# 13. PHYSICAL PROPERTIES OF BASALTS, GABBROS, AND ULTRAMAFIC ROCKS FROM DSDP LEG 37[1]

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## ABSTRACT

The physical properties of bulk and grain density, porosity and water content, electrical resistivity, thermal conductivity, and compressional and shear wave velocity have been measured on a large collection of DSDP Leg 37 basalts and a few gabbros and serpentinized peridotites. The mean properties for the basalts are: bulk density 2.795 g/cm³; porosity, 7.8%; electrical resistivity at 25°C, 220 ohm-m; thermal conductivity, 3.97 mcal/cm/sec/°C; compressional velocity at 0.5 kbar pressure, 5.94 km/sec; and Poisson's ratio, 0.295. The means for the gabbros are: 2.957 g/cm³; 1.0%; 1710 ohm-m; 5.7 mcal/cm/sec/°C; 7.07 km/sec; 0.310. The means for the serpentinized peridotites are: 2.699 g/cm³; 0.8%; 7160 ohm-m; 7.6 mcal/cm/sec/°C; 6.18 km/sec; 0.350.

## INTRODUCTION

This article summarizes and discusses the physical properties measurements made on DSDP Leg 37 basalts, gabbros, and serpentinized peridotites, and presents some detailed data not given in other chapters. The properties examined are: bulk and grain density, porosity and water content, electrical resistivity, thermal conductivity, and compressional and shear wave velocity.

The shipboard measurements of density, porosity, electrical resistivity (at atmospheric temperature and pressure), and seismic velocities (at 0.5 kbar pressure) are given in the site reports, so only a summary and interpretation are given here. Measurements of electrical resistivity at higher temperatures and pressures are given here. The seismic velocity results at high pressures are presented and discussed in detail by Hyndman (this volume) and Christensen (this volume). The thermal conductivity at room temperature and low pressure by J. Jolivet for the computation of heat flow is described by Hyndman et al. (this volume). Very high pressure and temperature measurements of thermal conductivity and electrical resistivity of basalts are presented by Schloessin and Dvorak (this volume). The electrical properties of basalts as a function of frequency are presented by Katsube et al. (this volume).

The measured samples were mini cores 2.5 cm diameter × 2.5 to 5.0 cm length weighing about 50 g. It must be remembered that the core recovery in the basement parts of the Leg 37 holes was only about 20%. In addition, although we have attempted to include all rock types and states of weathering and alteration, only part of the recovered core is sufficiently unfractured to be suitable for measurement. Much of the drilled section not recovered probably is broken and fractured volcanic material with some sediment. Thus, the mean physical properties given may be a poor representation of the average values for the section drilled. This problem has been discussed by Hyndman (this volume).

Histograms showing the distribution of values for each physical property are given in Figure 1 for the basalts and the mean values and standard deviations for basalts, gabbros, and serpentinized peridotites are given in Table 1. The velocity and density results of Table 1 include data for four gabbros and two serpentinites from Christensen (this volume).

For previous general discussions of the physical properties of basalts, see Somerton et al. (1963) and Nafe and Drake (1968).

## BULK AND GRAIN DENSITY

The bulk density was determined onboard ship at atmospheric pressure and temperature. The samples (about 50 g) were weighed on a simple Ohaus Centogram four-beam scale with a capacity of 311 g and a resolution of 0.01 g. Calibration with standard weights during calm seas indicated an accuracy of ±0.02 g. The reproducibility also is about ±0.02 g. The volumes of the samples were determined from the differences in weights in air and suspended in distilled water with a thin wire. As a check, the volumes also were determined from the sample dimensions. The latter method has lower accuracy. The accuracy of the density determinations is about ±0.15% or ±0.005 g/cm³, which is adequate for most purposes. Higher measurement accuracy could be obtained with a better balance mounted on gimbals, but the problems of complete

[1]Contribution 599 of the Earth Physics Branch.

