

## 15. THE PHYSICAL PROPERTIES OF BASALT CORE SAMPLES FROM DEEP SEA DRILLING PROJECT LEG 78B HOLE 395A<sup>1</sup>

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### ABSTRACT

A detailed study has been made of the physical properties of core samples from Deep Sea Drilling Project Hole 395A. The properties include: density, porosity, compressional and shear wave velocity, thermal conductivity, thermal diffusivity, and electrical resistivity. Of particular importance are the relations among the parameters. Most of the variations in the basalt properties follow the porosity, with smaller inferred dependence on pore structure, original mineralogy differences, and alteration. The sample measurements give very similar results to (and extend previous data from) Mid-Atlantic Ridge drillholes. The sample data from this site and previous data are used to estimate relations between porosity and other large-scale physical properties of the upper oceanic crust applicable to this area. These relations are important for the analysis and interpretation of downhole logging measurements and marine geophysical data.

### INTRODUCTION

To complement the downhole measurements made in DSDP Hole 395A during Leg 78B, a series of new physical properties measurements have been made on the core obtained during the earlier drilling. These measurements include: density, porosity, compressional and shear wave velocity, thermal conductivity, thermal diffusivity, and electrical resistivity. Previous measurements of density, porosity, and compressional wave velocity on samples from the hole have also been reported in Melson, Rabinowitz, et al. (1979) and Schreiber and Rabinowitz (1979). It is emphasized that the sampling is not representative of the section drilled, first, because of the very poor and biased core recovery, and second, because the more friable, fractured, and weathered high-porosity samples could not readily be measured, particularly after 5 y. of drying and deterioration in storage.

The relations among these parameters, particularly those involving porosity or density, are important—when considered together with downhole logs and marine geophysical data, they permit estimates of the bulk *in situ* physical properties of the upper oceanic crust. This chapter first presents a summary of previous and new sample physical properties data from this hole and then discusses relations among the parameters applicable to the upper crust at this site.

### DENSITY AND POROSITY

The bulk density of 132 water-saturated basalt mini-cores was measured by DSDP Leg 45 scientists (Melson, Rabinowitz, et al., 1979), giving a mean of  $2.84 \pm 0.01$  g cm<sup>-3</sup>. Four additional measurements ( $2.78 \pm 0.03$ ) were reported by Schreiber and Rabinowitz (1979), and 12 more were measured in this study (Table 1) ( $2.84 \pm 0.03$ ). The overall mean is  $2.84 \pm 0.01$  g cm<sup>-3</sup>. The mean grain or matrix density computed from the bulk density and porosity of 60 samples by the shipboard Leg 45 scientists is  $2.957 \pm 0.003$  g cm<sup>-3</sup>. The much smaller variation in grain density indicates that most of the variation in the bulk density reflects differences in porosity rather than in mineralogy. As indicated below, the porosity estimates obtained later in shore laboratories are much lower and appear to be less reliable. A value of  $2.95$  g cm<sup>-3</sup> is taken to be the representative grain density.

The porosity of the 60 samples previously measured on board the ship were obtained from the weight loss upon slow vacuum oven drying of samples that had been kept saturated from the time of recovery. The mean is  $5.2 \pm 0.3\%$ . The values are accurate to about  $\pm 0.5\%$  porosity. There undoubtedly is some bias to values that are too low because of incomplete drying, especially for low-porosity samples, that have very low permeability. However, the lack of correlation between calculated grain density and porosity suggests that the error is not significant.

Porosity measurements in this study were attempted on 12 additional samples that had been shelf dried for over 5 y. The samples were resaturated by first placing them in a vacuum and then immersing and leaving them in seawater for several weeks. Then they were dried in a

<sup>1</sup> Hyndman, R. D., Salisbury, M. H., et al., *Init. Repts. DSDP, 78B*: Washington (U.S. Govt. Printing Office).

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Table 1. Physical properties of Hole 395A basalts.

Sample (interval in cm)	Density, $\rho$ (g cm <sup>-3</sup> )	Porosity, $\phi$ (%)		$V_p^a$ (km s <sup>-1</sup> )	$V_s^a$ (km s <sup>-1</sup> )	Poisson's ratio, $\sigma$	Thermal conductivity, $K$ (W m <sup>-1</sup> K <sup>-1</sup> )	Thermal diffusivity, $\kappa$ (mm <sup>2</sup> s <sup>-1</sup> )	Heat capacity, $C_p$ (J g <sup>-1</sup> K <sup>-1</sup> )	Electrical resistivity, $R_0$ (ohm-m)
		Measured	Inferred							
395A-5-1, 96-99	2.90	1.1	2.6	6.00	2.93	0.34				
395A-9-2, 34-36	2.88	2.2	3.6	6.01	3.31	0.28	1.74	0.63	0.95	388
395A-14-3, 53-55	2.86	1.7	4.7	6.12	3.15	0.32	1.72	0.68	0.88	230
395-22-1, 52-54	2.86	2.0	4.7	6.25	3.28	0.31	1.81	0.65	0.97	763
395A-26-1, 9-11	2.87	2.4	4.1	6.17	3.39	0.28	1.77	0.75	0.84	535
395A-32-2, 90-92	2.54	7.6	21.2	4.83	2.48	0.32				
395A-38-1, 75-76	2.87	2.4	4.1	5.83	3.17	0.29	1.73	0.70	0.86	489
395A-40-1, 138-140	2.83	5.4	6.2	5.95	3.38	0.26	1.71	0.64	0.95	247
395A-48-1, 36-38	2.85	2.3	5.2	6.03	3.14	0.31	1.79	0.64	0.97	411
395A-55-1, 44-46	2.87	1.4	4.1	6.04	3.23	0.30	1.67	0.64	0.90	977
395A-59-2, 97-99	2.92	0.8	1.6	6.37	3.43	0.30	1.93	0.72	0.92	940
395A-63-4, 38-40	2.88	1.0	3.6	6.38	3.31	0.32	1.82	0.72	0.88	426
Mean	2.84	2.5	5.5	6.00	3.18	0.30	1.77	0.68	0.91	484 <sup>b</sup>
$\pm$ s.d.	0.10	2.0	5.1	0.40	0.26	0.02	0.07	0.04	0.05	

Note: Blank spaces indicate no measurement made.

<sup>a</sup> = at 40 MPa.<sup>b</sup> Geometric mean.

vacuum oven. In these very low permeability rocks the values appear to be significantly too low, based on the very low mean porosity of  $2.5 \pm 0.6\%$ , on the very low implied grain density of  $2.89 \pm 0.02$  g cm<sup>-3</sup>, and on a clear correlation of grain density with porosity that does not exist in earlier shipboard data. The problem arises primarily from the difficulty of complete resaturation after 5 y. of shelf drying, and from the difficulty in subsequent complete vacuum drying. A drying temperature of 50°C for 48 hr. was used. Longer time-drying experiments by Hyndman and Drury (1977) indicate that this time should be just sufficient. The calculated porosities with only 24 hr. of drying were lower by about 10% of the values. Because of these difficulties, the new measured porosity estimates were discounted; those used in this study were obtained from the sample bulk densities, assuming a constant grain density of 2.95 g cm<sup>-3</sup>. As indicated later, there may be small real variations in grain density down the hole, but their effect on estimated porosities is probably negligible.

