Table V. Experimental Values for κ_0 and ϵ (Eq. (11)) for Several Materials. From Ref. 12.

Material	κ_0 (cm^2/s)	ϵ (MPa^{-1})	$\kappa(100 \text{ MPa})/\kappa_0$
Teflon	0.0012	3.0×10^{-4}	1.03
Halite (NaCl)	0.031	3.1×10^{-4}	1.03
Quartz (l to c axis)	0.031	1.7×10^{-4}	1.02

Using published data for the specific heat and density of quartz, the calculated fractional change in conductivity for quartz is about 0.8%, a number comparable to those in the last column of Table IV.

The conclusion reached from this review is that the pressure dependence of the conductivity of nonporous rocks or minerals is not significant at pressures $\lesssim 100$ MPa. The variation of conductivity with temperature is a much more pronounced effect. In Figure 1 experimental values of conductivity versus temperature are shown for several rocksalt samples obtained from the site of the proposed Waste Isolation Pilot Plant near Carlsbad, NM. In the range 300-400K, the rate of change of conductivity with

temperature is $\sim 0.3\%/K$, and thus a 10K variation in temperature will change the conductivity about 3%. Since temperatures can easily vary within a 10K band, we feel that a $\sim 3\%$ variation in thermal conductivity is not significant for waste repository analysis. In addition the scatter of data shown in Figure 1 for samples from different elevations and different bore holes is much greater than the probable pressure-induced variation in conductivity. We now turn to a review and analysis of the conductivity of inhomogeneous porous rock which can exhibit a significant pressure dependence in some circumstances.

C. Thermal Conductivity of Inhomogeneous Rock

The theory of heat conduction in inhomogeneous media is naturally much more complicated than the homogeneous theory and, as a result, a number of simplified models have been advanced to explain various experimental observations. General discussions of these models have been given in recent reviews by Pratt¹³ and Parrott and Stuckes.¹⁴ Experiments have been conducted by Woodside and Messmer on the thermal conductivity of both unconsolidated sand¹⁵ and porous consolidated sandstone.¹⁶ They tested the various theories against their data and discussed effects of static pressure and gas overpressure on conductivity. Brailsford and Major have also compared experimental sandstone conductivity data to a theoretical relationship.¹⁷ Walsh and Decker have related the thermal conductivity of low porosity granites to the measured porosity and interpreted their results with a theory involving an effective or "crack" porosity derived from strain versus pressure data.¹⁸ In the remainder of this section we shall review the theories which have been applied to sandstone data and examine what pressure dependences of conductivity are predicted by each.

One of the simplest models of porosity was borrowed from the work of Maxwell in electromagnetic theory.¹⁹ In this analysis, Maxwell considered a dilute solution of spherical dielectric particles imbedded in a continuous host material

with a different dielectric constant and derived an electrostatic expression for the dielectric constant of the composite material. Brailsford and Major have reviewed the Maxwell derivation and applied it to the case of thermal conduction.¹⁷ The same theory is frequently applied to the analysis of solar selective absorbers consisting of small metallic particles embedded in a dielectric matrix. In this case the theory is called the Maxwell Garnett theory after J. C. Maxwell Garnett,²⁰ who applied the Maxwell wave theory to composite media. The major assumption in the Maxwell model is that the conductivity of each spherical inclusion is uniform but that the inclusions are widely separated so that the temperature distribution in the vicinity of each inclusion is not affected by the presence of other inclusions. We shall consider a system consisting of a continuous rock host material with a conductivity λ_r with inclusions containing a gas of conductivity λ_g . The result of the Maxwell theory for the effective conductivity λ_{er} of the composite with a continuous rock phase is then given by

$$\lambda_{er} = \lambda_r \left[1 - 2f \left(\frac{1-x}{2+x} \right) \right] / \left[1 + f \left(\frac{1-x}{2+x} \right) \right] \quad (11)$$