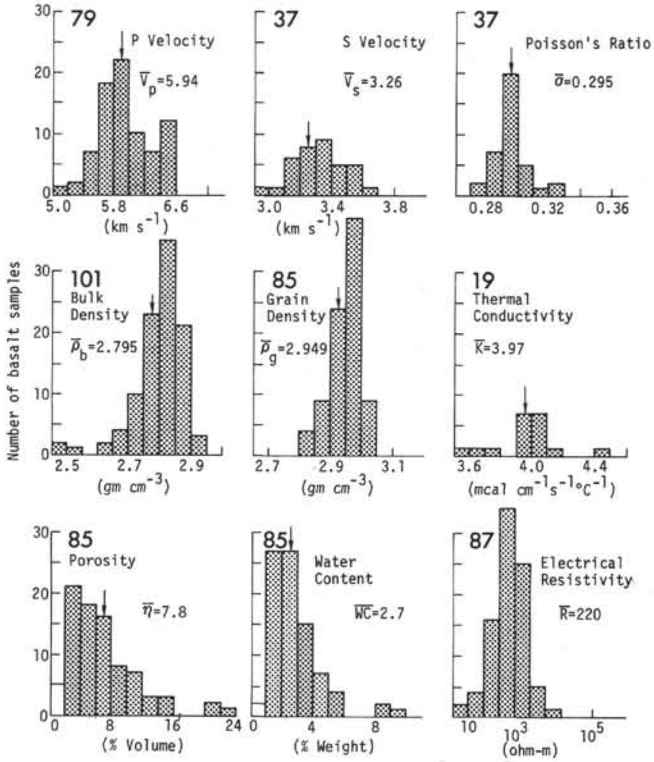


Figure 1. *Histograms showing frequency of occurrence of basalt samples for the physical properties of: bulk and grain density, porosity, water content, electrical resistivity, compressional and shear velocity. The mean value and number of samples are given for each property.*

water saturation and the amount of surface water will limit the real accuracy. The samples all were seawater saturated on the assumption that they were in this state on the sea floor. The water was wiped from the surface of the samples and the surface allowed to just dry (a few minutes) before weighing.

The grain or mineral density of the samples was estimated using the measured porosity (below). This involves simply subtracting the mass and the volume of the pore fluid and recomputing the density. Since the porosity must be underestimated, the computed grain density also will tend to be low. Thus, we estimate asymmetrical error limits of +0.014 to -0.007 g/cm³.

The bulk densities of 101 basalt samples (mainly from Site 332) range from 2.463 g/cm³ for highly vesicular material to 2.945 g/cm³ for low porosity material with a mean and standard deviation of individual values of 2.795 ±0.082 g/cm³ (Figure 1). Ten gabbros have a mean of 2.957 ±0.068 g/cm³ and 4 serpentinized peridotites 2.713 ±0.087 g/cm³. The grain or mineral density of the basalts ranges from 2.804 to 3.010 g/cm³ with a mean of 2.949 ±0.045 g/cm³. This range is considerably smaller than for the bulk density indicating that much of the variation in bulk density arises from variation in porosity rather than from differences in mineralogy. Allowing for the underestimation of the porosity, the true grain density should be close to 3.00 g/cm³.

The mean bulk density of 10 gabbros is 2.957 ±0.068 g/cm³ and the estimated mean grain density of 4

**TABLE 1**
**Physical Properties of DSDP Leg 37 Rocks**

| Property | Number of Samples | Mean | Standard Error of Mean | Standard Deviation Individual Values |
|---|---|---|---|---|
| **Basalts** | | | | |
| Comp. velocity (0.5 kbar) (km/sec) | 79 | 5.94 | 0.04 | 0.34 |
| Shear velocity (0.5 kbar) (km/sec) | 37 | 3.27 | 0.02 | 0.15 |
| Poisson's ratio | 37 | 0.295 | 0.002 | 0.011 |
| Bulk density (g/cm$^3$) | 101 | 2.795 | 0.008 | 0.082 |
| Grain density (g/cm$^3$) | 85 | 2.949 | 0.005 | 0.045 |
| Thermal cond. (mcal/cm/sec/°C) | 19 | 3.97 | 0.04 | 0.17 |
| Porosity (% vol) | 85 | 7.8 | 0.4 | 4.1 |
| Water content (% wt) | 85 | 2.7 | 0.2 | 1.5 |
| Elect. resist. (ohm-m) | 87 | 220[a] | | |
| **Gabbros** | | | | |
| Comp. velocity (0.5 kbar) (km/sec) | 10 | 7.07 | 0.05 | 0.15 |
| Shear velocity (0.5 kbar) (km/sec) | 10 | 3.72 | 0.03 | 0.11 |
| Poisson's ratio (0.5 kbar) | 10 | 0.310 | 0.004 | 0.011 |
| Bulk density (g/cm$^3$) | 10 | 2.957 | 0.021 | 0.068 |
| Grain density (g/cm$^3$) | 4 | 3.024 | 0.010 | 0.020 |
| Thermal cond. (mcal/cm/sec/°C) | 1 | 5.7 | | |
| Porosity (% vol) | 4 | 1.0 | 0.2 | 0.5 |
| Water content (% wt) | 4 | 0.37 | 0.07 | 0.15 |
| Elect. resist. (ohm-m) | 6 | 1710[a] | | |
| **Serpentinized Peridotites** | | | | |
| Comp. velocity (0.5 kbar) (km/sec) | 3 | 6.18 | 0.29 | 0.50 |
| Shear velocity (0.5 kbar) (km/sec) | 3 | 2.96 | 0.13 | 0.23 |
| Poisson's ratio (0.5 kbar) | 3 | 0.350 | 0.001 | 0.002 |
| Bulk density (g/cm$^3$) | 5 | 2.699 | 0.036 | 0.081 |
| Grain density (g/cm$^3$) | 1 | 2.868 | | |
| Thermal cond. (mcal/cm/sec/°C) | 1 | 7.6 | | |
| Porosity (% vol) | 1 | 0.8 | | |
| Water content (% wt) | 1 | 0.3 | | |
| Elect. resist. (ohm-m) | 1 | 7160 | | |

