The Mean Density of The Oceanic Crust

by

Greg Steven Raskin

Geophysics

Submitted in Partial Fulfillment of the Requirements of the

University Undergraduate Fellows Program

1982-1983

Approved by:

Richard L. Carlson

Richard L. Carlson

May 1983

ABSTRACT

The Mean Density of the Oceanic Crust. (May 1983)

Greg Steven Raskin

Faculty Advisor: Dr. Richard L. Carlson

There now exist sufficient numbers of recovered marine and ophiolitic samples to correlate their measured velocities and densities with marine seismic survey data in order to determine the mean density of the oceanic crust. Two methods are proposed: 1) assigning general lithologies to the various seismic layers of the crust based upon the ophiolite model, and 2) determining a velocity-density relationship for oceanic crustal rocks and subsequently converting the seismic velocity structure to density structure.

The mean densities and velocities of unaltered basalts, dolerites, and gabbros have the following values: 2.82±0.09 g/cm^3, 5.88±0.38 km/s; 2.84±0.08 g/cm^3, 6.38±0.44 km/s; and 2.92±0.09 g/cm^3, 7.05±0.32 km/s, respectively. Applying these mean densities to eight different layered models yields a density of the igneous oceanic crust of 2.90 g/cm^3. This is essentially a first order approximation and does not include formation porosity.

In determining a velocity-density relationship for crustal rocks, it was necessary to use two velocity ranges. For velocities less than 6.65 km/s (the grain velocity of basalt), velocity and density were calculated parametrically as functions of porosity. And, for velocities greater than 6.65 km/s, a linear regression of

the type $\rho = A + B/V_p$ was fit to all samples in order to reflect changes in lithology and grain size. Applying these equations to the layer velocities of the aforementioned models yields a mean crustal density of 2.89 g/cm^3.

ACKNOWLEDGEMENTS

I wish to thank my friend and advisor Dr. Richard L. Carlson for giving me this opportunity.

Thanks are also due Dr. Michael Brown for his help and support.

I also with to give special thanks to my friends James Rayhorn and Kenneth Wilks for many helpful suggestions and comments during the year.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

INTRODUCTION

The vast majority of available information concerning the oceanic crust at depth has been obtained from seismic refraction surveys. Early studies modeled the crust as having a simple three layer structure. Raitt [1963] determined the average layer velocities and thicknesses which are given in Figure 1. Dredge hauls conducted along ridge crests and fracture zones recovered basalts, gabbros, and serpentinites as well as the sediments known to constitute layer one. Various models of the oceanic crust were proposed, concluding that layers two and three were composed of basalt and either serpentinized peridotite or gabbro respectively. However, shear and compressional wave velocity data from laboratory samples could not strongly support any model [Christensen and Salisbury, 1975].

Considerable interest has been generated by recent studies of ophiolites. Many authors have noted the strikingly similar seismic profiles of ophiolites and the oceanic crust [Christensen and Salisbury, 1975; Spudich and Orcutt, 1980]. In addition, the lithological structures encountered, as well as their formational histories, conform to known crustal formation processes. Indeed, these same studies conclude that ophiolites do represent a valid model of the oceanic crust.

This thesis follows the format of Journal of Geophysical Research.

