SEISMIC VELOCITIES, DENSITIES, ELECTRICAL RESISTIVITIES, POROSITIES AND THERMAL CONDUCTIVITIES OF CORE SAMPLES FROM BOREHOLES INTO THE ISLANDS OF BERMUDA AND THE AZORES

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Abstract. A detailed study of physical properties was made on core samples from 1 km deep boreholes into islands of Bermuda and the Azores. The properties measured are seismic velocity, density, electrical resistivity, porosity and thermal conductivity. The properties of subaqueous lava flow samples from Bermuda and the Azores are very similar to those for deep ocean tholeiites. Subaerial flows and pyroclastics from the Azores have quite different properties, reflecting their higher porosity. Lamprophyric intrusive sheets from Bermuda also have significantly different and unusual properties, notably lower velocity and higher density than the flows. It is concluded that aseismic ridges, islands and seamounts may have significantly different physical properties from normal upper oceanic crust if they contain subaerially erupted material as on Sao Miguel, Azores or are produced by intrusion of lamprophyric sheets or dykes into older oceanic crust as inferred for Bermuda.

Introduction

Laboratory data on the physical properties of rocks are an essential element in the interpretation of surface geophysical measurements, and for structural and petrological models. Much work has been done on oceanic crustal samples, particularly those from the Deep Sea Drilling Project, that gives a reasonably clear picture of the upper part of the normal oceanic crust (e.g. Christensen and Salisbury, 1975; Hyndman and Drury, 1976.) However, there are some anomalous oceanic areas of particular significance. These are the aseismic ridges and oceanic islands, which reflect higher than normal vulcanism and are referred to as hot spots or hot spot traces. They may arise as the surface expression of deep mantle plumes (Wilson, 1965; Morgan, 1971) or from propagating lithospheric fractures (e.g. Shaw and Jackson, 1973). In either case, compared to normal oceanic crust they are characterized by greater crustal thickness and by different composition.

This article reports seismic velocities, densities, electrical resistivities, porosities, and thermal conductivities for basaltic samples from approximately 1 km deep boreholes into the islands of Bermuda and São Miguel, Azores (Fig. 1 and 2) and discusses the geophysical significance of these measurements.

An 802 m deep borehole was drilled into the islands of Bermuda at 32°22'N, 64°42'W (Fig. 1) at an elevation of 10 m during the summer of 1972 for Dalhousie University and Lamont-Doherty Geological Observatory. The hole was cored with almost complete recovery for geological and geophysical studies. The islands of Bermuda lie approximately 1100 km east of the North American coast, about one third of the distance to the mid-Atlantic ridge. They have a maximum elevation of 30 m and occupy the southern 7% of the 116 km^2 platform that makes up the top of the Bermuda pedestal. Only limestones are exposed, the volcanic platform lies at an average of about 80 m below sealevel (Gees and Medioli, 1970). Bermuda is the northernmost and largest of three seamounts of the Bermuda Rise. From potassium-argon dating, the Bermuda seamount appears to have been formed 33 m.y. ago by the intrusion of numerous lamprophyric sheets into oceanic crust that was then about 80 m.y. old (Reynolds and Aumento, 1974; Gees, 1969). The surrounding seafloor is now about 110 m.y. old (Larson and Pitman, 1972). This age of the seamount is substantiated by the age of the oldest limestones overlying the volcanics (see

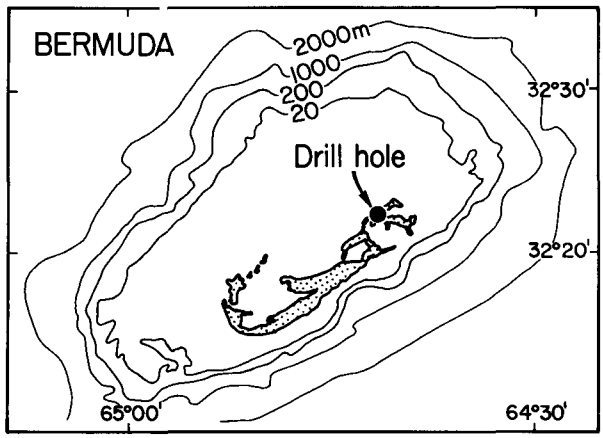


Fig. 1. Location of the Bermuda borehole.

Wilson, 1963) and by the low heat flow and the subsidence history of the platform (Hyndman et al., 1974a). Temperatures in the drilled section are low, reaching only about 38°C at the hole bottom (Fig. 3). In addition to radioactive dating, heat flow and heat production studies, extensive rock magnetic studies have been performed on the core (Ade-Hall et al., 1973) as well as detailed petrology (Aumento and Ade-Hall, 1973) and geochemistry (Aumento et al., 1976; and F. Aumento and B. Gunn, personal communication). A preliminary report of laboratory velocity measurements was presented by Barrett et al. (1973).

The borehole first penetrated 36 m of limestone, then 772 m of volcanic rocks made up of over 1000 igneous units. The whole core shows evidence of hydrothermal alteration both in petrology and in high Curie temperatures. Of the igneous units, 64% were altered subaqueously extruded tholeiite lavas and 36% were thin lamprophyric sheets of unusual composition, e.g. low silica, high Ti, K etc. It appears that any subaerial volcanics that were present have been eroded by a subsequent uplift before final subsidence. Such uplift could have been produced by the intrusion of the sheets.

During the summer of 1973, a 981 m deep borehole was drilled into the island of São Miguel, Azores at 25°31'W, 37°48.9'N (Fig. 2) at a surface elevation of 72 m for Dalhousie University and Lamont-Doherty Geological Observatory. Core recovery was almost complete except in the upper 148 m. The Azores are a group of 9 islands aligned in a NW-SE chain which crosses the mid-Atlantic ridge near 39°N. The region is marked by a change in trend of, and by a broadening of the ridge into a large platform (Krause and Watkins, 1970) and a positive regional gravity anomaly (e.g. Kaula, 1972). The Terceira rift, along which the main islands are located, trends from the ridge toward Gibraltar, probably defining a plate boundary that originated about 45 m.y. ago (Krause and Watkins, 1970). São Miguel is the largest island and lies 400 km east of the mid-Atlantic ridge crest. The drill site was located on the lower southern flank of the volcano Agua de Pau, which has erupted 5 times in the past 4,600 years, the last in 1563. Numerous hot springs and sporadic seismic activity indicate the volcano to be only dormant at present.

Extrusive lavas make up 72% of the Azores core in 140 flows averaging 5 m in thickness (Muecke et al., 1974). The main rock types are alkali basalts, hawaiites and mugearites. Three trachyte flows make up 6% of the total. The upper 763 m consists of 3 distinct subaerial volcanic eruptive sequences. This section is underlain by a 107 m transition sequence of basaltic sandstones and basaltic breccias, then by pillows and massive basaltic flows of subaqueous origin.

Temperatures were measured in the hole at intervals during drilling until the hole was terminated by the eruption of steam. Only bottom hole measurements appeared to represent in-situ temperatures, (Fig. 3) giving gradients up to 250°C/km. Temperatures in the hole nearer the surface were dominated by in-hole convection (see Muecke et al., 1974). The geothermal gradient is dominated by probably near-horizontal water flow so a meaningful geothermal heat flux cannot be ascertained. The temperature at the bottom of the hole was 200°C. Two K-Ar dates of 117,000 ± 24,000 yr at 57 m and 280,000 ± 140,000 yr at 950 m (Muecke et al., 1974) and that all of the core was normally magnetized presumably in the Brunhes polarity epoch that extends to 690,000 yr ago, indicate the youth and rapid formation of this volcanic edifice. It appears that the island has subsided over 900 m in less than 0.5 m.y. probably in isostatic response to the increasing volcanic load. A

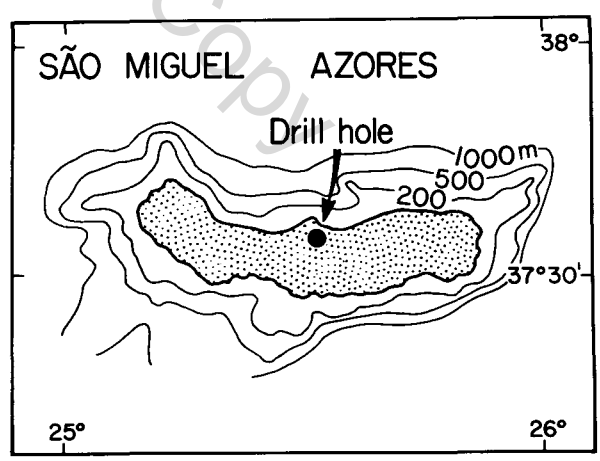


Fig. 2. Location of the Azores borehole.

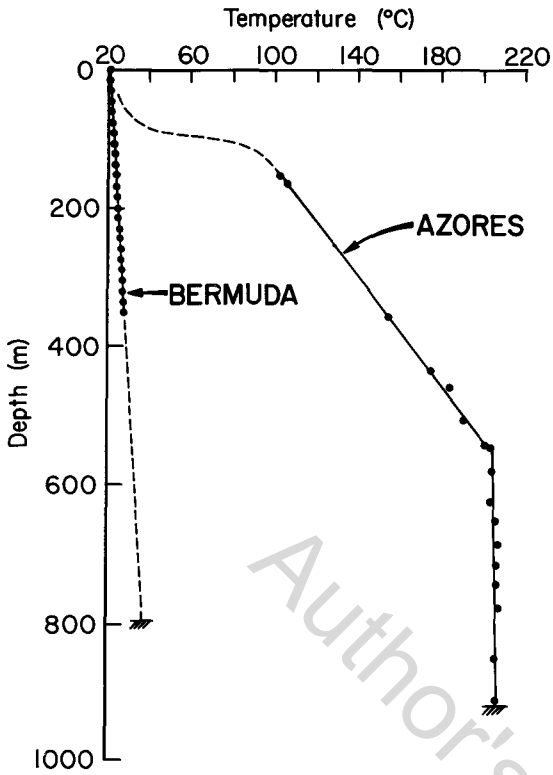


Fig. 3. Temperature-depth profiles in the Bermuda and Azores boreholes.

detailed description of the petrology of the core has been made by McGraw (1976) and extensive rock magnetic studies have been made (J.M. Hall, N.D. Opdyke and W. Lowrie, personal communication).