[a]Geometric mean.

samples 3.024 ±0.020 g/cm³. Thus, the higher density of the gabbros compared to the basalts comes primarily from their lower porosity.

The, mean bulk density of five serpentinized peridotites is 2.699 ±0.081 g/cm³ and the grain density of one sample 2.868 g/cm³.

## POROSITY AND WATER CONTENT

The porosity (percent pore volume) and water content (percent water by weight) was measured by weighing the minicores seawater saturated and after drying. The samples are kept as wet as possible after recovery so they have close to in situ water content. Some drying of the large cores is necessary for labeling and the cores generally were left in the core laboratory several days before the minicores were drilled. However, as seen from the slow drying of the minicores (see below) appreciable water probably is lost only from the surface few millimeters and not from the central part from which the minicores were drilled. The minicores were kept in seawater. No significant water appears to be lost by this treatment as the original weight is similar (and in fact generally greater) to that after drying and resaturating by the following procedure. The samples are evacuated in a vacuum desiccator, then immersed in seawater while still under vacuum, and finally squeezed at 1 kbar in a pressure chamber for several hours in a seawater-filled container (less than 25% of the water could be returned to the samples by the first step, the vacuum soaking which is commonly used, without the high pressure step).

The samples were dried simply in a vacuum desiccator sitting on a hot plate, the sample temperature being about 60° to 70°C. Drying was surprisingly slow. Figure 2 shows 18 consecutive weighings during 65 hr of drying for three samples ranging in porosity from 2% to 5%. An asymptote is approached only after about 60 hr. To show the drying process more clearly, the percent loss in weight also is plotted as a function of the square root of time. The loss of material by diffusion from a semi-infinite flat surface is proportional to the square root of time (no asymptote). Thus, as long as this plot is linear, the loss of water has not extended to a depth in the sample that is a major fraction of the minicore radius. The asymptotic approach to the dry weight appears to be quite rapid from the time the plot deviates from linearity. High porosity samples dry more rapidly than those of low porosity as expected, since high porosity rocks should have higher permeability. All of the samples were dried for at least 72 hr, which should remove at least 90% of the water. We estimate the accuracy of the porosity (percent of volume) to be from +0.4 to -0.2 plus +10% to -5% of value. The vacuum drying of the rocks leaves the dissolved salts behind so a small correction is needed. We have corrected for the residual salt assuming an original salinity of 35‰.