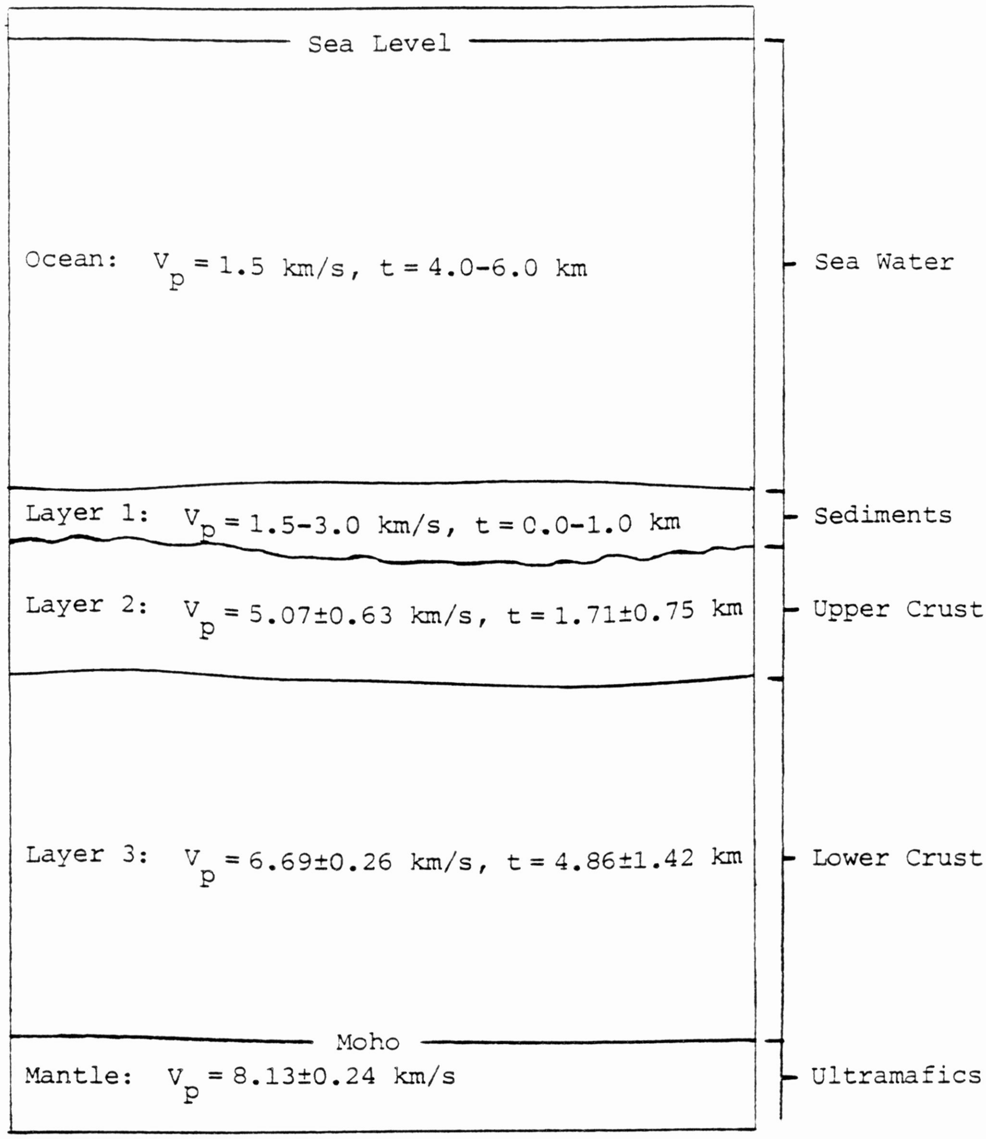

Fig. 1. Average oceanic crust. Layer thicknesses, velocities, and uncertainties given as determined by Raitt [1963].

The Deep Sea Drilling Project has produced numerous samples from cores penetrating hundreds of meters into the igneous crust. We now find ourselves in a position to correlate the recovered samples to the seismic refraction data and determine the mean density of the oceanic crust.

For the purposes of this study, the oceanic crust is defined as the igneous crust. The denisty of this crust is an important quantity in various geodynamic considerations, most notably the bouyancy and instability arguments concerning subduction. Within the literature, one can find the density of the oceanic crust cited from anywhere between 2.80 and 2.90 gm/cm^3 [Watts, 1978; Turcotte and Schubert, 1982]. In short, there has been neither standardization within the calculations and models, nor any rigorous study aimed at determining the density of this crust. It is the aim of this study to determine the mean density of the igneous oceanic crust in order to provide a standard value for all subsequent calculations and models.

THE OCEANIC CRUST AND OPHIOLITES: A COMPARISON

A large number of samples have been recovered directly from the ocean crust. The Deep Sea Drilling Project (DSDP) has provided numerous cores penetrating up to 589 meters into the igneous crust, principally at sites 332A (230m), 332B (589m), 333A (300m), 395A(580m), 396B (256m), 417A (209m), 417D (366m), and 418A (544m). The vast majority of these samples are basalts with the remainder being dolerites and gabbros. This high occurrence rate of basalt is explained in that although these cores represent significant penetrations into the upper igneous crust (layer 2), deep portions of layer 2 and layer 3 have not been sampled. However, this does indicate that the upper igneous crust is primarily basalt.

The widespread occurrence of basalts in the upper igneous crust conforms well with the ophiolite model. A generalized cross-section of an ophiolite would contain three major units [Figure 2]. The uppermost section is characterized by extrusives, namely pillow basalts and lava flows. The intermediate section is composed largely of basaltic vertical dikes which can become gabbroic or doleritic at depth. The dikes grade down into the lowest unit, which consists of cumulate and/or massive gabbros.

By examining the processes of oceanic crust formation, one can easily explain ophiolite formation. A generalized model of crust formation along a ridge axis involves simple accretion. New material from depth rises and the crust moves out from the ridge. The pillow basalts and lava flows found in the ophiolite sections