Density

The bulk densities of water saturated samples (stored in water from time of drilling) were determined by weighing the cores in air and obtaining the volume either by measuring the weight loss with the samples suspended in distilled water or by measurements of sample dimensions (Tables 1 and 2). The estimated accuracy is $\pm$ 0.01 g cm^{-3}. The grain or mineral densities for samples for which porosities are reported were determined simply by subtracting the mass and the volume of the pore fluid and recomputing the densities. Since the porosities tend to be underestimated because of incomplete drying (i.e. effective porosity), the grain density also will be too low. The estimated error limits are $\pm$ 0.02 to -0.01 g cm^{-3}.

The mean bulk density of 43 samples from Bermuda is 2.83 $\pm$ 0.02 g cm^{-3} with no systematic variation with depth (Fig. 4). However, the intrusive lamprophyric sheets have a significantly higher mean of 2.90 $\pm$ 0.03 g cm^{-3} (16 samples) and larger variation compared to the flows, with a mean of 2.79 $\pm$ 0.01 g cm^{-3} (27 samples), the higher mean perhaps reflecting the high density of the pyroxene in the former since the porosities are similar. Taking the mean sheet and flow densities and their relative proportions of 36% and 64% respectively in the drilled section gives a section mean of 2.83 g cm^{-3}, identical to the sample mean indicating that there was a representative sampling of sheets and flows. The mean grain density of 17 Bermuda samples is 2.85 $\pm$ 0.02 g cm^{-3}; the mean for 17 samples of the flows is 2.81 $\pm$ 0.01 and for 8 samples of the sheets is 2.92 $\pm$ 0.05 g cm^{-3}.

TABLE 1. Physical Properties of Samples from Bermuda

Property	Lava Flows		Lamprophyric Sheets		All Samples	
	Number	Mean	Number	Mean	Number	Mean
Compressional velocity (0.4 kbar) km s^{-1} (Vp)	12	5.97 ± 0.08	8	5.89 ± 0.07	20	5.94 ± 0.05
Shear velocity (0.4 kbar) km s^{-1} Vs)	12	3.22 ± 0.06	8	3.07 ± 0.05	20	3.16 ± 0.04
Poisson's ratio (σ)	12	0.295 ± 0.003	8	0.314 ± 0.006	20	0.303 ± 0.004
Bulk density, g cm^{-3}	27	2.79 ± 0.01	16	2.90 ± 0.003	43	2.83 ± 0.02
Grain density, g cm^{-3}	17	2.81 ± 0.01	8	2.92 ± 0.005	25	2.85 ± 0.02
Porosity, % vol.	17	2.6 ± 0.4	8	3.4 ± .05	25	2.9 ± 0.3
Electrical resistivity ohm m	17	1320 ± 400	6	340 ± 120	23	780 ± 210
Thermal conductivity W m^{-1} K^{-1}	32	2.08 ± 0.02	21	2.19 ± 0.07	53	2.12 ± 0.03

TABLE 2. Physical Properties of Samples from the Azores

Property	Subaerial		Subaqueous		All Samples	
	Number	Mean	Number	Mean	Number	Mean
Compressional velocity (0.4 kbar) km s^{-1} (Vp)	15	4.92 ± 0.17	3	5.60 ± 18	19	5.08 ± 0.15
Shear velocity (0.4 kbar) km s^{-1} (Vs)	15	2.49 ± 0.12	3	2.97 ± 0.09	19	2.60 ± 0.11
Poisson's ratio (σ)	15	0.328 ± 0.007	3	0.307 ± 0.003	19	0.323 ± 0.006
Bulk density, g cm^{-3}	41	2.60 ± 0.004	11	2.79 ± 0.04	56	2.64 ± 0.03
Grain density, g cm^{-3}	19	2.88 ± 0.003	3	2.85 ± 0.03	23	2.87 ± 0.02
Porosity	19	10.6 ± 1.7	3	4.4 ± 1.2	23	9.4 ± 1.5
Electrical resistivity ohm m	19	82 ± 38	3	230 ± 170	23	110 ± 50
Thermal conductivity W m^{-1} K^{-1}	31	1.75 ± 0.05	5	1.73 ± 0.04	40	1.75 ± 0.04

The mean bulk density of 56 samples from the Azores borehole is 2.64 ± 0.03 g cm^{-3}. There is some systematic increase in density with depth (Fig. 4). The mean is much lower and the scatter of values is more than double that for the Bermuda samples reflecting the wide variation and generally high porosity of the Azores rock types, particularly the pyroclastics in the subaerial section. The mean for the subaerial section is 2.60 ± 0.04 g cm^{-3} (41 samples). The mean for the igneous-sedimentary transition sequence is 2.71 ± 0.09 g cm^{-3} (4 samples), and the mean for the subaqueous sequence is 2.79 ± 0.04 g cm^{-3} (11 samples). As expected, the mean density of the subaqueous Azores rocks is the same as the mean of 2.79 for the flows from Bermuda, and for deep ocean tholeiitic basalts (e.g. Hyndman and Drury, 1976). The mean grain density of 23 Azores cores is 2.87 ± 0.02 g cm^{-3}, not greatly different from the Bermuda value, indicating that most of the bulk density differences arise from variations in porosity or water content. Bulk density is plotted as a function of porosity in Fig. 5 for both Bermuda and Azores samples. The intercept of about 2.85 gm cm^{-3} is in agreement with the above grain density estimates. This lower grain density and lower porosity, compared to the mean respective values for fresh deep ocean basalts (2.95 g cm^{-3} and 7.8%) suggests that initially free pore water in the Bermuda and Azores rocks is now bound in hydrous minerals, there being little change in total water in the alteration process.

For both the Bermuda and Azores core there is some sampling bias, low density fracture zones and rubble not being recovered, so the insitu density is lower than the means from the core. We believe that the effect is significant only for the upper several hundred meters of the Azores hole where core recovery was poor.

The densities of subaerial volcanic flow samples have been reported for the island of Hawaii, which is in a similar setting to São Miguel, by Manghnani and Woollard (1968) and Kinoshita (1965). They found that most surface and borehole samples had a density of 2.3 g cm^{-3}, in close agreement with the one near surface sample from the Azores borehole but considerably lower than the mean of 2.60 g cm^{-3} for the subaerial section. They found a number of dykes and plugs with much higher densities of 3.0 to 3.2 g cm^{-3}. Woollard (1951) estimated the mean density of the Hawaii island of Oahu to be 2.3 g cm^{-3} from that required to minimize the effect of topography on island gravity values. However, taking a larger area, including marine gravity and a more sophisticated model, Walcott (1970) and Watts and Cochran (1974) estimated a mean density of 2.80 g cm^{-3} for the Hawaiian Islands, close to our Bermuda mean and Azores subaqueous samples suggesting only the near surface rocks have low density (see also Worzel and Harrison, 1963). A similar density has been estimated from gravity measurements for the Great Meteor seamount (Watts et al., 1975). Lower values of about 2.5 g cm^{-3}, however, were estimated for the Caryn seamount by Worzel and Talwani (1959) and of 2.3 g cm^{-3} for the Jasper seamount off California by Harrison and Brisbin (1959). A possible explanation for this low density is that they may have carbonate caps. In a detailed gravity study of the Cook Islands, Robertson (1967) estimated that the island platforms had an average density of 2.32 g cm^{-3} with large central plugs of 2.88 g cm^{-3} (see also data in Malahoff, 1969). From Bermuda gravity measurements Woollard (1954) computes a

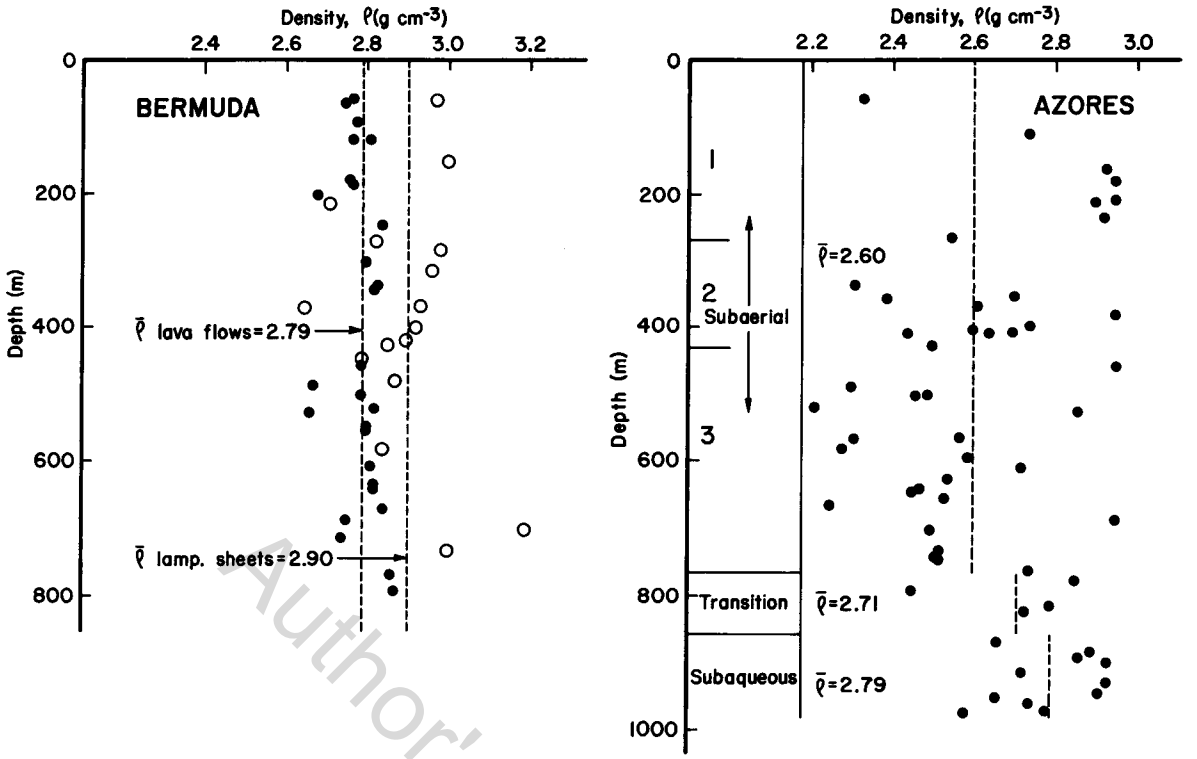


Fig. 4. Density as a function of depth in the Bermuda and Azores boreholes.

mean density of 2.80 for the seamount structure itself with lower density material, 2.50 g cm^{-3} of several km thickness on the adjacent sea floor.