where f = volume fraction of gas and $x = \lambda_g / \lambda_r$. An interesting observation concerning Eq. (11) is that it contains no details about the shape and distribution of voids other than their net volume fraction. Thus, although the theory is derived for spherical inclusions, it is usually applied to general systems

with arbitrarily shaped inclusions.¹⁸⁻²⁰ For a typical gas $\lambda_g \sim 1.3 \times 10^{-2}$ W/m-K at standard temperature and pressure²¹ while $\lambda_r \geq 2$ W/m-K.⁶ Thus $\lambda_g/\lambda_r \sim 10^{-2}$ and hence Eq. (11) may be approximated by treating x as a small quantity. The result is

$$\lambda_{er} \approx \frac{\lambda_r}{(1+f/2)} \left[(1-f) + \frac{3x}{2} \left(f + \frac{1-f}{1+f/2} \right) \right] \quad (12)$$

λ_{er}/λ_r is plotted in Fig. 2 (curve (1)) as a function of porosity for $0 < f < 0.4$. If it is assumed that the gas phase is continuous and the solid or rock phase is discontinuous, then the effective conductivity λ_{eg} can be deduced from Eq. (11) by interchanging λ_g and λ_r and replacing f by $1-f$. Porous rock with gas phase continuous is imagined to be composed of more or less discrete pieces of solid rock, each of which is completely surrounded by intergranular cracks. The result is

$$\lambda_{eg} = \lambda_r x [3-2f(1-x)] / [3x + f(1-x)] \quad (13)$$

This relation is plotted as curve (2) in Fig. 2. The assumption of a continuous gas phase yields a much more rapid decline of λ_e with increasing porosity.

In their experiments on porous sandstone, Woodside and Messmer found that the data could be fit quite well to a weighted geometric mean formula,¹⁶

$$\lambda_{em} = \lambda_g^f \lambda_r^{(1-f)}$$

$$= \lambda_r x^f \quad (14)$$

Equation (14) is plotted in Fig. 2 for $x = 10^{-2}$ (curve (3)). No theoretical basis was advanced for this model. It is evident that the decline of λ_e with increasing porosity in the geometric mean model is much greater than that predicted by the rock phase continuous Maxwell theory. Brailsford and Major¹⁷ fit the sandstone data of Sugawara and Yoshizawa²² to a modified Maxwell theory formula (Fig. 2, Curves (4), (5)). In this theory, the heterogeneous solid is imagined to be composed of both rock phase continuous and gas phase continuous Maxwell solids imbedded in a medium which has the effective conductivity of the medium. The Brailsford and Major model is defined by the equations,

$$\lambda_{e(BM)} = \lambda_r \left(\lambda_{er} / \lambda_r \right) \left(q + (q^2 + 9x)^{1/2} \right) / 4 \quad (15)$$

$$x = \lambda_{eg} / \lambda_{er}$$

$$q = 1 + (1-x)(1-3f_e)$$

Curve (4) in Fig. 2 is calculated with $f_e = f$; curve (5), which Brailsford and Major fit to the Sugawara and Yoshizawa data, is calculated with the assumption,

$$f_e = f[\alpha + (1-\alpha)f] \quad (16)$$

and $\alpha = 0.8$. Although Eq. (16) was proposed by Brailsford and Major as having general utility, it does not lead to a good fit of the Woodside and Messmer data for any physically reasonable value of α (a value of α for which $f_e \leq 1$ when $0 \leq f \leq 1$).

Figure 2. Relative effective conductivity λ_e/λ_r versus void volume fraction f as predicted by various theories: (1) Maxwell theory - solid phase continuous, λ_{er}/λ_r , Eq. (11). (2) Maxwell theory - liquid (or gas) phase continuous, λ_{eg}/λ_r , Eq. (13). (3) Geometric mean empirical model, λ_{em}/λ_r , Eq. (14). (4) Brailsford and Major effective medium model, $\lambda_{e(BM)}/\lambda_r$, Eq. (15). (5) Brailsford and Major modified model using Eq. (16) to calculate an effective void fraction. The value assumed for λ_g/λ_r in these calculations was 0.01.