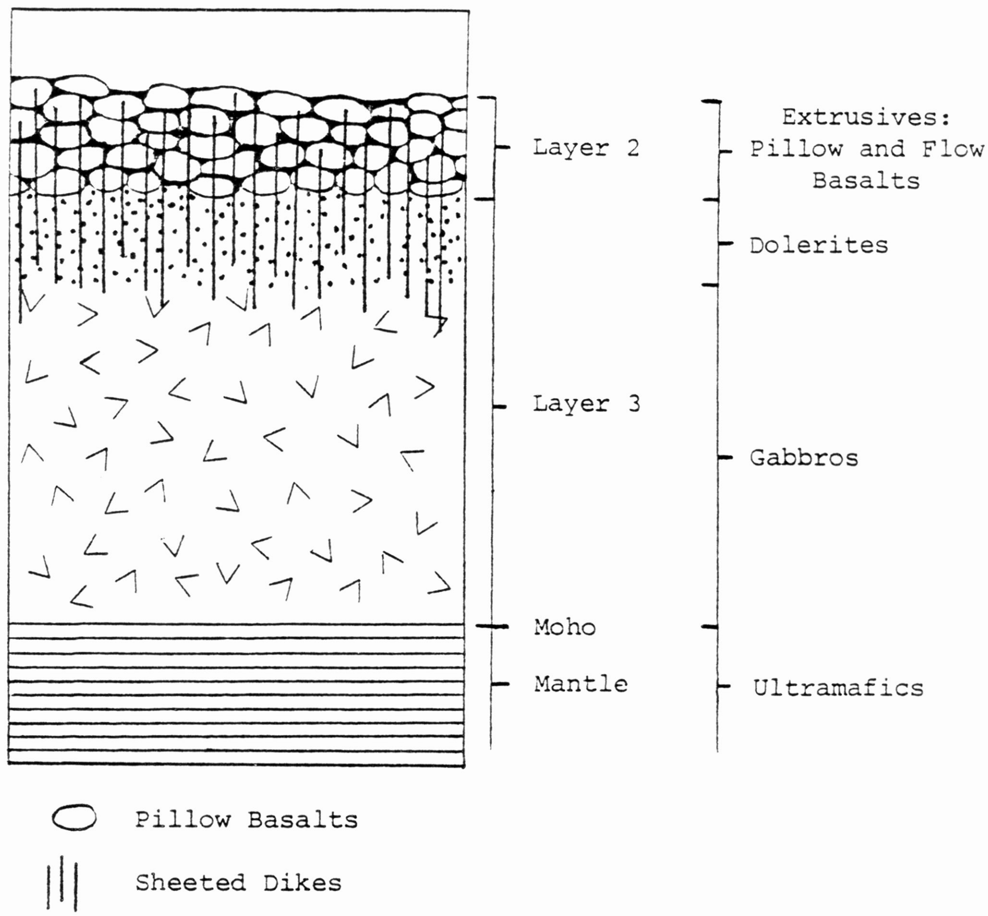

Fig. 2. Generalized cross-section of an ophiolite. Major lithologic units are presented, as is the seismic interpretation of the entire structure.

correlate quite well with dredge haul samples from the top of the crust [Christensen and Salisbury, 1975]. In addition the existence of pillow basalts dictates a sub-aqueous environment of formation. The vertical sheeted dikes appear to be feeder dikes to the surface. Their relatively thin one-sided structure and absence of country rock has been interpreted to have been formed in a tensional environment, as is thought to occur along the ridge axis [Gass-Masson Smith, 1963; Moores and Vine, 1971]. Christensen and Salisbury [1975] note that "the presence of sheeted dikes and cumulates implies continuous creation of void space at a site of tensional spreading; the filling of this space with tholeiitic magma suggests that this site was in the ocean basins". In addition, even such minor constituents as the plagiogranites found in ophiolites are often indistinguishable from their oceanic counterparts [Aldiss, 1981].

Another link between the ocean crust and ophiolites is found in their respective seismic velocity profiles. The classic seismic model of the oceanic crust involves a layer of sediments and two layers of igneous crust [Figure 1]. The seismic profiles of ophiolites appear quite similar to that of the oceanic crust [Christensen and Salisbury, 1975]. And if one considers that the ocean crust may be gradational instead of layered [Kennett and Orcutt, 1976], the ophiolites still prove to be a valid model [Spudich and Orcutt, 1980].

By studying the seismic character of the ophiolites in relation to their lithologies and structures, it is possible to hypothesize the gross structures and lithologies of the igneous ocean crust. This is accomplished by assigning rock types and/or structures to the various intervals of the seismic velocity profiles of ophiolites, and applying these to the profiles of the oceanic crust. A summary of these assignments to the classic two-layer models, as well as the subdivisions of layers two and three as suggested by Christensen and Salisbury [1975], Houtz and Ewing [1976], and Purdy [1983], is given in Table 1.