On the basis of our sample densities from Bermuda and the Azores and the above gravity estimates for seamounts and islands we conclude: The mean density for the bulk of oceanic volcanic islands and seamounts is about 2.8 g cm^{-3}, which is appropriate for subaqueous pillow lava flows with a few higher density dykes and sills. Density means as low as 2.3 g cm^{-3} and commonly 2.6 g cm^{-3} exist for the subaerially erupted volcanics which make up the upper part of active volcanic islands. As occurs in São Miguel, subaerially erupted volcanoes may extend to a kilometer or more below sea level in an active volcanic island because of loading subsidence (see Muecke et al., 1974). As in the case for Bermuda, most old seamounts will have the subaerial volcanics removed by erosion, so that only the higher density subaqueous volcanics remain (see discussion in Hyndman et al., 1974a). A thick low density coral cap, of course, exists on the tops of many shallow sea mounts. The fairly low density material apparently on the seafloor surrounding many islands and seamounts, probably consists of pyroclastics and volcanic sediments such as found in our Azores borehole transition sequence. The high density of the intrusive lamprophyric sheets could be a useful diagnostic parameter for seamounts or aseismic ridges formed by later intrusions into older seafloor, compared to those made up primarily by subaqueous lava flows, but the contrast probably is not sufficient to be detected by gravity measurements.

Porosity

Porosities (% pore volume) were determined by weighing the 2.5 cm dia. 5 cm long minicores seawater saturated (preserved in water from time of drilling) and after drying at abut 70°C under vacuum for two weeks. There was significant weight loss for the first 7 days but little thereafter. This treatment should remove most of the pore water except in very low permeability samples and it should produce only minor dehydration of hydrous minerals. However, we note that there probably is a continuum between free pore water and strongly bound water in hydrous minerals, some water being loosely bound. A small correction was made for the residual salt in the porosity estimate, assuming an original salinity of 35 ppt and that all of the salt was left behind on drying. We estimate the accuracy to be $\pm$ 10% to -5% of the value for porosity and water content.

The porosities of 20 samples from the Bermuda borehole range from 1.1 to 8.2% with a mean of 2.9 $\pm$ 0.3%. The mean for the lava flows of

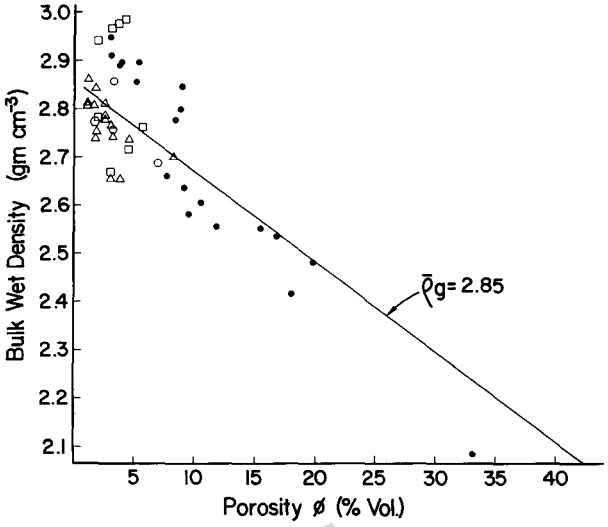


Fig. 5. Density as a function of porosity for Bermuda and Azores samples. (●) Azores subaerial; (O) Azores transition; (O) Azores subaqueous; (Δ) Bermuda flows; (□) Bermuda sheets.

2.6 ± 0.4% is lower than the mean of 3.4 ± 0.4% for the sheets. These values are significantly lower than the means for fresh young seafloor basalts, (e.g. 7.8%, Hyndman and Drury, 1976), indicative of some pore space filling by secondary minerals during hydrothermal alteration in the Bermuda rocks.

The porosities of 19 samples from the Azores borehole range from 3.0 to 33.1% with a mean of 9.4 ± 1.5%. The mean of the subaerial section is 10.6 ± 1.7%, while that for 3 subaqueous samples is 4.4 ± 1.2%, slightly higher than for Bermuda lavas and lower than young subaqueously extruded tholeiitic lavas. The porosity is presented as a function of depth for both boreholes in Fig. 6.

The permeabilities of 14 Bermuda samples by Boyle's Law gas expansion in a Hassler Holder were measured by Core Laboratories Ltd., Calgary. The all are less than 0.01 millidarcys the resolution limit of the method, except for one sample cut by a fracture with a value of 7.7. Low values are also suggested by the low drying rates. Of course water flow in the oceanic crust must be primarily in cracks and fractures. The permeability of the Azores subaerial rocks probably is much higher.

Seismic Velocities

Compressional and shear wave velocities were measured on 2.5 cm diameter, approximately 5 cm long water saturated samples to pressures of 6 kbars. The technique is essentially as described by Birch (1960) and Christensen and Shaw (1970). The accuracy estimated is 0.5% for compressional and ± 1.0% for shear wave velocities. The fluid pressure medium was excluded from the samples by a thin copper foil jacket and pore pressures were maintained at values much less than external pressure by placing 100 mesh screens between the samples and jackets.

The usefulness of laboratory measurements depends on how well the insitu conditions are simulated, notably the pressure, temperature, extent of water saturation and sample orientation if the rocks are anisotropic. It is also important to consider how representative the samples are of the section being investigated. In this study we wish to estimate the insitu velocities both in the sections penetrated by the boreholes and in other more general situations where the rocks may be similar such as in the upper few kilometers of an aseismic ridge. The velocities of most rocks increase significantly with increasing pressure, up to about 2.0 kbars (e.g. 7 km depth) primarily from the closing of microcracks and some pores. In addition to the effect of pressure insitu, the drilling process probably causes microcracks. These cracks should close at small confining pressure. We take the 0.4 kbar velocities to be representative of insitu values in the sections penetrated by the boreholes and the 2.0 kbar values as representative of depths greater than a kilometer in the crust. We

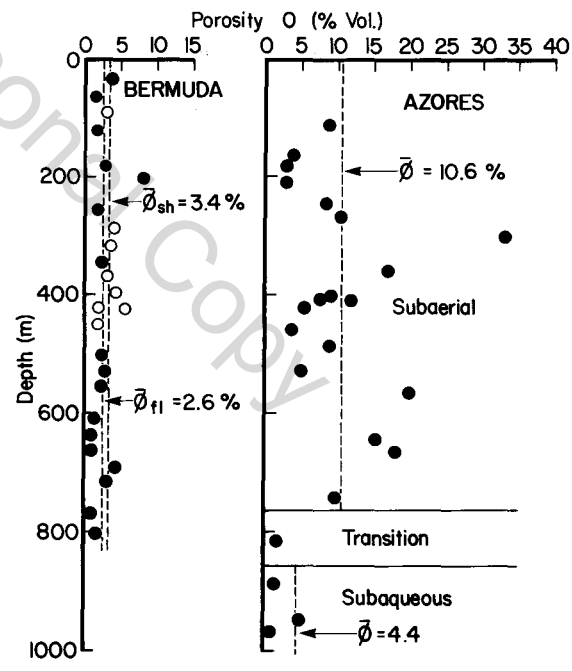
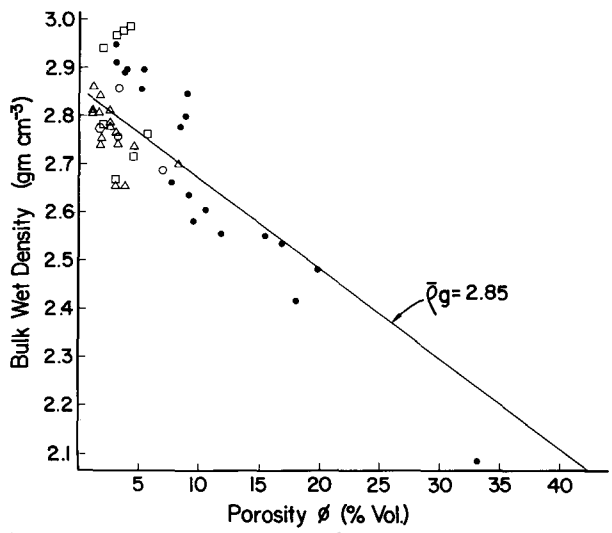


Fig. 6. Porosity as a function of depth in the Bermuda and Azores boreholes. Solid circles are Bermuda flows, open circles are Bermuda sheets.



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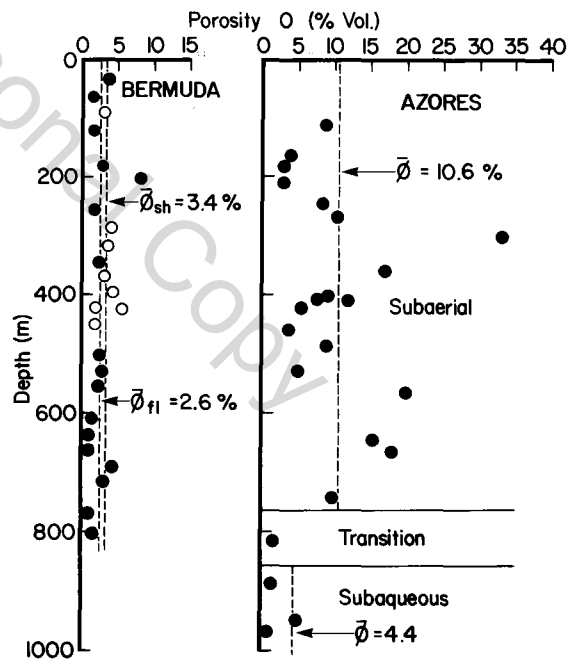

Fig. 6. Porosity as a function of depth in the Bermuda and Azores boreholes. Solid circles are Bermuda flows, open circles are Bermuda sheets.