Accurate porosity values are difficult to obtain in low permeability rocks. The $H_2O^-$ reported in chemical analyses involves nearly complete dehydration, but a significant part of the rock water may be lost during crushing and handling before analysis. In contrast,

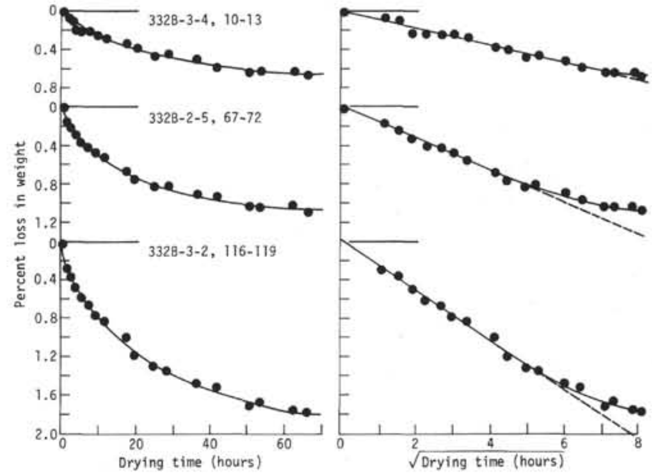


Figure 2. *Loss of weight of basalt samples with drying time at 65 °C with,* (a) *linear time scale,* (b) *square root time scale.*

whole rocks may start saturated at the initial weighing, but it takes a long time to dry them and pores that are not connected may never be evacuated. For accurate results the solid saturated rock must be weighed, then crushed with no loss of material, and finally dried and reweighed, or alternatively, the saturated rock could be crushed in a sealed container and the contents weighed before and after drying. Both procedures are difficult. Our method of weighing the saturated and vacuum-dried minicore seems the simplest procedure, gives sufficient accuracy for most purposes, and the sample subsequently can be used for other physical properties measurements.

Our drying at 60°C may partially dehydrate some of the clay minerals that occur frequently in the basalts, but the amount of water so lost probably is negligible since there are no obvious changes with drying seen in microscopic examination except for occasional shrinking away from the walls of vesicles of the clay minerals that coat them.

The basalt porosities are surprisingly high, 85 samples ranging from 1.7% to 24.5% (Figure 1) with a mean and standard deviation of individual values of 7.8 ±4.1%. The mean water content (percent weight) is 2.7 ±1.5%. It should be emphasized, as noted above, that the measured samples are representative of not more than 30% of the drilled section. The remainder, not recovered in the core, is probably fractured volcanic material which should have much higher large-scale porosity. In the upper part of layer 2 the porosity may be over 20%. The volume fraction of vesicles has an average similar to the porosity (M.F.J. Flower, personal communication), indicating that most of the pore space in the minicores is in vesicles. The mean mineral or grain density of each sample was determined using the measured porosities (see above). There is a small but significant correlation between porosity and computed grain density. Such a correlation probably is an artifact, which suggests that the porosity of the low porosity rocks is underestimated. Probably there is a significant fraction of the pore volume in the low porosity (low permeability) rocks that is not interconnected and thus is not dried by our procedure. Most of the pores and cracks in the high porosity vesicular rocks will be interconnected, so they are well dried. There also is the possibility that fine-grained, low porosity basalts trap significant amounts of gas on solidification so that some of the pores initially are gas rather than water filled.

The porosity of the gabbros is much lower than that of the basalts, four samples ranging from 0.8% to 1.7% with a mean of 1.0 ±0.4%. One serpentinized peridotite has a porosity of 0.8%.

## ELECTRICAL RESISTIVITY

Measurements of the electrical resistivity of ocean floor rocks are useful both for the interpretation of data of electromagnetic induction in the sea floor and for understanding the detailed composition and structure of these rocks. Natural electromagnetic induction has been used to estimate the resistivity with depth beneath the deep ocean floor (Cox et al., 1970; Greenhouse, 1972; Richards, 1970; Schneyer and Fonarev, 1968; Poehls and Von Herzen, 1975) and beneath oceanic islands (Hermance and Garland, 1968; Hermance, 1973; Klein, 1971; Elvers et al., 1970; Berdichevskiy and Chernyavskiy, 1970). The resistivities found for depths of a few tens of kilometers generally range between 1 and 6000 ohm-m, lower than those for most continental regions, and much lower than for most dry samples of probable oceanic crust and upper mantle rocks (Cox et al., 1970). Some oceanic basalts and ultramafic rocks recently measured, however, do have appropriately low resistivities either because of metallic mineral conduction or high porosity filled with low resistivity seawater (Stesky and Brace, 1973; Hyndman and Ade-Hall, 1974). In this study we have measured the resistivity of a large number of samples at atmospheric temperature and pressure and the variation of several at moderately high temperatures and pressures.