TABLE 1. Summary of Layered Models and Associated Lithologies

Reference	Layer	Thickness (km)	Vp (km/s)	Lithology
Raitt [1963]	2	1.71±0.75	5.07±0.63	Basalt
	3	4.86±1.42	6.69±0.26	Gabbro
Shor et.al. [1971]	2	1.49±0.98	5.19±0.64	Basalt
	3	4.62±1.30	6.81±0.16	Gabbro
Christensen and Salisbury [1975]	2	1.39±0.5	5.04±0.69	Basalt
	3	4.97±1.25	6.73±0.19	Gabbro
Christensen and Salisbury [1975] Sonobuoy Type 1	2	1.6	4.4	Basalt
	3a	1.2	6.4	Gabbro
	3b	4.8	2.1	Gabbro
Christensen and Salisbury [1975] Sonobuoy Type 2	2	1.6	4.4	Basalt
	3a	3.0	6.8	Gabbro
	3b	2.6	7.5	Gabbro
Houtz & Ewing [1976] Atlantic	2a	0.3±0.1	3.74±0.50	Basalt
	2b	1.0±0.1	5.13±0.38	Basalt
	2c	1.0±0.3	6.05±0.22	Dolerite
	3	4.97±1.25	6.83±0.21	Gabbro
Houtz & Ewing [1976] Pacific	2a	0.4±0.1	3.47±0.35	Basalt
	2b	0.8±0.1	5.28±0.39	Basalt
	2c	0.9±0.4	6.12±0.18	Dolerite
	3	4.97±1.25	6.90±0.17	Gabbro
Purdy [1983]	2a	0.38	5.30	Basalt
	2b	1.93	6.10	Dolerite
	3a	1.55	6.86	Gabbro
	3b	3.15	7.06	Gabbro

MEAN DENSITIES AND VELOCITIES OF OCEAN CRUSTAL ROCKS

In order to correlate the hand samples to the seismic data and thereby determine the mean density of the oceanic crust, it is necessary to determine the density and velocity properties of the major constituents. It is important to know how well we can define the density and velocity of a basalt, dolerite, or gabbro. For the purposes of this study, only relatively unaltered samples were considered. The data come from DSDP Sites 332A, 332B, 333A, 395A, 396B, 417A, 417D, and 418A as well as the Blow-Me-Down, North Oman, North Arm Mountain, and selected American Ophiolite suites [Hyndman, 1977; Aumento, et al., 1977; Melson et al., 1979: Dimitriev et al., 1979; Christensen, et al., 1979; Donnelly et al., 1980; Christensen et al., 1980; Salisbury and Christensen, 1978; Christensen and Smewing, 1981; Christensen and Salisbury, 1978; Christensen, 1978].

The distribution of wet-bulk densities of basalts, dolerites, and gabbros is shown in Figure 3. The left-skewed nature of these histograms is caused by a normal grain (matrix) density distribution with an independant porosity distribution overprint. Since the fluid (sea water) has a density far lower than that of the grains, the distribution can only skew to the left as increasing porosity will tend to lower the bulk density.

The distributions of compressional wave velocities at 1 kbar pressure are shown in Figure 4. Again there appears a small amount of left-skewness. A similar line of reasoning as given above accounts