TABLE 3. Mean Variations of Velocity with Pressure

	Pressure (Kbar) 0.2	0.4	0.6	0.8	1.0	2.0	4.0	6.0
	Bermuda Lava Flows (N = 12)							
Vp	5.95 ±.27	5.97 ±.27	6.00 ±.27	6.02 ±.27	6.03 ±.27	6.09 ±.26	6.17 ±.25	6.23 ±.24
Vs	3.20 ±.20	3.22 ±.20	3.23 ±.19	3.23 ±.19	3.24 ±.19	3.27 ±.18	3.30 ±.18	3.32 ±.17
σ	.296 ±.011	.295 ±.011	.296 ±.010	.297 ±.011	.294 ±.011	.298 ±.010	.299 ±.011	.301 ±.012
	Bermuda Lamprophyric Sheets (N = 8)							
Vp	5.86 ±.22	5.89 ±.21	5.92 ±.20	5.94 ±.20	5.96 ±.20	6.03 ±.20	6.13 ±.20	6.19 ±.20
Vs	3.05 ±.15	3.07 ±.15	3.08 ±.15	3.10 ±.15	3.11 ±.15	3.16 ±.15	3.21 ±.16	3.24 ±.17
σ	.314 ±.018	.314 ±.016	.314 ±.016	.313 ±.018	.313 ±.016	.311 ±.016	.310 ±.016	.311 ±.015
	Azores Subaerial (N = 15)							
Vp	4.88 ±.67	4.92 ±.66	4.95 ±.65	4.98 ±.65	5.00 ±.64	5.07 ±.62	5.21 ±.56	5.33 ±.52
Vs	2.45 ±.48	2.49 ±.46	2.53 ±.44	2.55 ±.43	2.57 ±.42	2.64 ±.38	2.73 ±.34	2.77 ±.31
σ	.332 ±.032	.328 ±.028	.325 ±.024	.324 ±.022	.319 ±.020	.314 ±.015	.311 ±.015	.313 ±.013
	Azores Subaqueous (N = 3)							
Vp	5.58 ±.30	5.60 ±.31	5.62 ±.31	5.63 ±.31	5.64 ±.31	5.70 ±.32	5.77 ±.29	5.83 ±.27
Vs	2.96 ±.17	2.97 ±.16	2.98 ±.16	2.98 ±.16	2.99 ±.15	3.01 ±.15	3.04 ±.15	3.06 ±.15
σ	.303 ±.006	.307 ±.006	.307 ±.006	.303 ±.006	.303 ±.006	.307 ±.006	.310 ±.000	.310 ±.000

stress that laboratory pressure can be related only approximately to equivalent depth in the crust. In our laboratory samples we have attempted to maintain close to zero pore pressure (see Hyndman and Drury, (1976) for a discussion of the importance of pore pressure), so the effective confining pressure is close to the external fluid pressure. The mean effect of pressure on velocities of Bermuda and Azores samples is given in Table 3 and Fig. 7.

Increasing temperature decreases the velocities of most rocks. Only a few measurements have been reported and over very limited temperature ranges. The effect on mafic rocks to 300°C ranges between -0.6×10^{-4} and -15×10^{-4} per °C (Birch, 1958; Hughes and Maurette, 1957; Nafe and Drake, 1968; Kroenke et al., 1976). Recent detailed measurements on basalts from the East Pacific Rise show -5×10^{-4} to be a representative value for oceanic basalts (Christensen, in preparation). Thus, the difference from a laboratory temperature of 25°C and the 200°C in the lower part of the Azores borehole (Fig. 3) requires a correction to the measured values of -0.1 km s^{-1}. The correction is negligible for the Bermuda borehole where the maximum is about 40°C (Hyndman et al., 1974a). The temperature in the crust under aseismic ridges and other old volcanic islands will rarely exceed the 200°C found in the Azores borehole. The temperature may be significantly higher at depth under active volcanic islands, requiring a greater correction.

All of the measured samples have been water saturated before measurement to simulate insitu conditions (Christensen and Salisbury, 1975). Velocities were measured in the vertical direction only since previous measurements have shown negligible anisotropy in extrusive oceanic volcanic rocks (e.g. Christensen and Shaw, 1970; Christensen and Salisbury, 1972).

Relations Between Velocity and Other Parameters

The relations between compressional and shear velocities and bulk density are important because they permit the estimation of density from seismic refraction data. Birch (1961) showed that there is one nearly linear velocity-density relation for all rocks with a particular mean atomic weight. The mean atomic weight of unaltered basaltic rocks varies only slightly so a single relation is expected. The relation might be different for altered rocks since the high water content associated with alteration will lower the mean atomic weight significantly perhaps without a corresponding decrease in bulk density. Christensen and Salisbury (1975) found a better fit with non-linear relations.

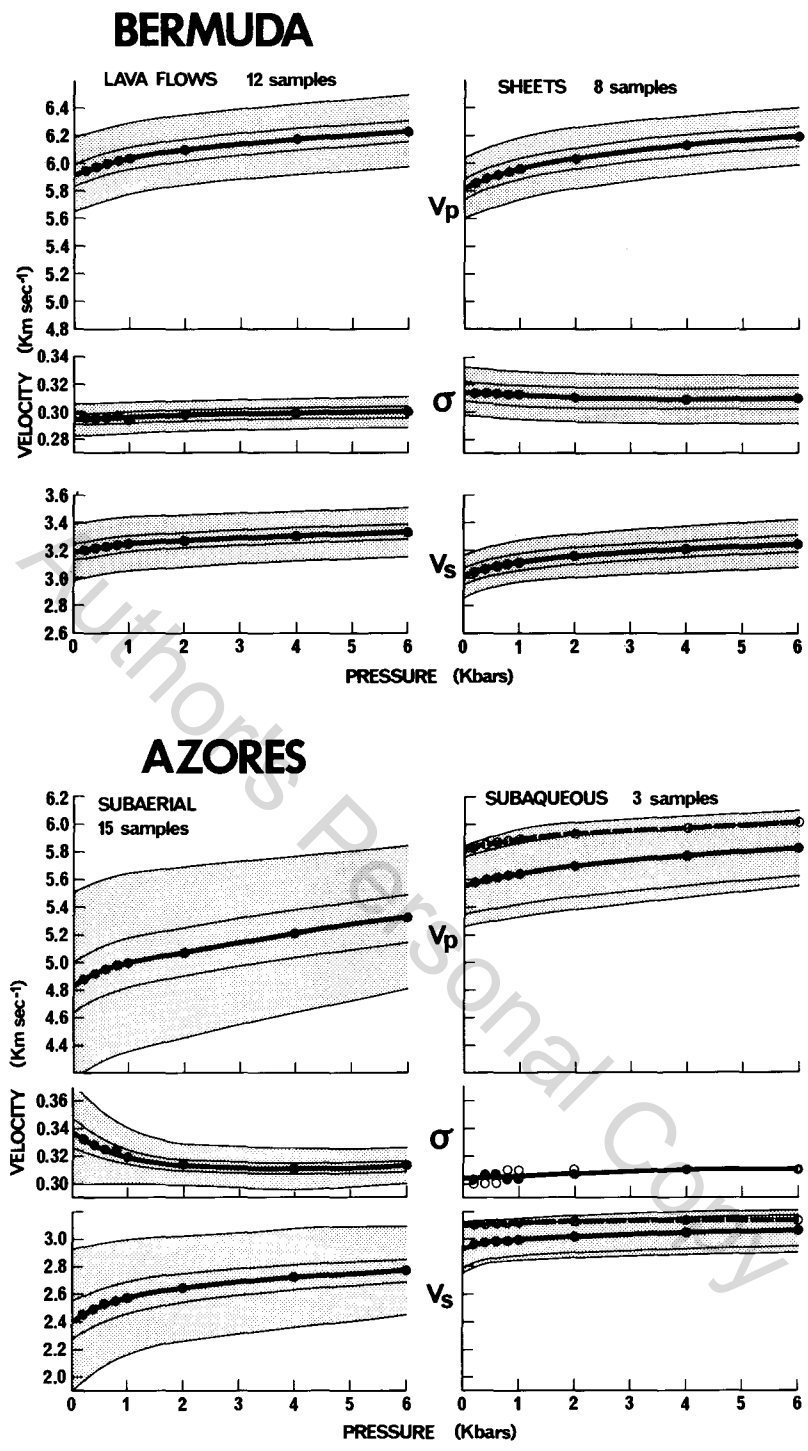

Fig. 7. Compressional and shear wave velocity as a function of pressure in the Bermuda and Azores samples. The inner bounds are the standard error estimate of the mean, the outer bounds are the standard deviation. The open circles are the Azores transition sample.