### Electrical Resistivity Versus Porosity and Pressure

The electrical resistivity of seawater (about 0.3 ohm-m) saturated minicores was measured at laboratory temperature of 22° to 24°C and atmospheric pressure onboard ship. A 50-Hz signal was used which should be too high for induced polarization effects and too low for capacitative effects. We have used a 1.00-v signal applied across the sample and a precision resistance in series, that is variable in steps so as to be the same order of magnitude of resistance as the sample, so there is approximately 0.5 v across the sample. The relative voltages were measured with a precision digital meter. Electrical contact was made by painting the ends of the samples with electrically conducting epoxy resin. The accuracy of resistance measurement is better than 1%, but variations in the state of the samples, particularly in surface water, make resistivities reproducible only to ±10%. The surface water is wiped off and the sample surface allowed to just dry (a few minutes) before the measurement.

The basalt resistivities all are very low, ranging from 6.3 to 5700 ohm-m with a geometric mean (mean on a logarithmic scale) of 220 ohm-m (Figure 1). The resistivity is strongly correlated with porosity (Figure 3) which suggests that ionic pore fluid conduction dominates in these samples. This mean is very similar to the 140 ohm-m found by Hyndman and Ade-Hall (1974) for Leg 26 DSDP basalts.

The electrical resistivity of a wide variety of water-saturated rocks in which conduction is primarily through pore fluid follows the simple Archie's Law dependence (Archie, 1942; Brace et al., 1965; Brace and Orange, 1968):

$$\frac{\sigma}{\sigma_f} = N^r$$

where $\sigma$ is the bulk rock conductivity, $\sigma_f$ the conductivity of the pore fluid, and $N$ is the porosity. The resistivity, $1/\sigma$, of seawater is about 0.20 ohm-m at laboratory temperature. The constant, $r$, depends on the nature of the fluid volumes in the rock. For pore conduction where the pores are roughly equidimensional, $r = 2$, i.e., for rocks with few cracks or at sufficiently high confining pressures for most cracks with small aspect ratios to be closed (e.g., Shankland and Waff, 1975). For rock with conduction primarily through fluid filled cracks $r \cong 1$. The measurements shown are at atmospheric pressure, but high pressure measurements

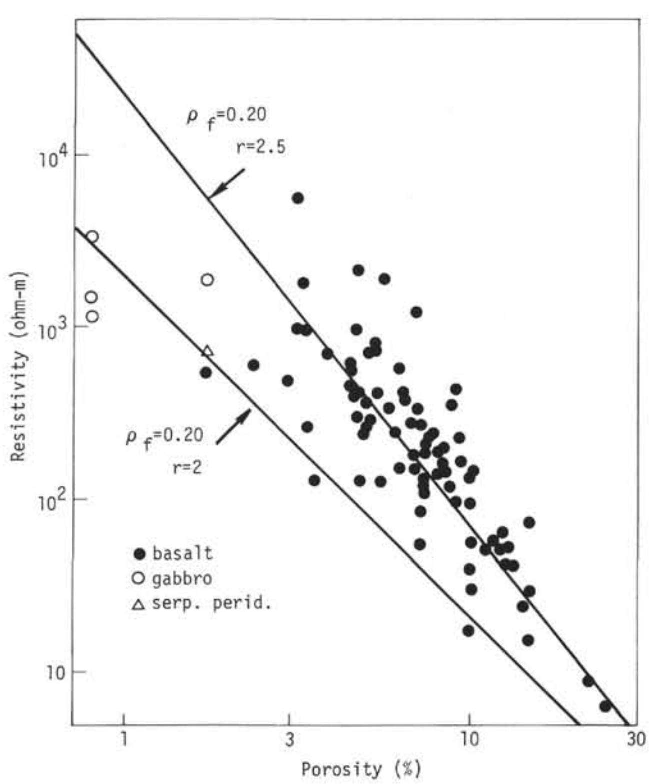


Figure 3. *Electrical resistivity of samples as a function of porosity at laboratory pressure and temperature. The lines give the predicted Archie's law dependence for a pore fluid resistivity of 0.20 ohm-m and r values of 2 and 2.5.*

(Figure 4) show very little pressure dependence of resistivity (also see Hyndman and Ade-Hall, 1974; Drury, 1976) so crack conduction should be negligible even at atmospheric pressure, and $r = 2$ is expected. Our measured basalt resistivities are slightly higher (crack conduction should give lower) but approximately follow the predicted $r = 2$ relation (Figure 3). The four gabbros and one serpentinized peridotite measured agree well with the relation.