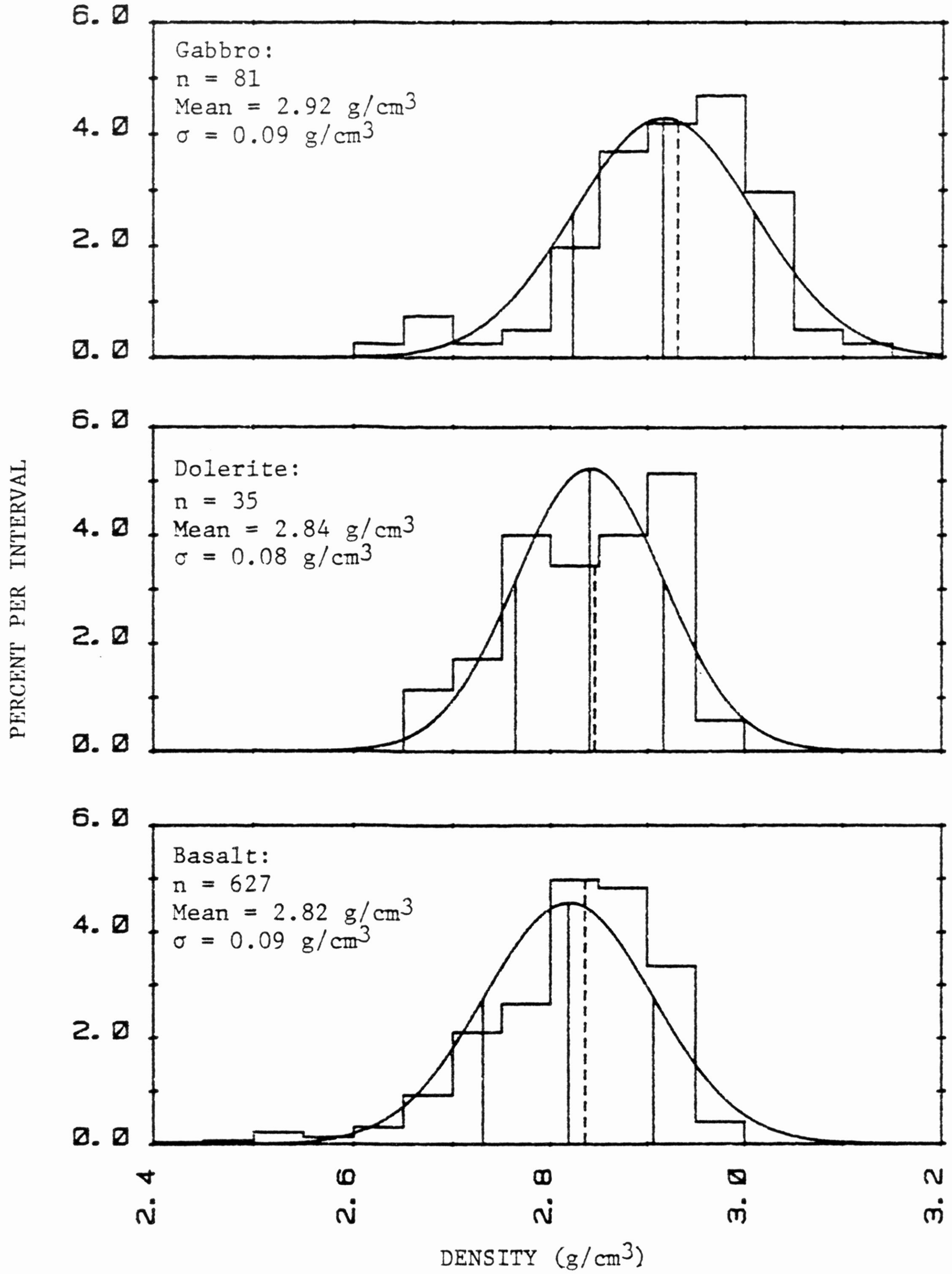

Fig. 3. Histogram of wet-bulk densities. Solid vertical lines represent mean and one standard deviation. Dashed vertical line represents the median.

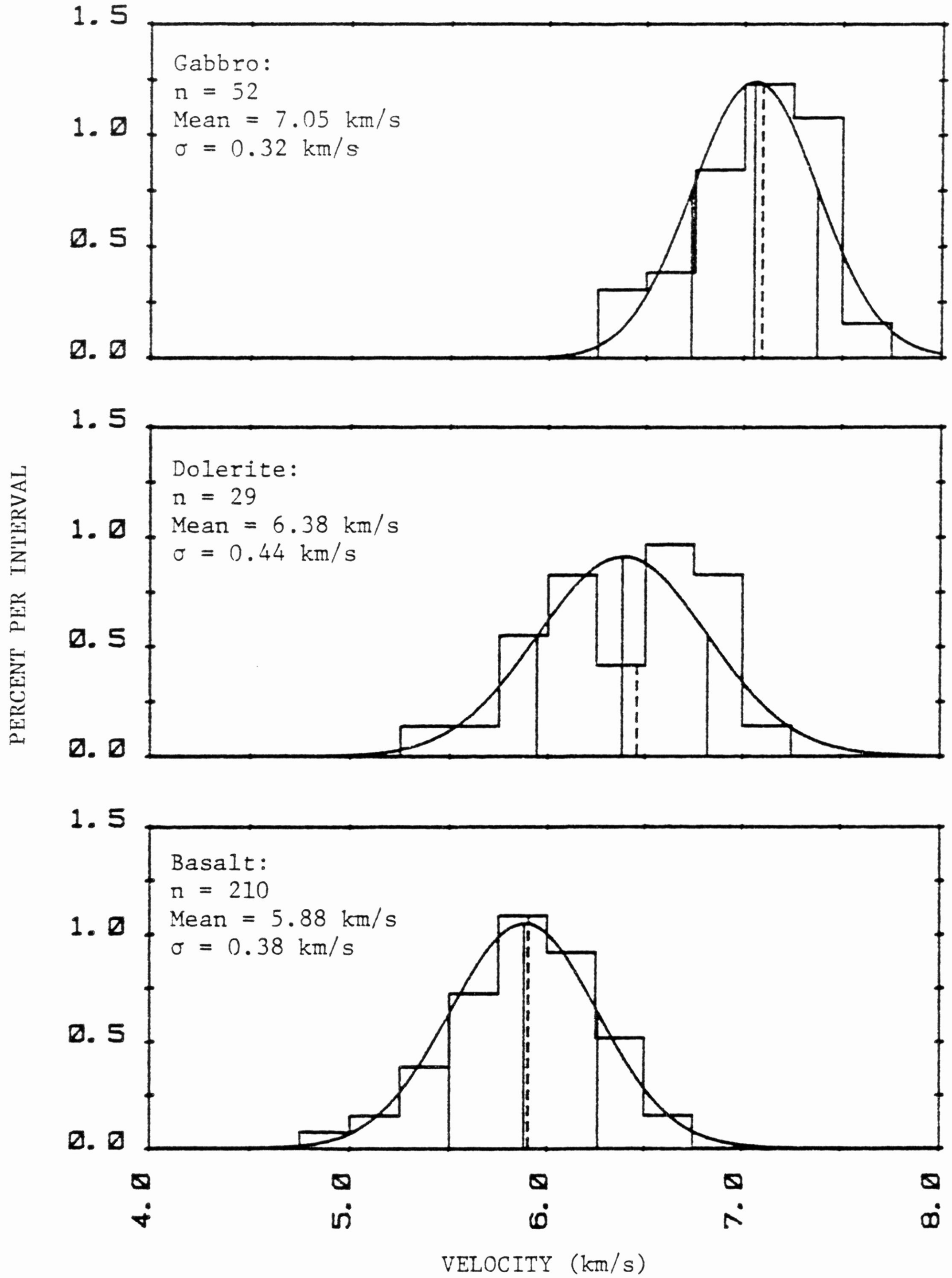

Fig. 4. Histogram of velocities at 1 kb pressure. Solid vertical lines represent mean and one standard deviation. Dashed vertical line represents the median.

for this skewness.