Careful examination of shipboard sample data indicates small but significant trends in the grain density through each volcanic unit. Most of the units defined by the chemistry and petrology of the cores and by the downhole logs exhibit a small increase in grain density with depth, especially in the upper part of the hole where there are more samples. These trends are not clearly evident in the sample density or porosity plots and thus appear to indicate an actual increase in average mineral density, probably reflecting a decrease in the amount of hydrated minerals with depth in each unit. This conclusion is substantiated by the small but probably significant trends in H<sub>2</sub>O<sup>+</sup> values shown by chemical measurements (Melson, Rabinowitz, et al. 1979). Probably the amount of hydrated minerals is related to the bulk porosity of the formation, because high-porosity horizons will provide greater permeability and thus access to water for hydration. As clearly indicated in the downhole logs (Mathews et al. this volume), there is a pronounced increase in large-scale porosity up through each unit as its eruptive cycle progresses.

## SEISMIC VELOCITIES

In this study, compressional and shear wave velocities were measured on 12, 2.5-cm-diameter, approximately 2.5-cm-long water-saturated samples to confining pressures of 600 MPa (6 kbar) (Tables 2, 3). The technique used was essentially that described by Christensen and Shaw (1970). The estimated accuracy is  $\pm 0.5\%$  for compressional and  $\pm 1.0\%$  for shear wave velocities. The fluid pressure medium was excluded from the samples by a thin copper foil jacket, and the pore pressures were maintained at a low value by placing a fine screen between the sample and the jacket. The compressional wave velocities of 6 water-saturated samples from Hole 395A had also been previously measured under a range of confining pressures by Schreiber and Rabinowitz (1979).

The 40 MPa (0.4 kbar) velocities are thought to represent best the *in situ* condition in the crustal section penetrated by the borehole. Although this confining pressure is somewhat higher than the *in situ* lithostatic pressure, it makes some allowance for closing microcracks and fractures generated by the drilling process. The *in situ* temperature is sufficiently close to that of the laboratory for no temperature correction to be needed.

Because there is a serious sampling bias (only competent samples are recovered and measured), it is important to relate velocity to other parameters that can be

Table 2. Compressional ( $V_p$ ) wave velocities.

Sample (interval in cm)	Wet- bulk density, $\rho$ (g/cm <sup>3</sup> )	Velocity (km/s at varying pressures [100 MPa or kbar])							
		0.2	0.4	0.6	0.8	1.0	2.0	4.0	6.0
395A-5-1, 96-99	2.90	5.96	6.00	6.02	6.04	6.06	6.11	6.15	6.18
395A-9-2, 34-36	2.88	5.98	6.01	6.03	6.05	6.06	6.11	6.15	6.19
395A-14-3, 53-55	2.86	6.08	6.12	6.15	6.18	6.21	6.30	6.39	6.46
395A-22-1, 52-54	2.86	6.21	6.25	6.27	6.29	6.31	6.35	6.40	6.46
395A-26-1, 9-11	2.87	6.14	6.17	6.20	6.21	6.23	6.29	6.38	6.48
395-32-2, 90-92	2.54	4.73	4.83	4.89	4.94	4.98	5.13	5.33	5.45
395A-38-1, 75-76	2.87	5.80	5.83	5.86	5.88	5.89	5.94	5.96	5.97
395A-40-1, 138-140	2.83	5.91	5.95	5.97	5.99	6.01	6.07	6.15	6.20
395A-48-1, 36-38	2.85	5.98	6.03	6.05	6.07	6.09	6.14	6.18	6.21
395A-55-1, 44-46	2.87	6.00	6.04	6.08	6.10	6.12	6.17	6.23	6.29
395A-59-2, 97-99	2.92	6.35	6.37	6.39	6.40	6.41	6.43	6.46	6.49
395A-63-4, 38-40	2.88	6.35	6.38	6.40	6.41	6.43	6.48	6.53	6.58

Table 3. Elastic constants and shear ( $V_s$ ) wave velocities.

Sample (interval in cm)	Pressure (100 MPa) (kbar)	$V_s$ (km/s)	$V_p/V_s$	Bulk modulus K (100 GPa) (Mbar)	Shear modulus $\mu$ (100 GPa) (Mbar)	Young's modulus E (100 GPa) (Mbar)	Lamé's constant $\lambda$ (100 GPa) (Mbar)
395A-5-1, 96-99	0.20	2.91	2.05	0.34	0.70	0.25	0.54
	0.40	2.93	2.05	0.34	0.71	0.25	0.55
	0.60	2.94	2.05	0.34	0.72	0.25	0.55
	0.80	2.94	2.05	0.34	0.72	0.25	0.56
	1.00	2.95	2.05	0.34	0.73	0.25	0.56
	2.00	2.98	2.05	0.34	0.74	0.26	0.57
	4.00	3.00	2.05	0.34	0.75	0.26	0.58
	6.00	3.01	2.05	0.34	0.76	0.26	0.58
395A-9-2	0.20	3.28	1.82	0.28	0.62	0.31	0.41
	0.40	3.31	1.82	0.28	0.62	0.32	0.41
	0.60	3.33	1.81	0.28	0.62	0.32	0.41
	0.80	3.34	1.81	0.28	0.63	0.32	0.41
	1.00	3.35	1.81	0.28	0.63	0.32	0.41
	2.00	3.38	1.81	0.28	0.64	0.33	0.42
	4.00	3.40	1.81	0.28	0.65	0.33	0.42
	6.00	3.41	1.82	0.28	0.66	0.34	0.44
395A-14-3, 53-55	0.20	3.09	1.97	0.33	0.69	0.27	0.51
	0.40	3.15	1.94	0.32	0.69	0.28	0.50
	0.60	3.18	1.93	0.32	0.70	0.29	0.50
	0.80	3.21	1.93	0.32	0.70	0.29	0.50
	1.00	3.23	1.92	0.31	0.71	0.30	0.51
	2.00	3.31	1.90	0.31	0.72	0.31	0.51
	4.00	3.37	1.90	0.31	0.74	0.33	0.52
	6.00	3.38	1.91	0.31	0.76	0.33	0.54
395A-22-1, 52-54	0.20	3.25	1.91	0.31	0.70	0.30	0.50
	0.40	3.28	1.91	0.31	0.71	0.31	0.50
	0.60	3.30	1.90	0.31	0.71	0.31	0.50
	0.80	3.31	1.90	0.31	0.71	0.31	0.51
	1.00	3.32	1.90	0.31	0.72	0.32	0.51
	2.00	3.36	1.89	0.31	0.72	0.32	0.51
	4.00	3.40	1.88	0.30	0.73	0.33	0.51
	6.00	3.43	1.88	0.30	0.75	0.34	0.52
395A-26-1, 9-11	0.20	3.36	1.83	0.29	0.65	0.32	0.43
	0.40	3.39	1.82	0.28	0.65	0.33	0.43
	0.60	3.40	1.82	0.28	0.66	0.33	0.44
	0.80	3.41	1.82	0.28	0.66	0.33	0.44
	1.00	3.42	1.82	0.28	0.67	0.34	0.44
	2.00	3.44	1.83	0.29	0.68	0.34	0.46
	4.00	3.46	1.84	0.29	0.71	0.34	0.48
	6.00	3.47	1.87	0.30	0.75	0.35	0.52
395A-32-2, 90-92	0.20	2.40	1.97	0.33	0.37	0.15	0.28
	0.40	2.48	1.95	0.32	0.38	0.16	0.28
	0.60	2.53	1.93	0.32	0.39	0.16	0.28
	0.80	2.56	1.93	0.32	0.40	0.17	0.29
	1.00	2.60	1.92	0.31	0.40	0.17	0.29
	2.00	2.69	1.91	0.31	0.42	0.18	0.30
	4.00	2.70	1.91	0.31	0.46	0.20	0.33
	6.00	3.13	1.85	0.29	0.59	0.28	0.40
395A-38-1, 74-76	0.40	3.17	1.84	0.29	0.59	0.29	0.40
	0.60	3.19	1.84	0.29	0.60	0.29	0.40
	0.80	3.21	1.83	0.29	0.60	0.30	0.40
	1.00	3.23	1.82	0.28	0.60	0.30	0.40
	2.00	3.28	1.81	0.28	0.60	0.31	0.40
	4.00	3.32	1.80	0.28	0.60	0.32	0.39
	6.00	3.33	1.79	0.27	0.60	0.32	0.39
	8.00	3.34	1.77	0.27	0.57	0.32	0.36
395A-40-1, 138-140	0.40	3.38	1.76	0.26	0.57	0.32	0.36
	0.60	3.41	1.75	0.26	0.57	0.33	0.35
	0.80	3.43	1.75	0.26	0.57	0.33	0.35
	1.00	3.45	1.74	0.25	0.57	0.34	0.35
	2.00	3.50	1.73	0.25	0.58	0.35	0.35
	4.00	3.53	1.74	0.25	0.60	0.35	0.37
	6.00	3.54	1.75	0.26	0.62	0.36	0.38
	8.00	3.11	1.92	0.31	0.65	0.28	0.47
395-48-1, 36-38	0.40	3.14	1.92	0.31	0.66	0.28	0.47
	0.60	3.16	1.91	0.31	0.66	0.28	0.47
	0.80	3.17	1.91	0.31	0.67	0.29	0.48
	1.00	3.18	1.92	0.31	0.67	0.29	0.48
	2.00	3.22	1.91	0.31	0.68	0.30	0.48
	4.00	3.26	1.90	0.31	0.69	0.30	0.48
	6.00	3.29	1.89	0.30	0.69	0.31	0.48
	8.00	3.20	1.88	0.30	0.64	0.29	0.45
395A-55-1, 44-46	0.40	3.23	1.87	0.30	0.65	0.30	0.45
	0.60	3.25	1.87	0.30	0.66	0.30	0.45
	0.80	3.28	1.86	0.30	0.66	0.31	0.45
	1.00	3.30	1.85	0.30	0.66	0.31	0.45
	2.00	3.35	1.84	0.29	0.66	0.32	0.45
	4.00	3.38	1.84	0.29	0.68	0.33	0.46
	6.00	3.39	1.86	0.30	0.70	0.33	0.48
	8.00	3.42	1.86	0.30	0.72	0.34	0.49
395A-59-2, 97-99	0.40	3.43	1.86	0.30	0.73	0.34	0.50
	0.60	3.44	1.86	0.30	0.73	0.35	0.50
	0.80	3.45	1.86	0.30	0.73	0.35	0.50
	1.00	3.46	1.85	0.29	0.73	0.35	0.50
	2.00	3.47	1.85	0.29	0.74	0.35	0.50
	4.00	3.49	1.85	0.29	0.75	0.36	0.51
	6.00	3.50	1.85	0.29	0.75	0.36	0.52
	8.00	3.28	1.94	0.32	0.75	0.31	0.54
395A-63-4, 38-40	0.40	3.31	1.93	0.32	0.75	0.32	0.54
	0.60	3.33	1.92	0.31	0.75	0.32	0.54
	0.80	3.35	1.91	0.31	0.75	0.32	0.54
	1.00	3.37	1.91	0.31	0.75	0.33	0.54
	2.00	3.42	1.89	0.31	0.76	0.34	0.54
	4.00	3.48	1.88	0.30	0.76	0.35	0.53
	6.00	3.51	1.87	0.30	0.78	0.36	0.54