Walsh and Decker have studied the effect of pressure and saturating fluid on the thermal conductivity of low porosity rock such as granite.¹⁸ The main thrust of their argument is that there are essentially two types of voids in low porosity rock; closed isolated voids and quasi-continuous intergranular cracks. The cracks are proposed to have the most effect on the thermal conductivity, and as the volume of these cracks decreases under pressure, the conductivity will increase. Walsh derives this effective void fraction due to cracks from isothermal strain data.²³ In Fig. 3 strain versus pressure data are shown for a sample of Casco granite.¹⁸ The rapid initial rise of the strain $\Delta V/V_0$ is believed to be caused by the closing down of the intergranular cracks. At higher pressures the strain becomes linear in an "intrinsic" region where all of the cracks are closed. In the Walsh analysis,²³ the porosity fraction due to cracks f_c is derived from the strain data by extrapolating the linear strain region back to $p = 0$ as shown in Fig. 3, where f_c = strain intercept of this linear extrapolation.

Since Walsh and Decker assume quasicontinuous cracks, they utilize the Maxwell fluid phase continuous relation, Eq. (13) but replace f by f_c . Using the strain data they derive the pressure-dependent porosity $f_c(p)$, as shown on Fig. 3, and with Eq. (13) calculate $\lambda_e(p)$ which is then compared to experiment. In the case of Casco granite with

overall porosity $f = 7 \times 10^{-3}$ and crack porosity $f_c = 4 \times 10^{-3}$, they find reasonable agreement between experiment and theory at pressures $p \geq 100$ MPa. A replotted curve of their experimental λ_e versus p data for an air-saturated sample is shown on Fig. 4 (curve (2)). In the case of water-saturated rock, they observe an overall higher conductivity with only $\sim 3\%$ change in conductivity at 100 MPa. Although this theory seems reasonably sound for low-porosity rock, it is not possible to say whether it also works for more porous rock such as sandstone. To the best of the author's knowledge, it has never been applied to other materials.

Figure 3. Strain versus pressure (solid line) for Casco granite (Ref. 18) showing construction used to determine crack porosity f_c . At $p = 0$, f_c is found by extrapolating the linear strain region to $p = 0$ (dashed line). At an intermediate pressure, p_1 , $f_c(p_1)$ is the distance between extrapolated and actual strain curves.

Figure 4. Relative effective conductivity λ_e/λ_r versus pressure from various models and experiments. (1) Geometric mean model with fluid rock phase, Eq. (17) for $f(p)$. (2) Data of Walsh and Decker¹⁸ for Casco granite under uniaxial stress. (3) Data of Woodside and Messmer¹⁶ for Berea sandstone air loaded. (4) Data of Woodside and Messmer for the same sample vacuum loaded. (5) Geometric mean model with "rigid" rock and $p_{\text{gas}} = p_{\text{rock}}$ and λ_g from Eq. (22).

In the general case, the variation of thermal conductivity with pressure for porous solids will depend on the way in which the porous regions collapse under pressure. At one extreme, the solid could behave as a dense liquid and transmit all external pressure to the void spaces which would collapse until the internal gas pressure in the voids equaled the external pressure. In this case the ideal gas law yields the following relation for the pressure-dependent porosity $f(p)$,

$$f(p) = f(p_0) p_0/p \quad , \quad (17)$$

where p_0 = atmospheric pressure (~ 0.1 MPa). With this relation and a model giving conductivity as a function of porosity, the dependence of conductivity on pressure can be found. For the geometric mean model, an increase in pressure to 1 MPa will cause the conductivity of 20% porous rock to increase from $0.4 \lambda_r$ to $0.87 \lambda_r$. At $p = 10$ MPa the conductivity is practically saturated at λ_r . Since the conductivity of gases will increase with increasing pressure,²⁴ the increase of λ_e with pressure will be even more rapid than that predicted from Fig. 2, curve (3), and Eq. (17). The variation of λ_{em} with p is shown as curve (1) in Fig. 4.