The basalt resistivities may be slightly higher because: (a) The pore fluid has a resistivity higher than the 0.20 ohm-m estimated for the seawater at the measurement temperature. A fluid resistivity of 1.0 ohm-m would fit the data. (b) The pores were not completely saturated before resistivity measurement. The resistivity of the samples was measured after drying for porosity measurement, then resaturating. The resaturation may not have been complete. (c) The pores are not randomly distributed throughout the samples, the interconnection being less than predicted by the theory. (d) Some of the water is loosely bound in clay minerals which increases the resistivity at low temperatures. Some work still in progress indicates that clay minerals indeed are important for conduction in these rocks. Values of $r$ as high as 2.4 have been reported for shaley sands (Waxman and Smits, 1968), but the distribution of fluid volumes in such rocks probably is quite different to that of our basalts. An $r$ value of 2.5 would fit our basalt data well.

**Electrical Resistivity Versus Temperature**

The low mean resistivity (geometric mean) of 220 ohm-m for the basalts is further emphasized by the rapid decrease in resistivity with increasing temperature up to about 200°C. The resistivity of saline solutions decreases rapidly with increasing temperature to about 200°C (Quist and Marshall, 1968). The decrease is slightly more rapid for seawater (e.g., Horne, 1969). Above 300°C the resistivity increases. A similar behavior is expected for fluid-saturated rocks. Examples of the temperature dependence of resistivity of Leg 37 basalts are shown in Figure 5. There is little difference for the different pressures of 1 bar, 0.1 kbar, and 0.35 kbar. It is seen that the temperature dependence for most samples is much greater than for seawater. We have not examined this phenomenon in detail, but suggest that the large temperature dependence comes for conduction in clay minerals.

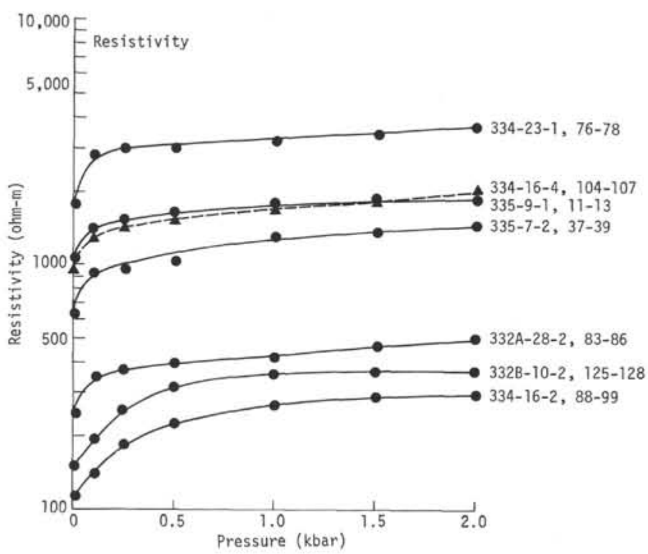


Figure 4. *Electrical resistivity of seawater saturated basalts as a function of pressure to 2.0 kbar at 23°C.*

**Electrical Resistivity With Depth in the Crust**

The resistivity as a function of depth in crustal layer 2 suggested by our data is shown in Figure 6, for low heat flow 1.0 $\mu$cal/cm²/sec (42 mW/m²) and high heat flow 2.0 $\mu$cal/cm²/sec (84 mW/m²) areas. These resistivities are a maximum, particularly in the upper part of layer 2 where there must be large-scale porosity that is not indicated by our samples (see Hyndman, this volume). The resistivity profile shown uses our mean basalt resistivity of 220 ohm-m at 23°C and the average temperature dependence of 5 of the 6 samples measured, omitting number 4 (Figure 5). The resistivity