measured or estimated for the upper oceanic crust. The relation between compressional and shear wave velocities and bulk density for ocean crust basalts was first extensively studied by Christensen and Salisbury (1975). This relation was based primarily on samples dredged or drilled from the top few meters of the crust that generally had extensive intergranular weathering porosity. It has proved to be an excellent representation of other basalts with primarily intergranular porosity (e.g., Hyndman et al., 1979). However, fresh ridge crest basalts with

primarily vesicular porosity appear to have velocities about 0.1 to 0.2  $\text{km s}^{-1}$  higher than this relation for a particular density (e.g., Christensen et al., 1979; Hyndman et al., 1979; Hyndman and Drury, 1976). Figure 1 shows that this 0.1 to 0.2  $\text{km s}^{-1}$  difference for both  $P$  and  $S$  wave velocities applies to the Hole 395A samples as well as those from nearby Hole 396B (Christensen et al., 1979). For comparison, the Wyllie time average equation commonly used for downhole velocity log interpretation in sedimentary sections gives density values that

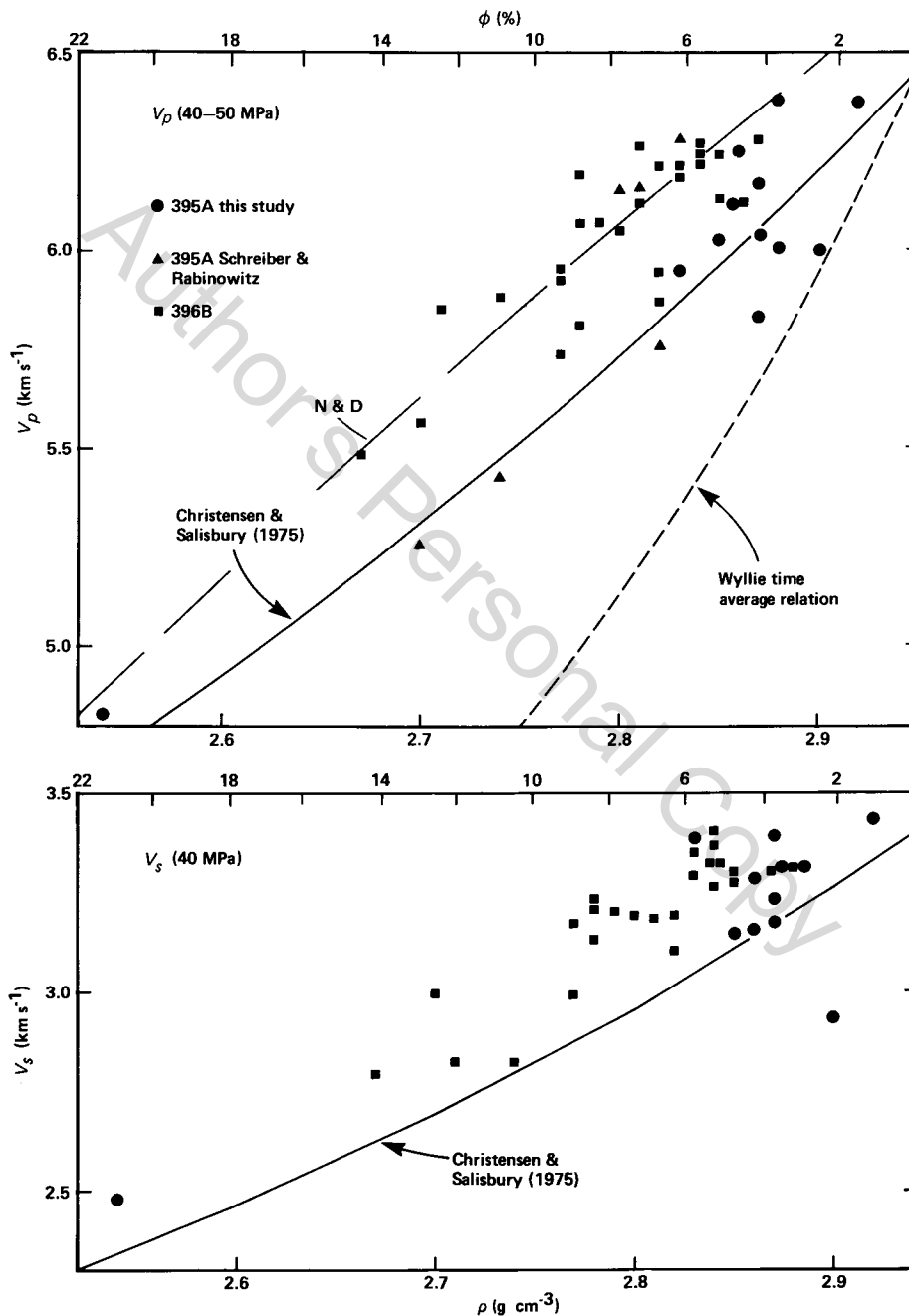


Figure 1. Compressional ( $V_p$ ) and shear wave ( $V_s$ ) velocity in  $\text{km s}^{-1}$ , at 40 to 50 MPa as a function of density ( $\rho$ ) and porosity ( $\phi$ ) for Holes 395A and 396B. (The best fits to DSDP data by Christensen and Salisbury [1975], the Wyllie time average relation [commonly used in well logging], and the Nafe and Drake relation [N & D] for  $V_p$  are shown.)