At the other extreme, the solid could behave as a completely rigid body, with little change in void space dimensions as pressure is applied. Since the rock is presumed to be deep in the ground, the gas will be assumed to be at the same pressure as the solid. Then, all variation of λ_e with p will be caused by the variation of λ_g with p since λ_r is practically constant. The thermal conductivity of a gas is given approximately by⁸

$$\lambda_g \approx \frac{1}{3} c_v \bar{v} l \quad , \quad (18)$$

where c_v = specific heat, $\bar{v}$ = average molecular velocity, and ℓ = mean free path. For a gas in a pore with a diameter d , the mean free path is related to its value ℓ_0 in an infinite gas by

$$1/\ell = 1/\ell_0 + 1/d \quad . \quad (19)$$

Then, designating the infinite gas conductivity by λ_{g0} , the expression for $\lambda_g(p)$ becomes

$$\lambda_g(p) = \lambda_{g0}(p)d/(\ell_0+d) \quad . \quad (20)$$

ℓ_0 is given by²⁴ $\ell_0 = 1/\sqrt{2}n\sigma^2$, where n = number of gas molecules per cm^3 and σ = scattering cross section. Using the ideal gas law, $p = nk_B T$, where k_B = Boltzmann constant,

$$\lambda_g(p) = \lambda_{g0}(p) \frac{pd}{pd+\epsilon} \quad , \quad (21)$$

where $\epsilon = k_B T/\sqrt{2}n\sigma^2$. When d is large $\lambda_g(p) \rightarrow \lambda_{g0}(p)$ and when d is small $\lambda_g(p) \rightarrow \lambda_{g0}(p) pd/\epsilon$. Since $pd/\epsilon = d/\ell_0$, and $d \ll \ell_0$ when Eq. (17) is required, we can conclude that $\lambda_g(p) \lesssim \lambda_{g0}(p)$ and hence use of λ_{g0} will yield an upper limit for the gas conductivity.

Data tabulated by McLaughlin²⁴ show that the variation of λ_g with pressure for nitrogen is linear in the region $0 \leq p \leq 100$ MPa. A fit to the tabulated data shows

$$\lambda_{g0} = \lambda_0 + kp \quad , \quad (22)$$

where $\lambda_0 = 2.7 \times 10^{-2}$ W/m-K and $k = 6.81 \times 10^{-2}$ W/m-K-MPa. This form for λ_g can be used in any of the models previously described. Figure 4 (curve (5)) shows the results of a calculation with the geometric mean model. For these calculations, the initial porosity was taken to be 20%. The dashed line in Fig. 4 shows an experimental λ versus p relation reported by Woodside and Messmer for Berea sandstone with 22% measured porosity and saturated with air (curve (3)).¹⁶ In this experiment, the pressure was applied hydraulically to a rubber-coated sample whose interior was connected to a gas pressure source. The gas pressure was maintained at 1 atm (0.1 MPa) for this test. A similar curve for an evacuated sample subjected to hydraulic pressure is shown by the dash-dot line in Fig. 4 (curve (4)). The similarity between the two experimental curves was interpreted by Woodside and Messmer as evidence that the increase in conductivity with pressure is caused predominantly by improved intergranular contact at high pressure and not by an increase in gas conductivity. In another experiment

with a hydraulic pressure of 27.6 MPa in excess of the N_2 loading pressure, little change was observed in the conductivity up to an N_2 pressure of 6.9 MPa. The experiment was taken as confirmation of the prediction that the increase in gas conductivity with pressure will cause only a small change in the net conductivity.