For the purposes of this study, the skewness is considered insignificant and the parent populations are approximated as normal distributions. In addition, the statistics of the normal distribution are better understood and developed than that of the quartile type distribution. Table 2 summarizes the mean densities and velocities of the basalts, dolerites, and gabbros as determined in this study.

TABLE 2a. Summary of Average Densities of Crustal Rocks

Rock Type	Number of Samples	Mean (g/cm^3)	Standard Deviation (g/cm^3)
Basalt	627	2.82	0.09
Dolerite	35	2.84	0.08
Gabbro	81	2.92	0.09

TABLE 2b. Summary of Average Velocities

Rock Type	Number of Samples	Mean (km/s)	Standard Deviation (km/s)
Basalt	210	5.88	0.38
Dolerite	29	6.38	0.44
Gabbro	52	7.05	0.32

VELOCITY-DENSITY RELATIONSHIPS FOR OCEAN CRUSTAL ROCKS

The strong correlation between compressional wave velocities and densities of sediments and rocks has been well established for nearly three decades [Nafe and Drake, 1957; Nafe and Drake, 1963; Ludwig, Nafe, and Drake, 1970; Hamilton, 1978; Christensen et al., 1980; Hamilton and Bachman, 1982]. Hamilton [1978] suggests that the application of velocity-density relationships to the seismic record should aid in the determination of layer densities within the crust. However, instead of using a visual best fit [Ludwig, Nafe, and Drake, 1970] to various sediments, sedimentary, metamorphic and igneous rocks, it is desirable to produce a statistical best fit to the rocks expected within the crust.

It is important to determine in what manner density should vary with velocity. In order to accomplish this, it is easiest to see how each varies with a common parameter, in this case porosity. Wyllie, Gregory, and Gardner [1956] showed that

$$\frac{1}{V_F} = \frac{\phi}{V_f} + \frac{(1 - \phi)}{V_g} \tag{1a}$$

where V_F is the velocity of the formation, V_f is the velocity of the fluid, V_g the grain or matrix velocity, and ϕ the fractional porosity. Equation (1a) can be rewritten parametrically as

$$\frac{1}{V_F} = \frac{1}{V_g} + \left(\frac{1}{V_f} - \frac{1}{V_g} \right) \cdot \phi \tag{1b}$$

or solving for ϕ

$$\phi = \left(\frac{1}{V_F} - \frac{1}{V_g} \right) \div \left(\frac{1}{V_f} - \frac{1}{V_g} \right) \tag{1c}$$

It is also known that density varies with porosity as

$$\rho_F = \phi \cdot \rho_f + (1 - \phi) \cdot \rho_g \tag{2a}$$

in which ρ_F, ρ_f, ρ_g, and ϕ are analogous to V_F, V_f, V_g, and ϕ from (1a). Similarly, equation (2a) can be rewritten parametrically as

$$\rho_F = \rho_g + (\rho_f - \rho_g) \cdot \phi \ . \tag{2b}$$

By substituting (1c) into (2b) for ϕ, the function for ρ_F becomes

$$\rho_F = \rho_g + (\rho_f - \rho_g) \cdot \left(\frac{1}{V_F} - \frac{1}{V_g} \right) \div \left(\frac{1}{V_f} - \frac{1}{V_g} \right) \tag{3a}$$

If one notes that ρ_g, V_g, ρ_f, and V_f are constants, then (3a) can be rewritten as

$$\rho_F = A + B/V_F \tag{3b}$$

in which

$$A = \rho_g - \frac{B}{V_g} \tag{4a}$$

and

$$B = \frac{(\rho_f - \rho_g)}{\left(\frac{1}{V_f} - \frac{1}{V_g} \right)} \tag{4b}$$

Therefore, it is expected that density will vary linearly with the reciprocal of velocity. Applying a linear regression of this type to all recovered samples of marine or ophiolitic origin [Figure 5], (3b) is found to be

$$\rho_F = (3.81 \pm 0.02) + (-5.99 \pm 0.11) \div V_F \quad g/cm^3 \tag{5}$$

for 483 samples, $r^2 = 0.86$, and rms error = 0.07 g/cm^3.