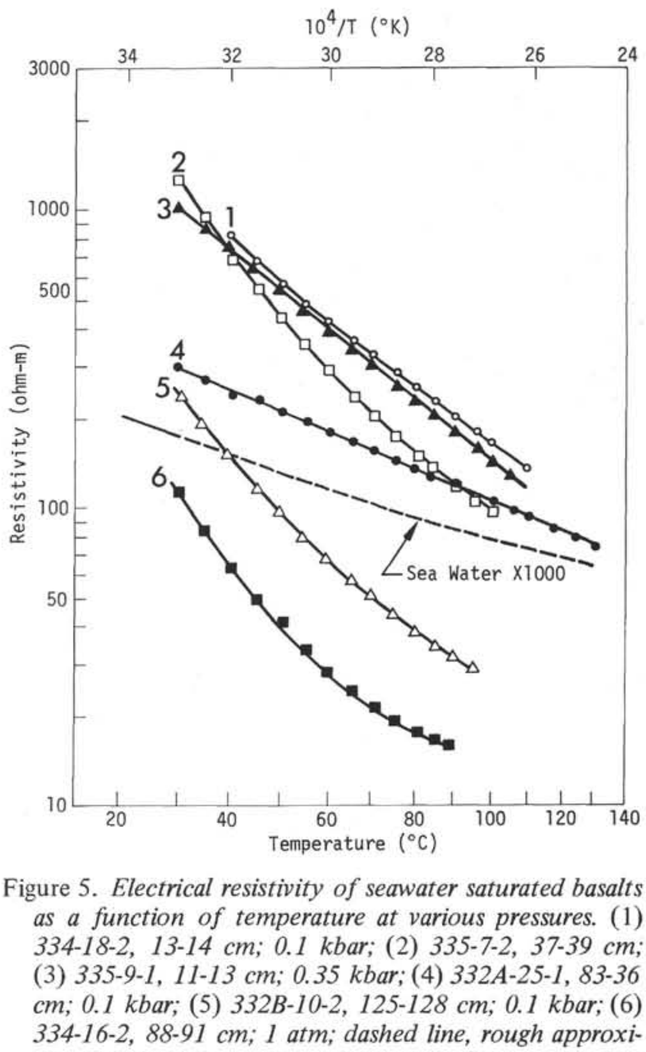


Figure 5. *Electrical resistivity of seawater saturated basalts as a function of temperature at various pressures.* (1) *334-18-2, 13-14 cm; 0.1 kbar;* (2) *335-7-2, 37-39 cm;* (3) *335-9-1, 11-13 cm; 0.35 kbar;* (4) *332A-25-1, 83-36 cm; 0.1 kbar;* (5) *332B-10-2, 125-128 cm; 0.1 kbar;* (6) *334-16-2, 88-91 cm; 1 atm; dashed line, rough approximation of 35% seawater extrapolated from data below 35°C, times 1000.*

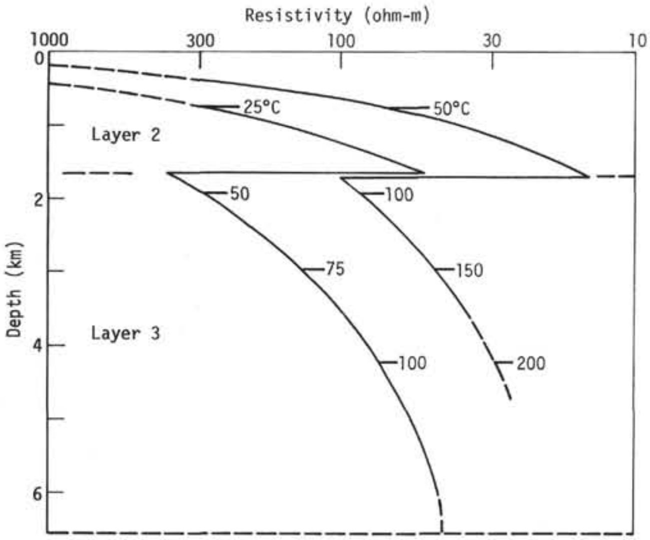


Figure 6. *Suggested electrical resistivity as a function of depth in the oceanic crust for areas of low and high heat flow* (1 and 2 $\mu$cal/cm$^2$ sec).

profile for layer 3 uses the mean gabbro resistivity at 23°C of 1710 ohm-m and assumes the same temperature dependence as for the basalt.

Our measured resistivities and estimates for the crust may be compared with the values mainly between 10 and 100 ohm-m found from geomagnetic induction measurements at variation periods from 0.5 to 2 hr (see above). Variations of these periods and the estimated resistivities will penetrate (skin depth) from 10 to several 100 km depth so, of course, are influenced by the upper mantle as well as by the crust. The resistivity in the uppermost mantle probably is much higher down to depths of at least several tens of kilometers, because there is little water present.