With a liquid such as water in porous rock, the conductivity becomes much higher since $\lambda_{H_2O} \sim 25 \lambda_{air}$ and $\lambda_{H_2O} \geq 0.2 \lambda_{rock}$. In general, the Maxwell relations, Eq. (11) or Eq. (13), do not predict the observed conductivity when the measured porosity and known values of λ_{rock} and λ_{H_2O} are used in them. Liquids may only partially saturate porous rock such as sandstone which then must be considered as three-component systems of rock, liquid, and air. The presence of nearly incompressible liquid in rock will reduce the variation of conductivity with increasing pressure.

D. Discussion

From the above review it appears that the dominant factors in determining the pressure dependence of thermal conductivity are porosity and liquid content of the rock. A reliable model of heterogeneous conductivity does not exist, but the weighted geometric mean formula, Eq. (14), should provide an estimate of the conductivity if the porosity and conductivity of the nonporous rock are known. The change of conductivity with pressure is expected to increase as either the porosity increases or the liquid content decreases. The effect

of pressure on porous rock is to reduce the size of the pores and improve the intergranular contact. Both of these effects produce an increase in thermal conductivity. If the pores are free to collapse until the trapped gas pressure equals the external pressure, then Fig. 4 shows that the rise in conductivity will be extremely rapid.

In the case of pressure application, the conductivity of rock always rises and hence the surface temperature of a waste canister embedded in the rock would be expected to fall. This represents more effective cooling of the canister and a generally safer condition. A more important effect, and one which would shift conductivity in the opposite direction, would be the evaporation of included water in the vicinity of the canister. For example, if a canister is buried in water-saturated 20% porous shale with a rock conductivity of 1.62 W/m-K, then the net initial conductivity predicted by the geometric mean model, Eq. (14), using $\lambda_{H_2O} = 0.5$ W/m-K, is $\lambda_e = 1.33$ W/m-K. If the water is converted to a saturated vapor at 100K, the conductivity becomes $\lambda_{H_2O} \approx 0.025$ W/m-K and the effective conductivity changes to $\lambda_e \approx 0.7$ W/m-K, a reduction of the conductivity by about a factor of two. This conductivity reduction would result in decreased heat dissipation in the rock surrounding the canister with a consequent increase in canister surface temperature. This effect of water boil-out has probably been observed in experiments on Eleana argillite reported in the

FY77 Eleana Project Annual Report.²⁵ In these experiments, significant drops in conductivity were observed for samples held near 100°C for 24 hours.

E. Comparison with Recent Experiments

Recently, A. Lappin has reported measurements on the thermal conductivity of welded and nonwelded tuff from G Tunnel at the Nevada Test Site (NTS).²⁶ In order to compare the results of Lappin's experiment with the theory of Sec. III-C, we have selected his 500K data which should not have been greatly affected by the initial water in the rock. The measurements were performed at NTS by the transient heat flow technique.¹² The results are summarized in Table VI.

Table VI. Typical Thermal Conductivity and Porosity of Welded and Nonwelded Tuff from G Tunnel at NTS. From A. R. Lappin

Tuff Type	Porosity (%)	Measured Conductivity (W/m-K)
Welded	13	1.18
Nonwelded	44	0.41

The major problem in comparing these data to theory is that the zero porosity rock conductivity λ_r is unknown. Lappin has suggested using a λ_r for granite ≈ 4 W/m-K. Touloukian et al.

show λ_r data for NTS granite in the range 1.5-1.8 W/m-K in the temperature range 750-800K.²⁷ Clark gives values $\lambda_r \sim 2.3$ W/m-K in the temperature range 473-573K for various eastern locations.²⁸ We have elected, somewhat arbitrarily, to treat λ_r as an adjustable parameter and determine it by fitting the geometric mean model (Eq. (14)) to the welded tuff data in Table VI. The model then predicts the nonwelded effective conductivity which can be compared to the measured value in Table VI. The results are shown in Table VII.

Table VII. Results of Comparing the Data from Table VI to the Geometric Mean Model, Eq. (14), at 500°C.
 $\lambda_{air} = 0.056$ W/m-K.