Since basalt constitutes nearly 80% of the parent population, it is important to test if the curve was biased in their favor. One

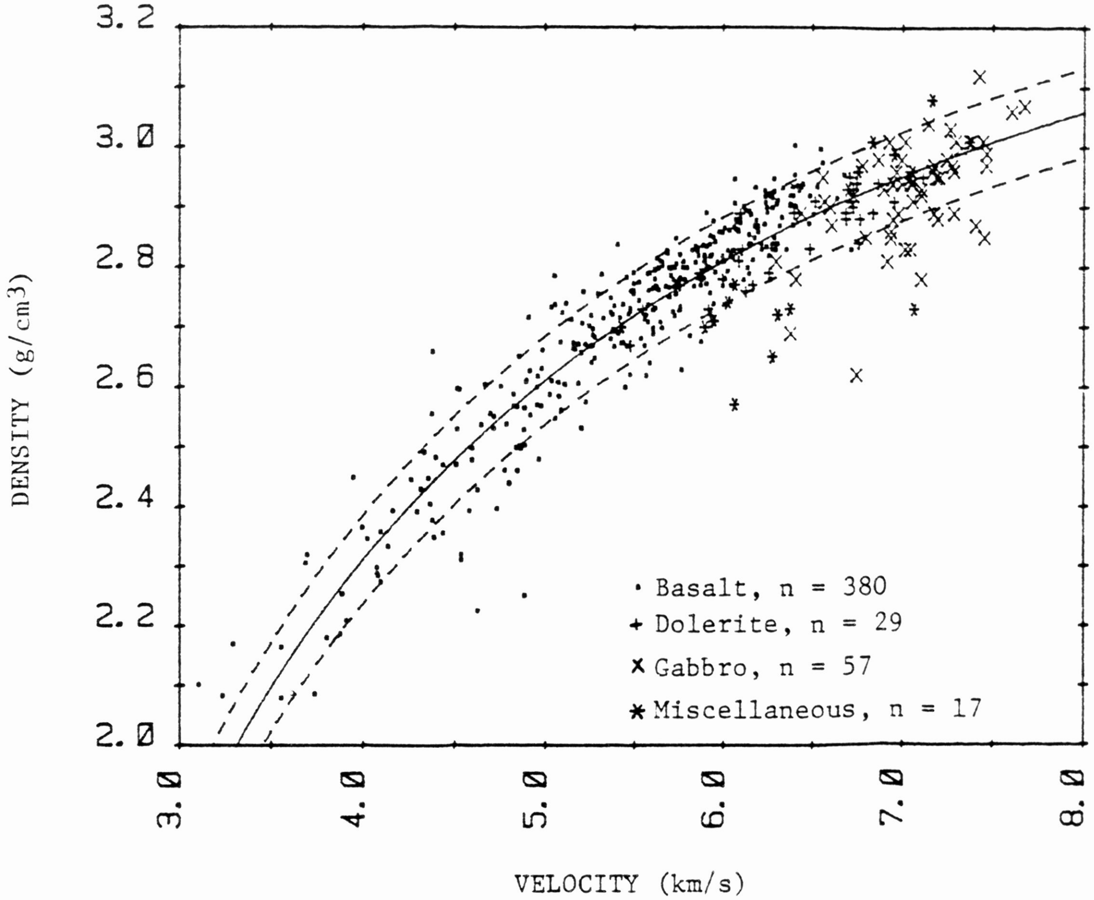

Fig. 5. Linear regression fit to all samples: $\rho = (A + B/V_p)$ g/cm^3. Dashed lines indicate rms error envelope. 483 samples total. A = 3.81 ± 0.02 and B = -5.99 ± 0.11 km/s. Rms error = 0.07 g/cm^3 for all samples. Coefficient of determination, r^2, equals 0.862.

measure of this is to calculate the rms error for each of the different rock types separately. Table 3 demonstrates that there is not any bias in favor of basalts at the expense of any other rock type. In fact, the rms errors are consistently less than the standard deviations from the mean for each individual rock type.

As previously stated, the goal is to determine a statistical best fit to the rocks expected within the crust. Many of the basalt samples were recovered at or near the surface of the crust from which they were exposed to sea water for several millions of years. These highly altered samples have low velocities and densities. Drilling results indicate that extensive alteration is confined to the top 50-70 meters and along narrow highly localized fractures [Christensen et al., 1978]. Christensen et al. [1979] and Fountain [1980] showed that within drilled basalts, the variation of velocity and density was controlled not by alteration, but by porosity. It can then be assumed that a low crustal velocity is due to large formation porosity, rather than alteration. Returning to the parametric velocity and density equations of (1b) and (2b), a theoretical relationship between velocity and density can be calculated.