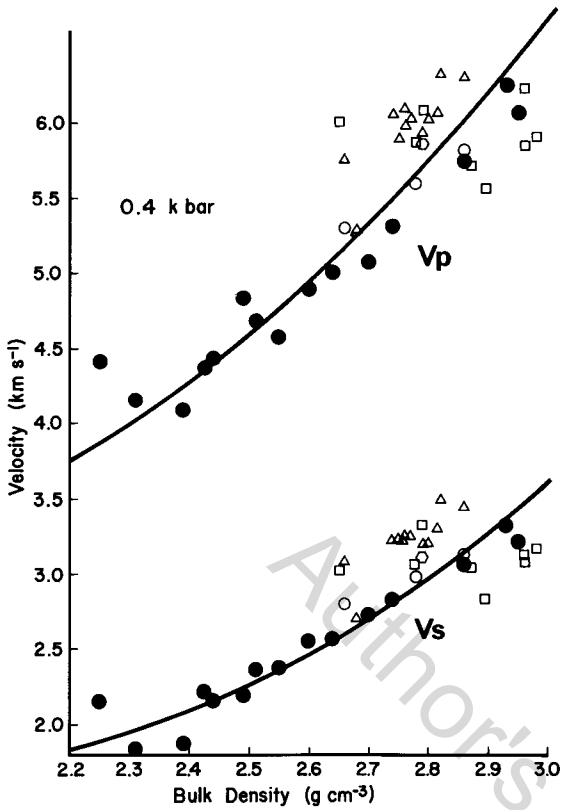

Fig. 8. Compressional and shear wave velocity as a function of bulk density for Bermuda and Azores samples. Symbols as in Fig. 5. The curves are from Christensen and Salisbury (1975).

The relation for the Bermuda and Azores samples is shown in Figure 8 along with the non-linear best fits for 77 sea floor basalts from the Deep Sea Drilling Project given by Christensen and Salisbury. An additional 79 samples from deep holes on the mid-Atlantic ridge fit their relations well although the shear velocities for the latter are slightly higher for a particular density (Hyndman and Drury, 1976). Only the Azores subaerial volcanics have a significant range of density but they define similar relations. The one transition and 3 subaqueous Azores samples also lie close to the lines. The lamprophyric sheet samples from Bermuda generally fit the relation although they scatter widely. In contrast, all of the altered tholeiitic lavas show much less scatter and have both compressional and shear velocities averaging 0.3 km s^{-1} above the relation of Christensen and Salisbury (1975). This difference may reflect a lower mean atomic weight produced by the extensive hydrothermal alteration (see Aumento and Ade-Hall, 1973), and possibly the filling of vesicles with alteration products. The contrast between the Bermuda lavas and fresh sea floor tholeiites is also evident, for example, in the high thermal conductivity (see below). It is important to note that the Azores subaerial samples, which have low velocities primarily because of high porosity in flows and pyroclastics, have the same velocity-density relation as the deep ocean samples reported by Christensen and Salisbury (1975) in which low velocities arise primarily from low temperature weathering. Thus, the relation should be valid for the upper oceanic crust, of any age of either normal ridge origin, or an origin associated with a hot spot.

The relation between Poisson's ratio (a measure of the ratio of compressional to shear velocity) and compressional velocity for Bermuda and Azores samples is shown in Figure 9. Only the Azores subaerial samples have a sufficient range to define a trend. Poisson's ratio increases systematically with decreasing velocity below about 5.0 km s^{-1}, in agreement with the general trend found for many seafloor tholeiitic basalts obtained by dredging or shallow drilling (Christensen, unpublished data; Hyndman, 1979). But the trend is in contrast to that of increasing Poisson's ratio with increasing velocity found by Hyndman and Drury (1976) for young fresh basalts from Deep Sea Drilling Project, 0.5 km deep holes on the mid Atlantic ridge, in the velocity range 5.5 to 6.5 km s^{-1}. The opposite dependences probably reflect the different nature of the porosity structure associated with the velocity variation. The young fresh tholeiites from the mid Atlantic ridge deep holes have porosity primarily in nearly spherical vesicles that are poorly connected, while in the Azores subaerial samples the porosity is primarily along well connected grain boundaries (see discussion in Hyndman, 1979).

On a plot such as Figure 10 (see below) the mid-Atlantic ridge rocks have a mean pore aspect ratio of 1/2 compared to about 1/10 for Azores subaerial volcanics. Two highly vesicular basalts from the Lau ridge with a 5.0 km s^{-1} compressional velocity (Christensen and Salisbury, 1975) also have a low Poisson's ratio of 0.27. In contrast, the low velocity samples from dredging or shallow drilling have porosity primarily from low temperature weathering which occurs along grain boundaries giving well connected spaces. Similarly, the low velocity Azores subaerial volcanics have well interconnected porosity, particularly in the pyroclastic sections and well fractured flows. We thus emphasize that variations in Poisson's ratio of unaltered basaltic rocks with velocity probably arise mainly from the porosity structure rather than from differences in mineralogy. Alteration products such as chlorite have high Poisson's ratios so Poisson's ratios of basalts may increase with very extensive hydrothermal alteration that also decreases the compressional velocity.

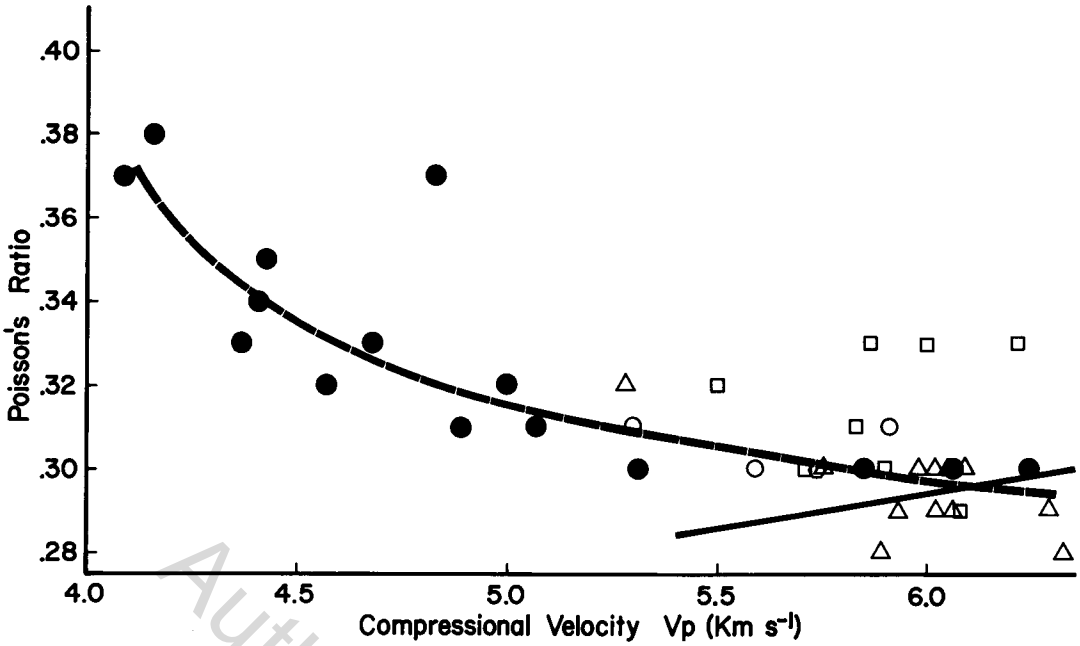

Fig. 9. Poisson's ratio as a function of compressional velocity for Bermuda and Azores samples at 0.4 kbar pressure. Symbols as in Fig. 5. The dashed line illustrates the general trend. The solid line is the relation of Hyndman and Drury (1976) for basalts from deep drilling on the mid-Atlantic ridge.

Figure 10 shows the relation between compressional velocity and porosity for the samples. A knowledge of the porosity and its form in the upper oceanic crust and in islands is important for an understanding of alteration processes, and for rough estimates of the permeability available for hydrothermal circulation processes (e.g. Francis, 1976; Whitmarsh, 1978). The relation may be useful for the estimation of upper crustal porosities from seismic refraction measurements, although much of the upper crustal pore space probably is in large fractures and voids so is not represented by small samples. The relation of Figure 9 is largely based on the data for the Azores subaerial rocks. A good correlation has been expected from the relation between velocity and density (Fig. 8) and the relation between density and porosity (Fig. 5). Figure 10 also shows the predicted velocity-porosity relations for various pore aspect ratios (or elongations) from the non-interaction theory of Toksöz et al. (1976) as applied to the upper oceanic crust by Whitmarsh (1978). A mean aspect ratio of 1/10 gives the best fit in rough agreement with the large scale upper crustal values suggested by Whitmarsh.

Mean Elastic Parameters For Drilled Sections

Compressional and shear wave velocities for Bermuda samples as a function of depth are shown in Figure 11. There is no systematic variation. There are, however significant differences in both velocities and in Poisson's ratio between the lava flows and lamprophyric

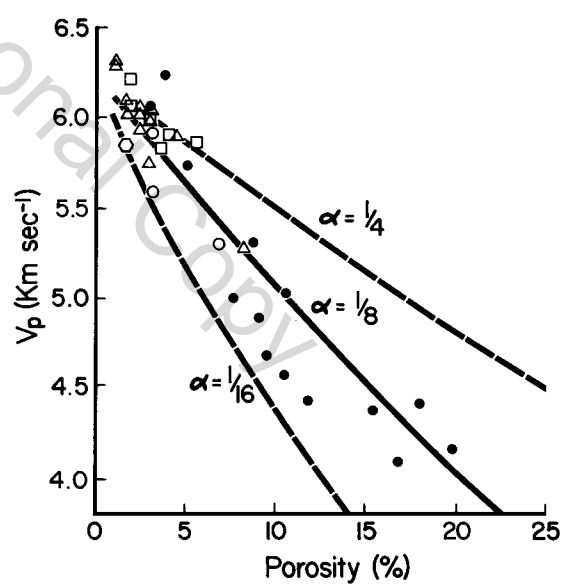

Fig. 10. Compressional velocity at 0.4 kbar pressure as a function of porosity. Symbols as in Fig. 5. The curves are the predicted relations for different pore aspect ratios from Whitmarsh (1978).