are much too high or porosity values too low for a particular velocity (Fig. 1) and thus will give very poor porosity estimates from downhole velocity data. In contrast, the Nafe and Drake (1961) curve for marine sediments gives densities that are somewhat too low or porosities that are too high.

Another important parameter for the interpretation and use of seismic data is Poisson's ratio (a function of the ratio of compressional to shear wave velocity). This parameter provides diagnostic information on rock type, but for the upper oceanic crust that has only small variations in mineralogy its more important dependence is on the amount and nature of the porosity. It is convenient to plot Poisson's ratio as a function of compressional wave velocity, but density or porosity give very similar relations. Hyndman (1979) defined two trends roughly linear over a limited range: (1) decreasing Poisson's ratio with increasing poorly connected vesicular porosity (small pore aspect ratio), and (2) increasing Poisson's ratio with increasing well-connected grain boundary or fracture porosity (large pore aspect ratio). Figure 2 shows that the new data from Hole 395A and that from nearby Hole 396B (Christensen et al., 1979) have a scatter that is too large to define any meaningful trends, although they are consistent with those trends previously proposed.

### THERMAL PROPERTIES

The thermal conductivity of 10 seawater-saturated basalt samples was measured on a divided bar instrument of the type described by Jessop (1970). Calibration was with fused silica using the conductivity values of Ratcliffe (1959). Measurements on individual disks were reproducible to  $\pm 2\%$  and the estimated accuracy is better than  $\pm 5\%$  (Table 1). The geometric mean thermal conductivity is  $1.77 \pm 0.02 \text{ W m}^{-1} \text{ K}^{-1}$  at  $25^\circ\text{C}$ . This mean is slightly higher than that for samples from other near-

by areas on the Mid-Atlantic Ridge: Hole 396, 1.71 (Erickson and Hyndman, 1979); Leg 37 holes, 1.66 (Hyndman and Drury, 1976). However, no friable or highly vesicular samples were measured in this study so the difference is probably the result of sampling bias.

The effect of pressure on the thermal conductivity of basalts is negligible for upper crustal conditions, but the effect of temperature is significant: a decrease of about  $0.005 \text{ W m}^{-1} \text{ K}^{-1}$  per degree (see review in Hyndman and Drury, 1976). In the section of the crust penetrated by Hole 395A, the effect is not important, but at the bottom of Layer 2 where the temperature may be over  $200^\circ\text{C}$  (Becker et al. this volume) the conductivity may be significantly lower.

The most important factor affecting *in situ* formation thermal conductivity in the upper crust is the large-scale porosity not represented in the samples. Of secondary importance are variations in the basalt composition such as olivine content and alteration. Many mathematical models have been proposed for the relation between porosity and thermal conductivity (e.g., review in Drury and Jessop, in press; Robertson and Peck, 1974). There is undoubtedly some dependence on the pore structure; for example, interconnected grain boundary and fracture porosity will give somewhat lower conductivity than the same, but poorly connected vesicular porosity. The most complete study of the relation between porosity and the thermal conductivity of basalts was by Robertson and Peck (1974). Their data are for subaerial volcanic island basalt samples, that may have a slightly different composition and porosity structure from the deep-sea basalt samples, but their results should be applicable to the latter. Their results also should be generally applicable to the bulk upper oceanic crust, although their (and the deep-sea) samples had mainly vesicular porosity, so the large-scale more interconnected bulk porosity of the upper oceanic crust may produce slightly lower bulk conductivity.

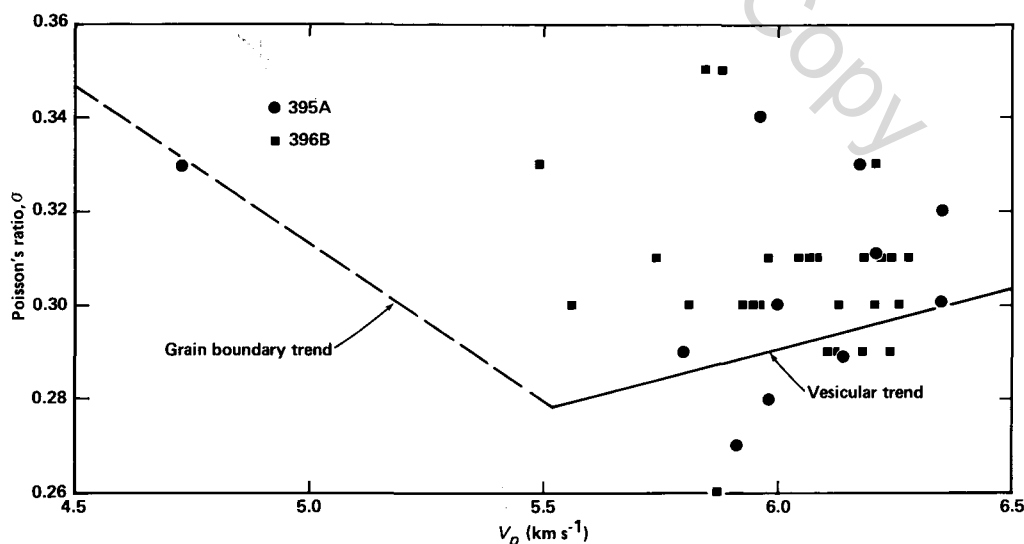


Figure 2. Poisson's ratio ( $\sigma$ ) as a function of compressional wave velocity ( $V_p$ ) for samples from Holes 395A and 396B. (The trends suggested by Hyndman [1979] for vesicular small pore aspect ratio porosity and for grain boundary large pore aspect ratio porosity are shown.)

Robertson and Peck (1974) found that none of the theoretical relations that they tried gave good agreement with the measured conductivities over a wide range of porosities. However, the empirical relation  $K = (1.36 - 0.60 \phi)^2 \text{ W m}^{-1} \text{ K}^{-1}$ , where  $\phi$  is the porosity, gave excellent agreement with measured values for water-saturated unweathered basalts with from 2 to 75% porosity and about 7% olivine content. They found that after porosity, the olivine content was the next most important factor. Although not well determined, the olivine content of the Hole 395A samples averaged about 7% (Melson and Rabinowitz et al., 1979; Dungan et al., 1979). As shown in Figure 3, 10% olivine content gives conductivities about 3% higher, and 5% olivine gives conductivities 3% lower.