Tuff Type	Porosity-f (%)	λ_r (W/m-K)	Measured λ_e (W/m-K)	Calculated λ_e (W/m-K)
Welded	13	1.86*	1.18	
Nonwelded	44	1.86*	0.41	0.40**

* Fit value of λ_r

** Computed value of λ_e for nonwelded tuff

The model evidently predicts a nonwelded tuff conductivity which is extremely close to the experimental value. The 1.86 W/m-K value of λ_r determined from the welded tuff data is in good agreement with the data reported by Touloukian but only about half of Lappin's ambient temperature assumed value.

The 25°C values of λ_e reported by Lappin are 1.62 and 0.61 W/m-K for welded and nonwelded tuff respectively. The extra conductivity at 25°C is probably due to the presence

of moisture. If we assume that the volume fraction of water is f_w and air f_g , then the natural extension of Eq. (14) to a three-component rock, water, air system is, with $f = f_g + f_w$

$$\lambda_e = \lambda_r (1-f) \lambda_g^{f_g} \lambda_w^{f_w} \quad (23)$$

If it is assumed that all of the water is driven off at 500°C so that $f_w = 0$ at that temperature, and that the variation of λ_r with T is small, then Eq. (23) may be used to derive a value for f_w . The ratio $\lambda_e(25^\circ\text{C})/\lambda_e(500^\circ\text{C})$ is given by

$$\frac{\lambda_e(25)}{\lambda_e(500)} = \lambda_g(25)^{(f-f_w)} \lambda_w^{f_w} / \lambda_g(500)^f \quad (24)$$

Solving for f_w yields

$$f_w = \frac{\ln \left[\frac{\lambda_e(25)}{\lambda_e(500)} \right] + f \ln \left[\frac{\lambda_g(500)}{\lambda_g(25)} \right]}{\ln \left[\frac{\lambda_w(25)}{\lambda_g(25)} \right]} \quad (25)$$

Using $\lambda_g(25) = 0.026$ W/m-K, $\lambda_g(500) = 0.056$ W/m-K, and $\lambda_w(25) = 0.605$ W/m-K, Eq. (25) predicts $f_w = 13\%$ and 23% for welded and nonwelded tuff respectively.

Lappin has determined values for densities, porosities, and water content of the two tuff types. He reports typical values as shown in Table VIII.

Table VIII. Nominal Values of Density, Porosity and Water Content for Welded and Nonwelded Tuff as Reported by Lappin

Tuff Type	ρ (gm/cm ³)	Porosity (%)	Water wt%	Water vol%*	f_w from Eq. (25) (%)
Welded	2.2	10	6	13.2	13
Nonwelded	1.9	35	18	34.2	23

* Water vol% = (water wt%) $\left[\rho / \rho_{H_2O} \right]$

The last column of Table VIII shows the prediction from Eq. (25) using the previous f values of 0.13 and 0.44 for welded and nonwelded tuff respectively. Thus, Lappin's experimental data indicates that both types of tuff are essentially saturated with water at 25°C while the geometric mean formalism leads to a lower predicted initial water content for the nonwelded tuff.

On the basis of this limited data, we conclude that the three-component geometric mean formula, Eq. (23), is in semiquantitative agreement with the tuff experimental data. λ_r is considered as a zero porosity rock conductivity which must be determined experimentally from high temperature

conductivity. The prediction of λ_e as a function of porosity for a two-component rock-air system appears quite good. In the case of three-component rock-air-water systems, Eq. (25) appears to yield reasonably good results. More precise data on conductivity as a function of water content and porosity are needed to better define the limits of validity of Eq. (23).