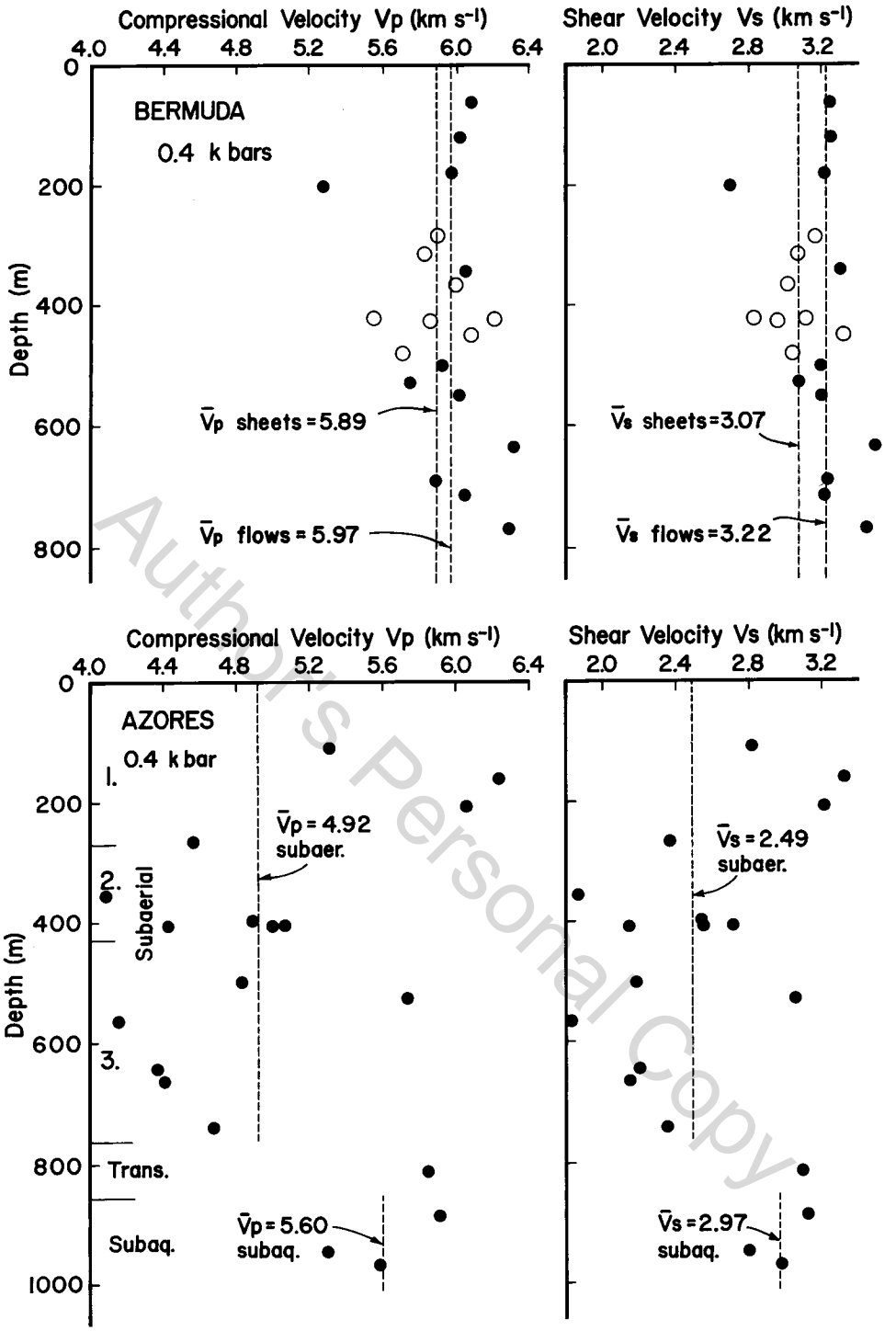

Fig. 11. Compressional and shear wave velocities at 0.4 kbar pressure as a function of depth in the Bermuda and Azores boreholes.

sheets. The mean compressional velocities are 5.97 ± 0.08 and 5.89 ± 0.07, mean shear velocities 3.22 ± 0.06 and 3.07 ± 0.05 km s^{-1} respectively. The mean Poisson's ratios are 0.295 ± 0.003 and 0.314 ± 0.006 (Table 1). The velocities for the flows are higher and Poisson's ratios lower than for the sheets even though the flow densities are lower (Fig. 4). Christensen and Salisbury (1975) show how appropriate differences in mineralogy can affect velocity and Poisson's ratio. The mean velocities for the lavas are higher than for most sea floor samples that have been measured, reflecting the low porosity of these hydrothermally altered rocks. The mean velocities and Poisson's ratio of the flows are almost identical to those for samples from deep drilling on the mid-Atlantic ridge (Hyndman and Drury, 1976). The mean compressional velocity of 5.94 km s^{-1} for the drilled section is somewhat higher than the velocity of 5.1 km s^{-1} estimated for under the Bermuda platform by Woollard (1954).

The compressional and shear wave velocities for the Azores samples as a function of depth are shown in Figure 11. There is no systematic variation with depth in the subaerial section which has a mean compressional velocity of 4.92, mean shear velocity of 2.49 km s^{-1} and mean Poisson's ratio of 0.328. The scatter of all parameters is large. The high Poisson's ratios are associated with low velocity (Fig. 9). The velocities of the 3 subaqueous samples are significantly higher than those for the subaerial samples, the mean compressional velocity being 5.60, mean shear velocity 2.97 km s^{-1} and Poisson's ratio 0.307. The latter is typical of the higher velocity subaerial samples, but is slightly higher than for either fresh sea floor basalts (e.g. Hyndman and Drury, 1976) or the hydrothermally altered Bermuda subaqueous lava flows of 0.295. The mean velocities of the Azores subaqueous samples also is slightly lower than for the latter two rock types. One sample from the transition zone has similar values to the subaqueous rocks.

Electrical Resistivity

The electrical resistivity of the Bermuda and Azores minicores were measured by applying a 0.5V, 10 or 50 Hz signal to the ends which had been painted with electrically conducting epoxy resin (see Hyndman and Drury, 1976). The precision of the measurements is better than 1% but variations in surface water makes the values reproducible only to about ± 10%. All of the data reported are for samples saturated with seawater except for one Bermuda sample measured dry to high temperature. The rocks beneath Bermuda and the Azores are undoubtedly water saturated. Seawater salinity probably is a good approximation for beneath Bermuda and for the subaqueous section penetrated by the Azores borehole. The salinity of the insitu fluid in the Azores subaerial section might be lower than that of seawater so that our measured resistivities may be too low. In order to relate the measured resistivities to variable insitu conditions and to investigate conduction mechanisms, the variation of resistivity with porosity, pressure and temperature were measured on a number of samples. The resistivity change with different degrees of dehydration and with varying signal frequency for some Bermuda and Azores cores have been presented by Drury (1977).

Resistivity as a Fuction of Porosity

The resistivity of rocks for which conduction is primarily through fluid filled pore spaces, is found to correlate closely with the rock porosity. The empirical relation is (Archie, 1942):

$$\rho = A\rho_f \, \emptyset^{-n}$$

where ρ_f is the resistivity of the fluid, $\emptyset$ is the porosity and A and n are constants depending primarily on the extent of interconnection between the pore spaces. A is usually close to 1. If the porosity consists of randomly spaced spherical pores n is about 2. (Brace et al., 1965; Brace and Orange, 1968; Shankland and Waff, 1974). Values of n greater than 2 are possible if the pores are less than randomly interconnected.

The resistivity (at atmospheric pressure and 22°C) of all Bermuda and Azores samples correlates well with porosity (Fig. 12) with an n value of 2.4, and a seawater resistivity of 0.2 ohm-m, in agreement with thin sections which indicate most of the porosity to be in roughly spherical pores perhaps with less than random connection. Deep sea floor basalts also have shown high n values of 2.5 (Hyndman and Drury, 1976) suggesting less than random interconnection. The correlation clearly implies conduction primarily through pore fluid in the Bermuda and Azores rocks.

Resistivity as a Function of Pressure

The effect of pressure on resistivity provides important diagnostic information on rock conduction mechanisms. If pore fluid conduction dominates, resistivity should increase with increasing pressure as microcracks and pores are closed, reducing the available conduction paths. If conduction through mineral grains dominates, resistivity should decrease with increasing pressure, as the conducting grains are brought into more intimate contact (e.g. Brace and Orange, 1968). The effective pressure in either situation is the confining pressure, i.e. the external pressure minus the internal pore pressure.

The problem of relating laboratory sample confining pressure to depth in the crust is discussed by Hyndman and Drury (1976).

The variation of resistivity of seawater saturated samples from Bermuda was measured to 2 kbars pressure at 20 to 22°C. The sample cylinders are placed between two stainless steel electrodes of the same diameter and the assembly encased in heat-shrink plastic tubing to exclude the hydraulic fluid pressure medium. The electrodes have perforations leading to small reservoirs, that hold the fluid squeezed from the sample with increasing pressure. Thus, the pore pressure is close to zero and the confining pressure close to the external pressure.

The resistivity increases with increasing pressure for all of the samples measured indicating that pore fluid conduction is the dominant mechanism. The increase is by about 50% to a pressure of 0.5 kbar and 100% to 2.0 kbar (Fig. 13).

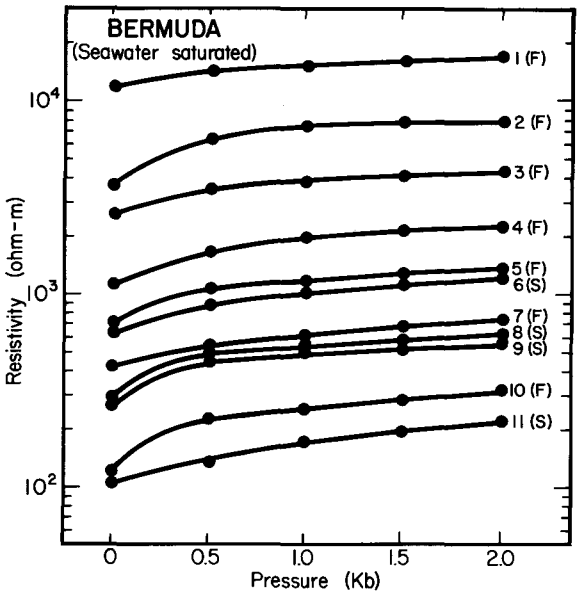

Fig. 13. Electrical resistivity as a function of pressure for Bermuda samples, at 23°C and seawater saturated.