Figure 3 shows sample conductivities versus density and inferred porosity for Hole 395A along with the data for Hole 396B and the mean values for Leg 37 samples from the Mid-Atlantic Ridge to the north. The porosity has been estimated from the density, assuming a grain density of  $2.95 \text{ g cm}^{-3}$ . These estimates of porosity are preferred to our direct measurements because of the difficulty of resaturating the samples for porosity measurement by direct drying (see earlier discussion). The Robertson and Peck relation gives slightly higher conduc-

tivity for a particular porosity or density than the geometric mean that is commonly found to be an adequate approximation for chip samples and some sediments (e.g., Sass, et al., 1971) (Fig. 3).

The third factor affecting the thermal conductivity deep-sea basalts, in addition to the porosity and the original mineral composition, is the degree of alteration and secondary mineralization. A progressive increase in conductivity with depth to over  $2.0 \text{ W m}^{-1} \text{ K}^{-1}$  resulting from increasing hydrothermal alteration is clearly seen in deep drillholes in Iceland and Bermuda (Oxburgh and Agrell, 1982; Hyndman et al., 1979). This type of alteration decreases the porosity, but the main factor is the change in mineralogy. However, it is not clear which secondary minerals are important. Low temperature alteration or weathering as seen in old ocean crust samples (e.g., Holes 417D, 418A, Fig. 3) has a similar effect of increasing the conductivity (Salisbury et al., 1979). The Hole 395A samples are generally little altered except near the bottom of the hole. However, there is one sample from near the bottom of the hole where alteration is important and its conductivity lies well above the Robertson and Peck (1974) relation (Fig. 3).

The thermal diffusivity of 10 seawater-saturated basalt samples was measured using the modified Angstrom

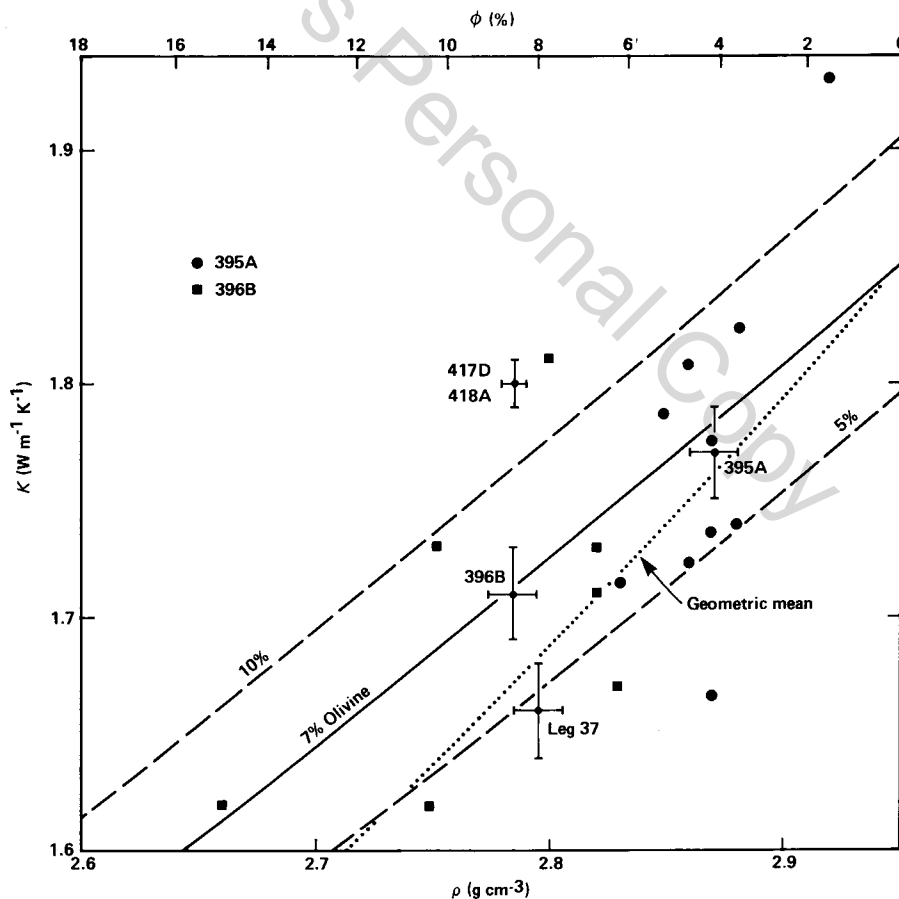


Figure 3. Thermal conductivity ( $K$ ) as a function of density ( $\rho$ ) and inferred porosity ( $\phi$ ) for samples from Holes 395A and 396B. (The Robertson and Peck [1974] empirical relation for basalts with 5, 7, and 10% olivine is shown along with the geometric mean relation.)

method as described by Drury et al. (in press) (Table 1). The sample disk is attached to a long rod of matching diameter and of similar and known thermal properties. The diffusivity is obtained by measuring the amplitude decrement and phase lag of a sinusoidal temperature wave that travels through the assembly (e.g., Sidles and Danielson, 1954; Kanamori et al., 1969; Savvides and Murray, 1978). The composite of sample plus rod is sufficiently long to approximate the infinite length assumed in Angstrom's original method (Angstrom, 1863). However, correction must be made for the reflection of heat from the thermal mismatch at the interface between sample and rod. An analogy with transmission line propagation has been used. The corrections are on the order of  $\pm 10\%$ . Calibration checks were made using fused silica. The reproducibility of measurements is about  $\pm 3\%$ , and the estimated accuracy is  $\pm 5\%$ . The mean sample temperature was  $27^\circ\text{C}$ . The diffusivity of rocks has frequently been estimated from mineral heat capacity estimates and from the thermal conductivity. However, as shown by Drury et al. (in press), such heat capacity estimates have an uncertainty commonly over  $10\%$ ; for higher accuracy, diffusivity must be measured directly.

The geometric mean diffusivity of the Hole 395A basalt samples is  $0.68 \pm 0.01 \text{ mm}^2 \text{ s}^{-1}$ . This measurement is in good agreement with the only previous direct measurement of tholeiitic basalt found in the literature— $0.67 \text{ mm}^2 \text{ s}^{-1}$  (Lindroth, 1976). By combining the thermal diffusivity, thermal conductivity, and density data, estimates of heat capacity can be obtained (Table 1). The

geometric mean of the 10 samples is  $0.912 \pm 0.014 \text{ J g}^{-1} \text{ K}^{-1}$ , which agrees well with the single value of  $0.924 \text{ J g}^{-1} \text{ K}^{-1}$  at  $25^\circ\text{C}$  quoted by Lindroth and Krawza (1971). By correcting for the porosity, the mean matrix or grain heat capacity is found to be  $0.864 \text{ J g}^{-1} \text{ K}^{-1}$ .

The dependence of heat capacity on porosity is readily estimated from the matrix heat capacity and density, and the relation between density and porosity, that is,  $C_p = (2.55 + 1.63\phi)/(2.95 - 1.93\phi)$ , where  $\phi$  is the porosity. This relation and the Robertson and Peck (1974) relation between porosity and thermal conductivity stated earlier then gives the relation between diffusivity and porosity as shown in Figure 4. The measured diffusivity values are consistent with this relation, although the range of porosity is too small to define it well.

### ELECTRICAL RESISTIVITY

The electrical resistivity of 10 basalt samples 2.5 cm in diameter and about 2.5 cm long was measured at about  $23^\circ\text{C}$  using a 1 volt 50 Hz signal. This frequency gives essentially the d.c. resistivity (e.g., Drury and Hyndman, 1979). As indicated earlier, the samples had been shelf-dried for 5 y.; we resaturated them by first placing them in a vacuum and then immersing and leaving them in seawater for several weeks to permit salinity equilibration between the seawater and the pore fluid. The seawater was microfiltered to prevent algal growth and had a resistivity of about 0.25 ohm-m at  $23^\circ\text{C}$ .