IV. Conclusions

The major conclusion of this investigation is that porosity plays a dominant role in determining the effect of pressure on thermal expansion and conductivity. As porosity increases, the rate of change of both these quantities with pressure increases. It is reasonable to expect that porosity will decrease with increasing depth, but without detailed knowledge of this variation it is not possible to quantitatively predict the depth or pressure dependence of the rock thermal properties. The calculated changes of thermal expansion in Table II show that dense rocks such as granite experience a decrease in expansion coefficient $\lesssim 5\%$ at a pressure of 100 MPa while porous rock types such as sandstone experience changes $\sim 25\%$. An estimate of the change of expansion under pressure can be made from data on compressibility (or bulk modulus) versus temperature through use of Eq. (5). In the case of water-bearing rock, any reduction in water content caused by boil-out will serve to increase the compressibility and consequently cause a

large shift in thermal expansion. Thus moisture content plays an important role in determining thermal expansion just as it does in determining conductivity.

The depth dependence of thermal conductivity can be estimated if the porosity, water content, and variation of these quantities with depth is known. If the porosity is gas filled, the weighted geometric mean formula, Eq. (14), is recommended for calculating the conductivity. A reasonable extension of Eq. (14) to rock with gas and water-filled porosity is given by Eq. (23). If possible, this formula should be verified by measuring the conductivity of rock with varying porosity and water content and then comparing the results to the predictions of Eq. (23). Since the various λ_e formulas discussed in Sec. II-B have not been extensively verified by comparing their predictions to a wide range of experimental results, it is not possible to use them with a high degree of confidence. The Woodside and Messmer experimental results and Lappin's tuff data support the validity of the weighted geometric mean formula. However, Eq. (14) and Eq. (23) have no theoretical foundation and thus there is no way to estimate what their limits of validity or applicability are.

The experimental problem of measuring the depth dependence of conductivity or expansion depends on how elastically the porosity collapses. If the collapse is completely inelastic, then the porosity will remain closed when a core sample is removed from a drill hole. In this

case, the functional dependence of conductivity and expansion with depth could be determined quite accurately from atmospheric pressure measurements on core samples. On the other hand, if the porosity collapse is partially elastic, then accurate properties determination will necessitate development of equipment capable of determining properties in the presence of an applied hydrostatic pressure.

The development of equipment capable of measuring properties under hydrostatic pressure is expected to be difficult. Since rock samples will usually contain water, the pressure apparatus would have to be gas-tight. For thermal expansion, the pressure dependence can be derived from measurements of the isothermal compressibility as a function of temperature at several fixed pressures in the range 0-100 MPa. For thermal conductivity measurement, the radial heat flow method is probably the most adaptable technique for use at high pressure, and it has been employed by Kieffer et al. at pressures up to 4×10^3 MPa with samples approximately 1.3 cm in diameter by 4 cm high.¹² For inhomogeneous rock specimens, larger sample sizes will be required in order to average over the inhomogeneities. The requirement for gas-tight sealing will make the task of running thermocouple leads into the cell a difficult one.

As a result of the projected difficulty and expense involved in measuring the pressure dependence of thermal conductivity and expansion for inhomogeneous rock samples,

it is recommended that effort be directed to developing a more fundamental understanding of the effect of moisture content and porosity on thermal properties and to understanding the way porosity collapses as depth and pressure increase. A specific recommended course of action for waste repository thermal properties studies is the following.

- I. From drill samples, measure the following quantities as a function of sample depth
 1. Moisture content
 2. Porosity
 3. Density
 4. Thermal expansion versus temperature at atmospheric pressure
 5. Thermal conductivity versus temperature at atmospheric pressure
- II. From these measurements correlate expansion and conductivity with porosity and attempt to fit analytical expressions to the experimental data.

This program should prove to be useful if the porosity remains collapsed after samples are removed from the earth and brought up to atmospheric pressure. If the collapse of porosity is partially elastic, then some of the collapsed porosity will be reestablished upon sample removal and thus

atmospheric pressure conductivity measurements will underestimate the true conductivity.