## SEISMIC VELOCITY

The laboratory measurements of seismic velocity on Leg 37 samples are discussed by Hyndman (this volume) and Christensen (this volume). A large number of samples were measured onboard ship to 2.0 kbar pressure. The mean compressional velocity of 79 basalts at 0.5 kbar pressure is 5.94 ±0.04 km/sec, the mean shear velocity for 37 samples is 3.27 ±0.02 km/sec, and the mean Poisson's ratio 0.295 ±0.008. There is a good correlation of increasing velocity with increasing density. The upper part of layer 2 probably contains extensive fractured basaltic material, sediment, and large scale voids so the velocity is estimated to increase from about 3.5 km/sec near the surface to about 5.5 km/sec below 500 meters depth in regions with a pronounced layer 2a such as Sites 332 and 333.

The mean velocities and standard error of mean for 10 gabbros at 0.5 kbar for compression is 7.07 ±0.05 km/sec, for shear is 3.72 ±0.03, and Poisson's ratio is 0.310 ±0.04. About 2% of cracks or other water-filled voids or 10% of 4.5 km/sec material is required to lower this compressional velocity to the mean seismic refraction layer 3 velocity of 6.7 km/sec. The higher velocity, higher density, and higher Poisson's ratio of the gabbros compared to the basalts appear to arise primarily from the lower porosity of the gabbros.

The mean velocity for three serpentinized peridotites at 0.5 kbar is for compression, 6.18 ±0.29 km/sec, for shear 2.96 ±0.13, and Poisson's ratio is 0.350 ±0.001. The velocities are somewhat lower and the Poisson's ratios much higher than found by seismic refraction for layer 3. Thus, serpentinized peridotites probably can be only a minor constituent of layer 3.

## THERMAL CONDUCTIVITY

The thermal conductivity of 21 samples—19 basalts, 1 gabbro, and 1 serpentinized peridotite—were measured by J. Jolivet using a divided bar apparatus. The details are given in Hyndman et al. (this volume). The mean sample temperature was 21°C and there was an axial force on disk samples of 75 kg/cm². The accuracy of individual values is about ±5%. The mean of the basalts is 3.97 ±0.04 mcal/cm/sec/°C (1.66 ±0.02 W/m/K). This value is slightly higher than the mean of 3.70 (1.55) for Leg 26 basalts from the Indian Ocean (Hyndman et al., 1974) which have an age of about 100 m.y. However, these latter samples show the effects of surface weathering. The Leg 37 mean value is similar to the average of 4.0 (1.7) for basalts from the Mid-Atlantic Ridge near 45°N (Hyndman and Jessop, 1971) and low porosity basalts from Hawaii (Robertson and Peck, 1975). It is slightly lower than a basalt measured by Somerton et al. (1963). A value of 4.0 (1.7) thus seems representative for young sea floor layer 2 basalts with older weathered basalts having somewhat lower conductivity. It should be pointed out, however, that only consolidated basalts are recovered in the DSDP drill core. Up to 50% of the upper part of layer 2 may be fractured basalt, sediment, and voids, particularly where there is a well-defined low seismic velocity layer 2a present. We estimate, for example, a conductivity of 3.5 ±0.3 (1.5 ±0.1) for the upper 500 meters at Site 332 increasing to 4.0 ±0.2 (1.7 ±0.1) in layer 2b.

The one gabbro measured from Site 334 has a conductivity of 5.7 mcal/cm/sec/°C (2.4 W/m/K) which is slightly higher than some previously measured gabbros (e.g., 4.60 [1.93] in Birch and Clark, 1940). Our sample probably is less weathered or altered. A conductivity of 7.6 (3.2) was obtained for a serpentinized peridotite from Site 334 which is higher than for most serpentinites (e.g., 4.3 [1.8] for two MAR samples, Hyndman and Jessop, 1971) which are more completely serpentinized. Thus, a conductivity of 5.5 ±0.5 (2.3 ±0.2) is a reasonable estimate for layer 3 if it is composed primarily of gabbro. The conductivity of the lower part of layer 3 may be higher if there is a significant fraction of only partially serpentinized peridotites.

## ACKNOWLEDGMENTS

Mr. R. Iuliucci and the other Deep Sea Drilling Project technical staff on board D.V. *Glomar Challenger* provided invaluable assistance with the shipboard measurements. Financial support for some of the laboratory facilities was provided by the National Research Council of Canada. The work was done while one of us (R.D. Hyndman) was at Department of Oceanography, Dalhousie University, Halifax, N.S., Canada.

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