The geometric mean resistivity of the samples is 484 ohm-m, which is significantly higher than the means

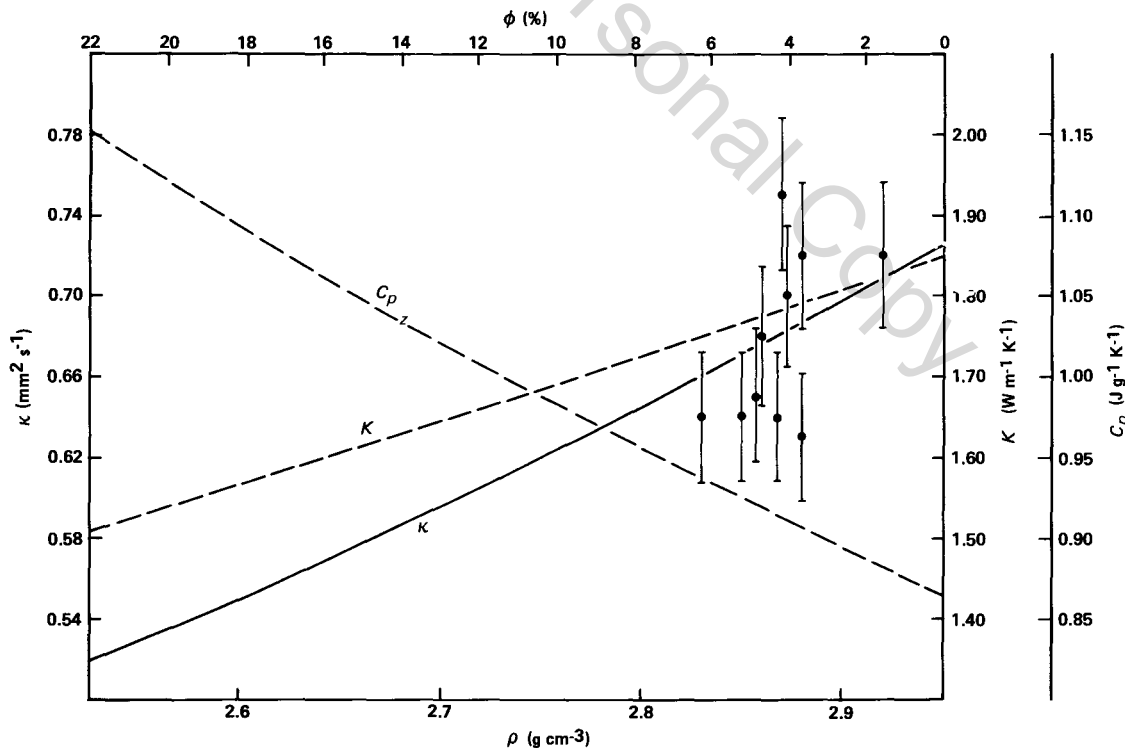


Figure 4. Thermal diffusivity ( $\kappa$ ) as a function of density ( $\rho$ ) and inferred porosity ( $\phi$ ) for Hole 395A samples. (The relations described in the text for diffusivity, conductivity [ $K$ ] and heat capacity [ $C_p$ ] are also shown.)

from previous studies of deep-sea basalt cores. The mean of 153 DSDP samples reported by Drury and Hyndman (1979) was 190 ohm-m. However, the mean porosity of the latter samples was 7%, much higher than that for the present samples, so the difference probably represents a sampling bias.

The electrical resistivity of fluid saturated rocks in which the matrix has a high resistivity follows the Archie (1942) relation

$$R_0/R_f = a \phi^{-m} \quad (1)$$

where  $R_0$  is the bulk resistivity,  $R_f$  is the fluid resistivity ( $R_0/R_f$  is called the formation factor),  $\phi$  is the porosity, and  $a$  and  $m$  are constants that depend on the pore structure (e.g., Brace et al., 1965). The seawater pore fluid has an estimated resistivity of about 0.25 ohm-m for the laboratory measurements and about 0.30 ohm-m *in situ* where the average temperature is about 6°C. The parameter " $a$ " is normally found to be about 1.0. The exponent constant " $m$ " is high (about 2) for poorly connected porosity and low (about 1) for well connected porosity (Brace et al., 1965). Deep-sea basalt core samples for which the porosity is largely vesicular and thus poorly connected have values of  $m = 2.0$  to  $2.5$  (e.g., Hyndman and Drury, 1976; Drury and Hyndman, 1979; Christensen et al., 1979; Hamano, 1980). However, there is considerable uncertainty because of the difficulty in measuring porosity and resistivity in resaturated samples. The relation between resistivity and porosity for Hole 395A samples is shown in Figure 5. The values of  $a = 1$  and  $m = 2$  give a good fit to the data, although there is considerable scatter. The porosity estimates are those

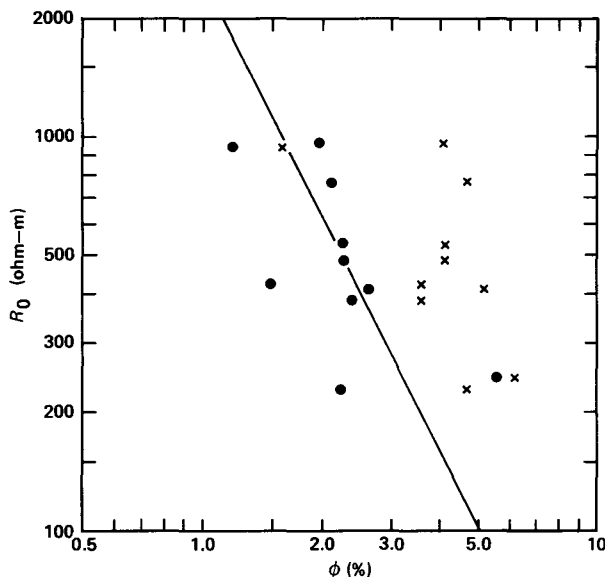


Figure 5. Electrical resistivity ( $R_0$ ) as a function of porosity ( $\phi$ ) for Hole 395A samples. (The solid circles indicate the porosities obtained by attempting to saturate the samples, then vacuum drying. This porosity is a measure of the fluid available for conduction during the resistivity measurement. The crosses show the porosity inferred from the density. The Archie's law relation, with  $a = 1$ ,  $m = 2$  and a fluid resistivity of 0.25 ohm-m, is also shown as a solid line.

obtained by drying after the resistivity measurement. As indicated earlier, the actual porosity is significantly higher. However, the drying estimate is an indication of the fluid available for conduction in the resistivity measurement for which there must have been incomplete saturation. The porosity that remained unfilled with seawater after the saturation process would behave as matrix during both the resistivity and drying porosity measurements. The porosity estimate from the density is also shown in Figure 5. Because the resistivity values would be even lower had saturation been complete, the difference emphasizes the difficulty of obtaining accurate porosity-resistivity relations for deep-sea basalts.