Resistivity as a Function of Temperature

The resistivity of most rocks changes rapidly with temperature. If conduction is through pore fluid the temperature dependence expected is of the form exhibited by the fluid itself. At moderate temperatures seawater resistivity decreases with increasing temperature (e.g. Horne, 1969). Its behaviour at higher temperatures probably is similar to NaCl solutions which have decreasing resistivity to about 300°C because of the increasing ionic mobility from decreasing viscosity. At still higher temperatures resistivity increases because of a decrease in ionic concentration due to density and association effects (e.g. Quist and Marshall, 1968). At very high temperatures or for dry rocks semiconduction through the mineral grains may dominate. The resistivity is then given by

$$\frac{1}{\rho} = \frac{1}{\rho o} e^{-E/kT}$$

where ρ_o is the limiting resistivity at very high temperature, E is the activation energy (in electron volts) and K is Boltzmann's constant. Previous measurements of sea floor basalts have shown a more rapid decrease in resistivity with temperature than for sea water or saline solutions even though pore fluid conduction in the basalts was indicated by the dependence on porosity and on pressure. The effect might arise through conduction partially through clay minerals (Hyndman and Drury, 1976).

The resistivity of seawater saturated basalts from Bermuda at a pressure of 0.3 kbar was measured up to 250°C (Fig. 14). One sheet

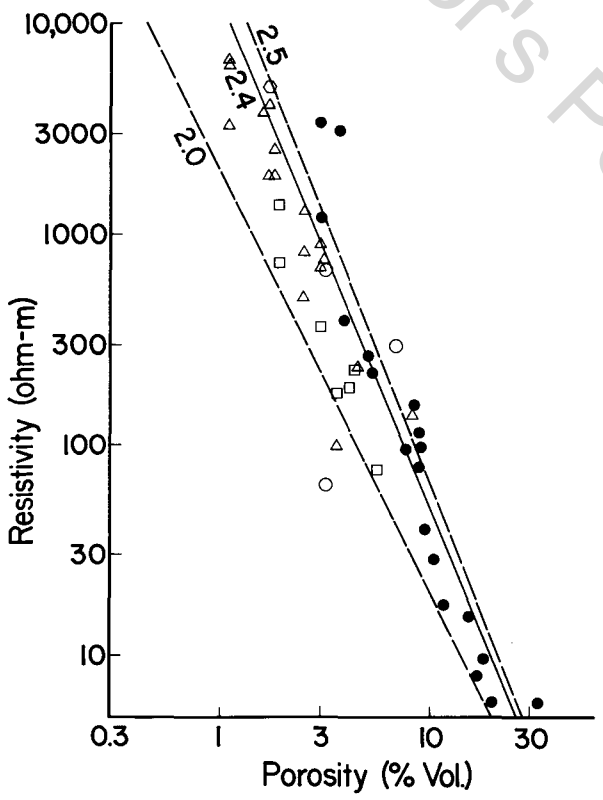

Fig. 12. Electrical resistivity as a function of porosity for Bermuda and Azores samples. Symbols as in Fig. 5. The lines are for different values of the exponent in Archie's Law.

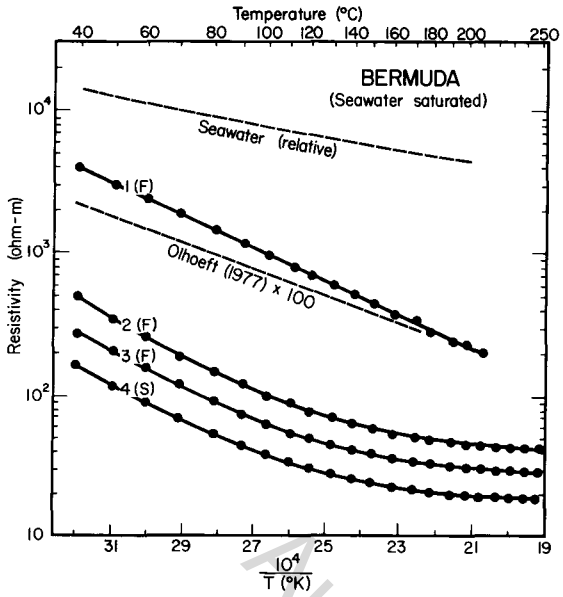

Fig. 14. Electrical resistivity as a function of temperature for Bermuda samples at 0.3 Kbar pressure, seawater saturated. The sample measured by Olhoeft (1977) had 14% porosity and was saturated with 1.7 ohm-m NaCl solution.

and two flow samples show a more rapid decrease in resistivity with increasing temperature than seawater at low temperatures, becoming similar to seawater above about 150°C. We suggest clay mineral conduction may be important in these samples (see Drury, 1977). One lava flow sample exhibits a linear decrease of log resistivity with inverse of temperature which suggests a semiconduction primary mechanism. This is a low porosity, high resistivity sample.

Clay minerals may be important to conduction in some basaltic rocks, producing through ion exclusion and cation exchange mechanisms, the more rapid change in resistivity with temperature than expected for pore fluid conduction. A more complex interaction appears to exist among pore fluid, clay, and mineral grain conduction to produce this behaviour (see the discussion in Drury, 1977). Clay mineral conduction probably occurs but is less pronounced for subaerial basalts than for subaqueous tholeiite pillows lavas. Olhoeft (1977) also found that some process in addition to simple pore fluid conduction was necessary to explain the effect of temperature on the resistivity of a Hawaiian tholeiitic basalt. (See Fig. 14).

One Bermuda sample was measured dry to 400°C (Fig. 15). It exhibits a linear dependence indicating semiconduction in the mineral grains. All of the pore water should have been removed and the clays dehydrated between 100 and 200°C (e.g. Deer et al., 1966). The activation energy is 0.5 eV and $\rho_o = 10^{-2}$ ohm-m, close to the values of 0.34 to 0.99 eV and 10^{-1} to 10^3 ohm-m found for deep ocean tholeiites by Schloessin and Dvorak (1976) to 1000°C. In the ocean crust mineral semiconduction probably dominates above about 300°C, if such temperatures occur. Pore fluid and clay mineral conduction dominate at lower temperatures.

Mean Resistivities for Drilled Sections

The mean resistivity of 23 Bermuda samples at 22°C and atmospheric pressure is 780 ohm-m with no trend with depth (Fig. 16). The mean for 17 lava flows is 1320 ohm-m and for 6 sheets is 340 ohm-m. The mean resistivity for the sheets is slightly lower reflecting their higher mean porosity. The mean resistivity is significantly higher than for deep ocean tholeiites recovered by the Deep Sea Drilling Project which have a mean of about 250 ohm-m (Hyndman and Ade-Hall, 1974; Hyndman and Drury, 1976; Drury, 1977). The higher resistivity of the Bermuda rocks reflects their reduced porosity resulting from hydrothermal alteration. The insitu mean resistivity of the drilled Bermuda section to 800 m depth should be close to the 800 ohm-m measured since the slightly higher average insitu temperature of about 30°C (Hyndman et al., 1974a) 8°C above

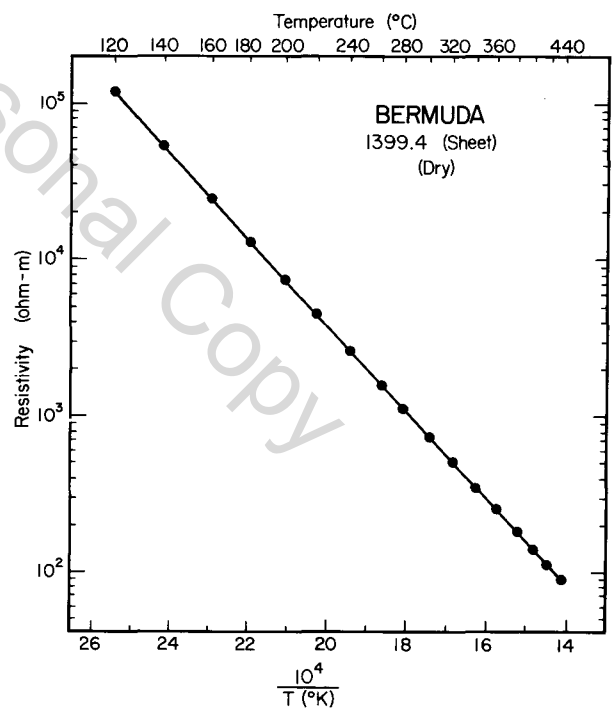

Fig. 15. Electrical resistivity as a function of temperature for one dry Bermuda sample at atmospheric pressure.

the laboratory temperature, will be about offset by the higher insitu pressure. At greater depth the resistivity should be significantly lower, the effect of the increasing temperature, 22°C km^{-1}, being more important than the effect of increasing pressure. At a depth of 5 km similar rocks would have a resistivity of about 200 ohm-m.

The mean resistivity of 23 Azores samples at 22°C and atmosphere pressure is 110 ohm-m. There is a general increase with depth (Fig. 16). The mean for 19 samples of the subaerial sequences is 82 ohm-m, for one transition zone sample it is 4900 ohm-m, and for 3 subaqueous samples it is 230 ohm-m. The very low resistivities of the subaerial rocks simply reflects their very high porosity. The subaqueous pillow lavas have resistivities very close to the mean for deep sea samples. The mean insitu resistivity of the drilled subaerial section should be about 5 ohm-m, significantly lower than the laboratory mean because of the much higher mean insitu temperature of about 140°C (Muecke et al., 1974). The subaqueous rocks under insitu conditions will have a resistivity of about 50 ohm-m. The temperature probably increases only slowly at depths greater than that penetrated by the borehole, unless a recent intrusion is approached, so the resistivity will be only slightly less than the 50 ohm-m,

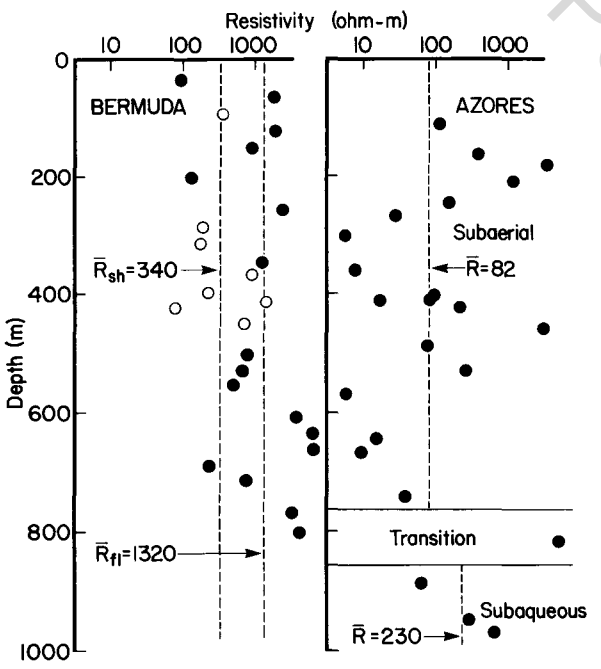

Fig. 16. Electrical resistivity as a function of depth in the Bermuda and Azores boreholes, for samples at 23°C and atmospheric pressure, seawater saturated. The dashed lines are the geometric means.

perhaps 10 ohm-m at 5 km, for similar pillow lavas.