Another area which needs to be investigated is the dynamic behavior of porous water-bearing rock in proximity to a bore hole drilled in the ground with a heat source in the hole. In this case, moisture can escape up the hole and the porosity can then collapse further, leading to an eventual increase in the thermal conductivity and decrease in the expansion coefficient. In a situation such as this, laboratory experiments on sealed samples might well produce data which are not applicable to an actual waste disposal scenario.

References

1. R. D. Cheverton, W. D. Turner, "Thermal Analysis of the National Radioactive Waste Repository, Progress through June 1971," ORNL-4726, Union Carbide Corp. December 1971, pp. 39-40.
2. F. W. Sears, Thermodynamics, The Kinetic Theory of Gases, and Statistical Mechanics, 2nd Ed., Addison-Wesley, Reading, MA, 1956, pp. 30-40.
3. S. P. Clark, Jr., Handook of Physical Constants, Revised Ed., Geologic Society of America, Inc., 1966, p. 170, Table 7-17.
4. P. W. Bridgman, The Physics of High Pressure, G. Bell and Sons, London, 1958, p. 169.
5. S. P. Clark, Jr., Op. Cit. pp. 23-24, Table 4-4.
6. V. S. Touloukian et al., Thermophysical Properties of Matter - Thermal Conductivity of Nonmetallic Solids, TPRC Data Series, Vol. 13. Plenum, NY, 1970, pp. 13a-14a.
7. O. L. Anderson, J. Geophys. Research 72, 3661 (1967).
8. N. W. Ashcroft, N. D. Mermin, Solid State Physics, Holt, Rinehart and Winston, New York, 1976, pp. 495-505.
9. E. I. Parkhomenko, Electrical Properties of Rocks, Plenum Press, New York, 1967, pp. 7-9.
10. P. W. Bridgman, Op. Cit., p. 322.
11. S. P. Clark, Jr., Op. Cit., p. 482.

12. S. W. Kieffer, I. C. Getting, G. C. Kennedy, J. Geophys. Research, 81, 3018 (1976).
13. A. W. Pratt, "Heat Transmission in Low Conductivity Materials," in Thermal Conductivity, Vol. 1, edited by R. P. Tye, Academic Press, NY, 1969, pp. 301-405.
14. J. E. Parrott and A. D. Stuckes, Thermal Conductivity in Solids, Pion, Ltd., London, 1975, pp. 129-147.
15. W. Woodside and J. H. Messmer, J. Appl. Phys. 32, 1688 (1961).
16. W. Woodside and J. H. Messmer, J. Appl. Phys. 32, 1699 (1961).
17. A. D. Brailsford and K. G. Major, Brit. J. Appl. Phys. 15, 313 (1964).
18. J. B. Walsh and E. R. Decker, J. Geophys. Research 71, 3053 (1966).
19. J. C. Maxwell, A Treatise on Electricity and Magnetism, Clarendon Press, Oxford, 1904, 3rd Edition, Vol. 1, p. 440.
20. J. C. M. Garnett, Philos. Trans. R. Soc. Lond. 203, 385 (1904).
21. Y. S. Touloukian et al., Thermophysical Properties of Matter - Thermal Conductivity of Nonmetallic Liquids and Gases, TPRC Data Series, Vol. 3, Plenum, NY, 1970.
22. A. Sugawara and Y. Yoshizawa, Austral. J. Phys. 14, 469 (1961).
23. J. B. Walsh, J. Geophys. Research 70, 381 (1965).

24. E. McLaughlin, "Theory of Thermal Conductivity of Fluids,"
in R. P. Type, Op. Cit., Vol. I, pp. 29-33.
25. A. R. Lappin, J. F. Cuderman, "Eleana Project Annual
Report for FY-77," Unpublished Sandia Laboratories
Report.
26. A. R. Lappin, Personal communication and to be published.
27. Y. S. Touloukian, et al., Op. Cit., Vol. II, pp. 817-819.
28. S. P. Clark, Jr., Op. Cit., pp. 468-469, Table 21-4.

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