## RELATIONS BETWEEN POROSITY AND OTHER PROPERTIES IN THE UPPER OCEANIC CRUST

One of the main applications of core sample data is to aid in the interpretation of downhole logs and marine geophysical data, and thus in the estimation of the bulk *in situ* physical properties in the upper oceanic crust. Porosity is the main factor controlling variations in the basalt physical properties discussed earlier. The detailed original basalt mineralogy and subsequent alteration are generally of secondary importance. Thus porosity-dependent relations applicable to the larger-scale upper oceanic crust as seen in the downhole logs are of particular importance. In the upper crust, roughly half of the porosity is in the form of small-scale, primarily vesicular porosity as measured in core samples, and half is larger-scale better connected fracture and rubble porosity as measured in logs and marine geophysical measurements (e.g., Salisbury et al., 1979). A series of such relations with a brief discussion of their applicability to the Leg 78B downhole logs is given as follows:

### Porosity-Density

The basic relationship between porosity ( $\phi$ ) and density ( $\rho$ , in  $\text{g cm}^{-3}$ ) is

$$\rho = \rho_g - \phi (\rho_g - \rho_f). \quad (2)$$

If the grain density  $\rho_g = 2.95 \text{ g cm}^{-3}$  and the fluid density  $\rho_f = 1.02 \text{ g cm}^{-3}$ , this becomes

$$\rho = 2.95 - 1.93\phi. \quad (3)$$

Sample data indicate that less than 20% of observed basalt density variation comes from grain or matrix density differences. The grain density may be slightly higher for very fresh basalts (e.g.,  $2.97 \text{ g cm}^{-3}$ ) and lower for moderately weathered or altered rocks from old ocean crust (e.g.,  $2.90 \text{ g cm}^{-3}$ ). The relation just presented should give *in situ* density from porosity to about  $\pm 0.04 \text{ g cm}^{-3}$  for the upper crust along young ocean ridges. Conversely, density should give porosity to about  $\pm 3\%$ .

### Porosity-Seismic Velocity

For porosities below 20%, the relations between porosity and compression ( $V_p$ ) and shear ( $V_s$ ) wave velocities (in  $\text{km s}^{-1}$ ) are



$$V_p = 2.33 + 0.81 (2.95 - 1.93\phi)^{3.63} \quad (4)$$

$$V_s = 1.33 + 0.011 (2.95 - 1.93\phi)^{4.85} \quad (5)$$

which can be simplified to the polynomials

$$V_p = 6.44 - 9.61\phi + 7.20\phi^2 \quad (6)$$

$$V_s = 3.42 - 6.54\phi + 7.01\phi^2 \quad (7)$$

These relations are derived from the Christensen and Salisbury (1975) velocity-density relations and the density-porosity relationship shown earlier. The Christensen and Salisbury relation was based on the somewhat weathered samples then available from shallow drillholes. Samples from deeper crustal drillholes with very little alteration and mainly vesicular porosity have velocities about  $0.2 \text{ km s}^{-1}$  higher for the same porosity (N.I. Christensen and R. Carlson, 1982, unpublished data; see also Fig. 1). However, the original Christensen and Salisbury relation appears to be a good fit to samples with mainly but not totally interconnected grain boundary porosity (e.g., Hyndman et al., 1979). Thus the larger-scale porosity of the crust may give velocities that are low compared to the relations already described (see also Kirkpatrick, 1979). The overall relation for the upper crust combining the two scales of porosity should thus be approximated by these equations to  $\pm 0.2 \text{ km s}^{-1}$  or  $\pm 2\%$  porosity.

#### Porosity-Poisson's Ratio

An approximate relationship between porosity and Poisson's ratio ( $\sigma$ ) for porosities between 5 and 20% is

$$\sigma = 0.24 + 0.5\phi \quad (8)$$

This relation has been derived from a few suites of basalt samples with primarily interconnected (nonvesicular) grain boundary porosity (e.g., Hyndman, 1979) taken to represent large-scale fracture and rubble porosity. It assumes that such porosity dominates and that its form is similar to that of the samples. The Poisson's ratio-velocity relations have been combined with a linear approximation to the porosity relation just discussed, i.e.,  $V_p = 6.40 - 8.19\phi$ . This relation is poorly defined and its applicability to the upper oceanic crust is very tentative. The Poisson's ratio estimates from porosity are certainly no better than  $\pm 0.02$ . The relation is consistent with sediment data (e.g., Nafe and Drake, 1961). The general trend to high Poisson's ratio values in the high porosity upper crust is suggested by some shallow seismic refraction data (e.g., Clowes and Au, 1982) and by downhole logging in the Costa Rica rift area (Newmark et al., 1982).

#### Porosity-Thermal Conductivity

The relationship between porosity and thermal conductivity ( $K$ , in  $\text{W m}^{-1} \text{K}^{-1}$ ) is

$$K = (1.36 - 0.60\phi)^2 \quad (9)$$

This relation from Robertson and Peck (1974) for vesicular basalts with about 7% olivine gives a matrix conductivity of  $1.85 \text{ W m}^{-1} \text{K}^{-1}$ . It may give conductivities that are slightly too low for crust with significant hydrothermal alteration, and too high for crust with large-scale fracture porosity. However, it should estimate the overall upper crustal conductivity to  $\pm 0.07 \text{ W m}^{-1} \text{K}^{-1}$ .

#### Porosity-Diffusivity

Thermal diffusivity ( $\kappa$ , in  $\text{mm}^2 \text{s}^{-1}$ ) may be calculated from porosity from the relation

$$\kappa = (1.36 - 0.60\phi)^2 / (2.55 + 1.63\phi) \quad (10)$$

This relation comes from the above conductivity equation and the relation between heat capacity and porosity,  $C_p = (2.55 + 1.63\phi) / (2.95 - 1.93\phi)$ . The equation assumes the matrix or grain heat capacity of  $0.864 \text{ J g}^{-1} \text{K}^{-1}$  obtained from the mean sample heat capacity (discussed before) corrected for porosity. The above equation should give upper crustal diffusivity values accurate to  $\pm 0.05 \text{ mm}^2 \text{s}^{-1}$ .

#### Porosity-Electrical Resistivity

Finally, the basic relationship between porosity and bulk resistivity ( $R_0$ , in ohm-m) is given by Archie's Law (equation 1). If the pore fluid is seawater at  $5^\circ\text{C}$ , this becomes

$$R_0 = 0.30\phi^{-2} \quad (11)$$

The equation, particularly the exponent, is very dependent on the pore structure. An exponent slightly larger than 2 has been found for a number of suites of vesicular basalt samples, but a lower value is expected for interconnected large-scale crustal porosity. Thus the above equation probably is a reasonable representation of the upper oceanic crust (see also discussion in Becker et al., 1982).

### DISCUSSION

The new physical properties measurements on Hole 395A core samples presented here are generally consistent with the few previous measurements on samples from this hole and with data from other Mid-Atlantic Ridge samples. The exceptions are the porosity and electrical resistivity values that are very sensitive to incomplete resaturation of the shelf-dried samples. The new measurements include thermal conductivity, for which there are few on mid-ocean ridge basalts, and thermal diffusivity, for which no previous measurements exist.

Many of the commonly used relationships between different physical properties, particularly the porosity-dependent relations for basalt samples, are rigorously applicable only on a small scale. We have attempted to estimate which of these relations are applicable to the upper oceanic crust and the limits within which their use is justified. It must be remembered, however, that the total porosity of the upper oceanic crust is about half

small-scale porosity and half larger-scale fracture and rubble porosity. For the latter, some of these relations will not apply. Thus for interpreting downhole logs and marine geophysical data, it may in some cases be necessary to derive and combine empirical relationships that are appropriate for different scales of investigation.

# ACKNOWLEDGMENTS

We wish to acknowledge the assistance of Ms. W. Bentkowski in the thermal conductivity measurements. The velocity measurements were supported by the U.S. Office of Naval Research Contract No. N-00014-80-C-0252. This chapter is contribution No. 1052 of the Earth Physics Branch, Pacific Geoscience Centre, Department of Energy Mines and Resources.