Thermal Conductivity

The thermal conductivity of samples from the Bermuda and Azores boreholes was measured on 32 mm diameter, 10 mm thick disks with a divided bar instrument of the type described by Jessop (1970). The samples were water saturated, at a mean temperature of 25°C and subject to an axial pressure of 25 bars. Corrections were applied for disk-bar contact resistance, lateral heat loss and small variations in disk diameter. Calibration was with crystalline quartz and fused silica using the conductivity values of Ratcliffe (1959). Measurements on individual disks were reproducible to ± 2% and we estimate the accuracy to be better than ± 5%.

The mean thermal conductivity of 53 samples from the Bermuda core (Fig. 17), 32 flows and 21 sheets is 2.12 ± 0.03 W m^{-1} K^{-1} (5.07 ± 0.07 mcal cm^{-1} s^{-1} °C^{-1}). The mean of sheets of 2.19 ± 0.07 (5.23 ± 0.16) is significantly higher than the 2.08 ± 0.02 (4.96 ± 0.05) for the flows. The variation indicated by the standard deviation of 0.31 (0.74) also is much greater than that for the flows of 0.13 (0.31). The conductivity for the sheets appears to reflect primarily the content of pyroxene which has relatively high conductivity (e.g. Clark, 1966). The mean for the flows is also much higher than the 1.7 (4.0) representative of most previously measured fresh oceanic tholeiitic lavas (e.g. Hyndman and Drury, 1976). The higher conductivity in this case appears to be associated with the extensive hydrothermal alteration of the Bermuda lavas that probably occurred at the time of the intrusion of the lamprophyric sheets.

The mean thermal conductivity of 40 samples from the Azores borehole (Fig. 16) is 1.75 ± 0.04 W m^{-1} K^{-1} (4.19 ± 0.10 mcal cm^{-1} s^{-1} °C^{-1}). The mean of 31 subaerial samples is 1.75 ± 0.04 (4.18 ± 0.11) and of 5 subaqueous lavas is 1.73 ± 0.04 (4.13 ± 0.10). The subaerial volcanics have a similar conductivity to that for similar rocks from Hawaii with porosity from 5 to 10% (Robertson and Peck, 1974). The large variation (s.d. = 0.25 (0.59)) reflects primarily large variations in porosity, although the composition also is highly variable, (Muecke et al., 1974). The subaqueous lavas have a much smaller scatter (s.d. = 0.09 (0.22)). The mean is slightly higher than for most fresh tholeiitic flows, reflecting the significant hydrothermal alteration which was actually still in progress at the time of drilling (see temperature profile, Fig. 3). We conclude from both the Bermuda and Azores cores that hydrothermal alteration increases the

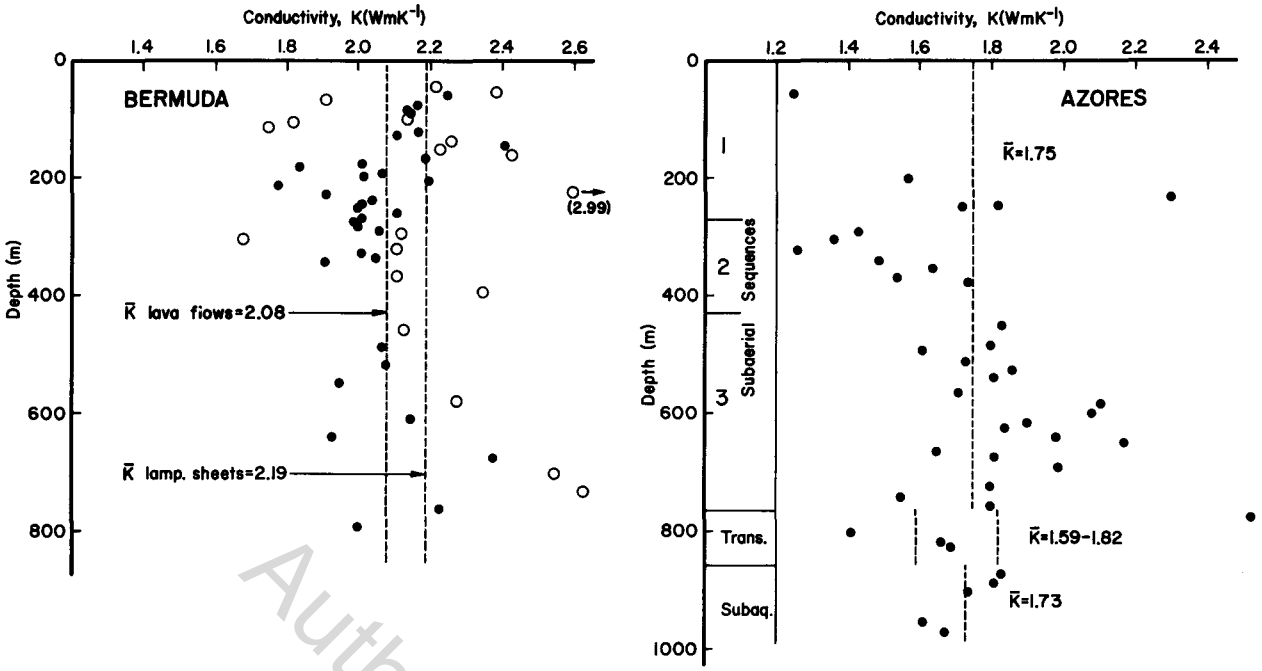

Fig. 17. Thermal conductivity as a function of depth for the Bermuda and Azores boreholes. The dashed lines are the simple arithmetic means.

conductivity of submarine tholeiitic lavas. In contrast, low temperature weathering appears to decrease the conductivity, in one case to 1.55 (3.7) (Hyndman et al., 1974b; Hyndman and Drury, 1976).

The effect of pressure on the thermal conductivity of basalts is quite small and can be neglected at least for upper crustal pressure of 0.5 kbar. The effect of temperature is significant being a decrease of about 0.054 (0.13) per 100°C increase (see review of data in Hyndman and Drury (1976)) implying a mean insitu conductivity at the bottom of the Azores borehole where the temperature is 200°C, of 1.64 (3.92). Such high temperatures and thus low thermal conductivity will occur locally in the upper crust of many young areas. The temperature effect can be neglected in old areas such as Bermuda at least for the upper few kilometers of the crust. The insitu conductivity may also be slightly lower than the sample means because of fractured sections of low conductivity, not represented in the core. In our study, we think these zones are important only for the upper part of the Azores borehole.

Conclusions

The physical properties of basalt samples from boreholes into the islands of Bermuda and the Azores (Tables 1 and 2) are generally similar to those from the deep ocean floor. The most significant differences are for samples from the Azores upper subaerial section. They have high porosity and as a consequence lower density, lower seismic velocities, higher Poisson's ratio, lower electrical resistivity and lower thermal conductivity as compared to deep ocean tholeiitic basalts. Mineralogical differences affect the measured physical properties but much less than differences in porosity. The lamprophyric sheets in the Bermuda section have higher density, lower seismic velocities, higher Poisson's ratio, and higher thermal conductivity than the lava flows from Bermuda or from the deep ocean floor. The hydrothermal alteration of the Bermuda lava flows appears to have reduced their density, particularly the grain density, reduced their porosity, increased their electrical resistivity and increased their thermal conductivity. This is assuming that their original properties before alteration were similar to samples from the deep ocean crust. The hydrothermal alteration apparently has not significantly affected the seismic velocities. The subaqueous Azores samples exhibit a similar but smaller difference compared to deep ocean lavas, probably from less pronounced hydrothermal alteration at the present 200°C insitu temperature.

The only physical properties of those measured, that are different for aseismic ridges and volcanic island seamount chains or 'hot spot traces' compared to normal oceanic

crust, that might be detected by surface geophysical measurements are the slightly higher density and higher Poisson's ratio compared to the surrounding oceanic crust, suggested by the Bermuda samples.

Acknowledgements. We wish to acknowledge the efforts of Dr. F. Aumento who was primarily responsible for the drilling effort in Bermuda and Dr. J.M. Hall who organized the drilling in the Azores. Dr. M.J. Keen supported and encouraged both projects. D.L. Barrett undertook some of the early velocity measurements on the Bermuda core and was of considerable assistance in the setting up of some of the measurement equipment. J. Hull assisted in the velocity measurements and analysis of the data. The Bermuda drilling was financed by Dalhousie University, the National Research Council of Canada and by the Lamont-Doherty Geological Observatory. Financial support for the Azores drilling came from the U.S. National Science Foundation (International Decade of Ocean Exploration), the Research Corporation, the National Research Council of Canada, Dalhousie University, and the Empressa Insular de Electricidade (Ponta Delgada) S.A.R.L. Financial support for the measurements was provided by the Office of Naval Research Contract N-00014-75-C-0502 and the National Research Council of Canada. The drilling and some of the measurements were undertaken while R.D. Hyndman and M.J. Drury were at Dalhousie University.

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