# REFERENCES

- Angstrom, A. J., 1863. New method of determining the conductivity of bodies. *Phil. Mag.*, 25:130-142.
- Archie, G. E., 1942. The electrical resistivity log as an aid in determining some reservoir characteristics. *J. Petrol. Technol.*, 5:1-8.
- Becker, K., Von Herzen, R. P., Francis, T. J. G., Anderson, R. N., Honnorez, J., Adamson, A. C., Alt, J. C., Emmerman, R., Kempton, P. D., Kinoshita, H., Laverne, C., Mottl, M. J., and Newmark, R. L., 1982. *In situ* electrical resistivity and bulk porosity of the oceanic crust, Costa Rica rift. *Nature*, 300:594-598.
- Brace, W. F., Orange, A. S., and Madden, J. R., 1965. The effect of pressure on the electrical resistivity of water-saturated crystalline rocks. *J. Geophys. Res.*, 70:5669-5678.
- Christensen, N. I., Hyndman, R. D., Hull, J. M., and Salisbury, M. H., 1979. Seismic velocities, electrical resistivities, densities, and porosities of basalts from DSDP Leg 46. In Dmitriev, L., Heirtzler, J. R., et al., *Init. Repts. DSDP*, 46: Washington (U.S. Govt. Printing Office), 383-388.
- Christensen, N. I., and Salisbury, M. H., 1975. Structure and constitution of the lower oceanic crust. *Rev. Geophys. Space Phys.*, 13: 57-86.
- Christensen, N. I. and Shaw, G. H., 1970. Elasticity of mafic rocks from the Mid-Atlantic Ridge. *Geophys. J.*, 20:271-284.
- Clowes, R. M., and Au, D., 1982. *In-situ* evidence for a low degree of S-wave anisotropy in the oceanic upper mantle. *Geophys. Res. Lett.*, 9:13-16.
- Drury, M. J., Allen, V. S., and Jessop, A. M., in press. The measurement of thermal diffusivity of rock cores. *Tectonophysics*.
- Drury, M. J., and Hyndman, R. D., 1979. The electrical resistivity of oceanic basalts. *J. Geophys. Res.*, 84:4537-4545.
- Drury, M. J., and Jessop, A. M. in press. The estimation of rock thermal conductivity from mineral content—an assessment of techniques. *Zbl. Geol. Paläontol.*, 1 (1-2).
- Dungan, M. A., Long, P. E., and Rhodes, J. M., 1979. The petrology, mineral chemistry, and one-atmosphere phase relations of basalts from site 395. In Melson, W. G., Rabinowitz, P. D., et al., *Init. Repts. DSDP*, 45: Washington (U.S. Govt. Printing Office), 461-478.
- Erickson, A. J., and Hyndman, R. D., 1979. Downhole temperature measurements and thermal conductivities of samples, Site 396 Deep Sea Drilling Project Leg 46. In Dmitriev, L., Heirtzler, J., et al., *Init. Repts. DSDP*, 46: Washington (U.S. Govt. Printing Office), 389-400.
- Hamano, Y., 1980. Physical properties of basalts from Holes 417D and 418A. In Donnelly, T., Francheteau, J., Bryan, W., Robinson, P., Flower, M., Salisbury, M., et al., *Init. Repts. DSDP*, 51, 52, 53, Pt 2: Washington (U.S. Govt. Printing Office), 1457-1466.
- Hyndman, R. D., 1979. Poisson's ratio in the oceanic crust—a review. *Tectonophysics*, 59:321-333.
- Hyndman, R. D., Christensen, N. I., and Drury, M. J., 1979. Seismic velocities, densities, electrical resistivities, porosities and thermal conductivities of core samples from boreholes into the islands of Bermuda and the Azores. In Talwani, M., Harrison, C. G., and Hayes, D. E. (Eds.), *Deep Drilling Results in the Atlantic Ocean: Ocean Crust*: Washington (Am. Geophys. Union Geophys. Monogr., Maurice Ewing Series 2), pp. 94-112.
- Hyndman, R. D., and Drury, M. J. 1976. The physical properties of oceanic basement rocks from deep drilling on the Mid-Atlantic Ridge. *J. Geophys. Res.*, 81:4042-4052.
- , 1977. Physical properties of basalts, gabbros, and ultramafic rocks from DSDP Leg 37. In Aumento, F., Melson, W. G., et al., *Init. Repts. DSDP*, 37: Washington (U.S. Govt. Printing Office), 395-401.
- Jessop, A. M., 1970. The effect of environment on divided bar measurements. *Tectonophysics*, 10:39-49.
- Kanamori, H., Mizutani, H., and Fujii, N., 1969. Method of thermal diffusivity measurement. *J. Phys. Earth*, 17:43-53.
- Kirkpatrick, R. J., 1979. The physical state of the oceanic crust: Results of downhole geophysical logging in the Mid-Atlantic Ridge at 23°N. *J. Geophys. Res.*, 84:178-188.
- Lindroth, D. P., 1976. Thermal diffusivity of six igneous rocks at elevated temperatures and reduced pressures. *U.S. Bureau of Mines, Report of Investigations*, 7954.
- Lindroth, D. P., and Krawza, W. G., 1971. Heat content and specific heat of six rock types at temperatures to 1000°C. *U.S. Bureau of Mines, Report of Investigations*, 7503.
- Melson, W. G., Rabinowitz, P. D., et al., 1979. *Init. Repts. DSDP*, 45: Washington (U.S. Govt. Printing Office).
- Nafe, J. E., and Drake, C. L., 1961. Physical properties of marine sediments. In Hill, M. N. (Ed.), *The Sea* (Vol. 3): New York, (Interscience), 794-815.
- Newmark, R. L., Anderson, R. N., and Zoback, M. D., 1982. Sonic and ultrasonic logging results from the upper km of the oceanic crust. *EOS*, 63:433-434. (Abstract)
- Oxburgh, E. R., and Agrell, S. O., 1982. Thermal conductivity and temperature structure of the Reydarfjörður borehole. *J. Geophys. Res.*, 87:6423-6428.
- Ratcliffe, E. H., 1959. Thermal conductivities of fused and crystalline quartz. *Brit. J. Appl. Phys.*, 10:22-25.
- Robertson, E. C., and Peck, D. L., 1974. Thermal conductivity of vesicular basalt from Hawaii. *J. Geophys. Res.*, 79:4875-4888.
- Salisbury, M. H., Stephen, R., Christensen, N. I., Francheteau, J., Hamano, Y., Hobart, M., and Johnson, D., 1979. The physical state of the upper levels of Cretaceous oceanic crust from the results of logging, laboratory studies and the oblique seismic experiment at DSDP Sites 417 and 418. In Talwani, M., Harrison, C. G., and Hayes, D. E., (Eds.), *Deep Drilling Results in the Atlantic Ocean: Washington* (Am. Geophys. Un. Geophys. Monogr., Maurice Ewing Series 2), pp. 113-134.
- Sass, J. H., Lachenbruch, A. H., and Munroe, R. J., 1971. Thermal conductivity of rocks from measurements on fragments and its application to heat flow determinations. *J. Geophys. Res.*, 76: 3391-3401.
- Savvides, N., and Murray, W., 1978. Apparatus for the measurement of thermal diffusivity featuring a low-frequency sine-wave generator and a digital phase meter. *J. Phys. Earth: Sci Instrum.*, 11: 941-947.
- Schreiber, E., and Rabinowitz, P. D., 1979. Acoustic wave velocity measurements of oceanic crustal samples—DSDP Leg 45. In Melson, W. G., Rabinowitz, P. D., et al., *Init. Repts. DSDP*, 45: Washington (U.S. Govt. Printing Office), 383-386.
- Sidles, P. H., and Danielson, S. 1954. Thermal diffusivity of metals at high temperatures. *J. Appl. Phys.*, 25:58-66.

Date of Initial Receipt: November 1, 1982

Date of Acceptance: May 13, 1983