By noting that the y-intercepts in both (1b) and (2b) have physical meaning, it is quite easy to calculate the grain velocity and density of basalt. Applying a linear fit of the type $1/V_p = A + B \cdot \phi$ [Figure 6] to samples recovered on Leg 37 of the DSDP [Hyndman, 1977; Aumento et al., 1977] produces the coefficients A = 0.150±0.002 s/km and B = 0.002 s/km%. Knowing

TABLE 3. Rms Errors Computed from Equation (5) for Crustal Rock Groupings

Rock Grouping	Number of Samples	RMS Error (g/cm^3)	Standard Deviation (g/cm^3)
Basalts	380	0.07	0.20
Unaltered Basalts	210	0.05	0.09
Dolerites	29	0.04	0.08
Gabbros	57	0.08	0.09
Miscellaneous	17	0.12	0.15
All Samples	483	0.07	----
Unaltered Samples	308	0.06	----

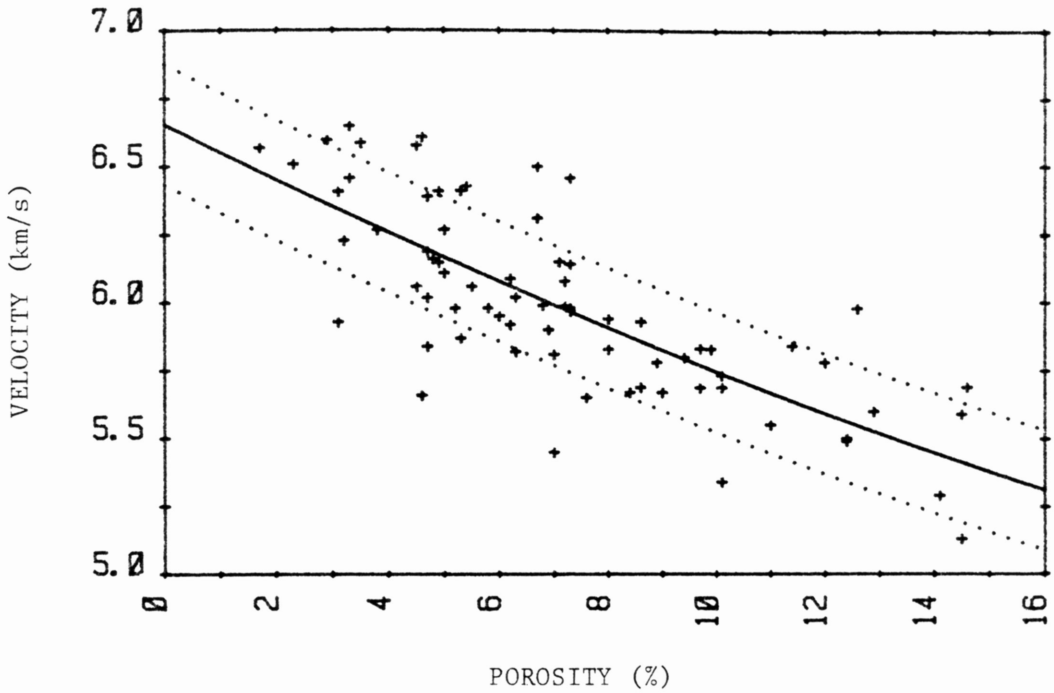

Fig. 6. Linear regression fit to Leg 37 samples: $1/V_p = A + B\cdot\phi$. Dotted lines indicate rms error envelope. 73 samples total. $A = 0.1503 \pm 0.0018$ and $B = 0.0024 \pm 0.0002$. Rms error = 0.22 km/s and $r^2 = 0.595$. All data from Aumento et. al. [1977] and Hyndman [1977].

$A = 1/V_g$ and $\sigma A = 0.002$ s/km, we find σV_g via the equation

$$(\sigma_f)^2 = \sum_{i=1}^{n} \left(\sigma_{xi} \frac{\partial f}{\partial xi} \right)^2 \qquad (6)$$

where $f = f(x_1, x_2, x_3, \ldots x_n)$ [Bevington, 1969]. Hence,

$$(\sigma_A)^2 = (\sigma V_g)^2 \cdot \left(\frac{-1}{V_g{}^2} \right)^2$$

$$\sigma V_g = \sigma A \cdot V_g{}^2.$$

Therefore the grain velocity of basalt is 6.65±0.08 km/s. Applying a linear regression fit of the form $\rho = A + B\phi$ [Figure 7] to the same data set produces the coefficients A = 2.93±0.01 g/cm³ and B = -0.016±0.002 g/cm³. Thus, the grain density of basalt is 2.93±0.01 g/cm³.

It appears that from equations (1b) and (2b) that the fluid velocity and density could be calculated from the regression coefficient B. This is not the case. There seems to be a systematic change in the grain properties of these samples with increasing porosity. Therefore, any fluid velocity or density calculation would represent some sort of average property of fractions of sea water and alteration by-products such as smektite. However, since this change appears to be systematic (linear), it does not affect the calculations of grain velocity and density.

Returning to the suggestion that alteration is primarily confined to the uppermost 50-70 meters of the crust, it is assumed that the fluid material at depth within the crust is sea water. Given the velocity and density of seawater (1.53 km/s and 1.025 g/cm² respectively [Weast, 1974]) and the grain velocity and density of basalt, the substitution of these values into equations (4a) and (4b)

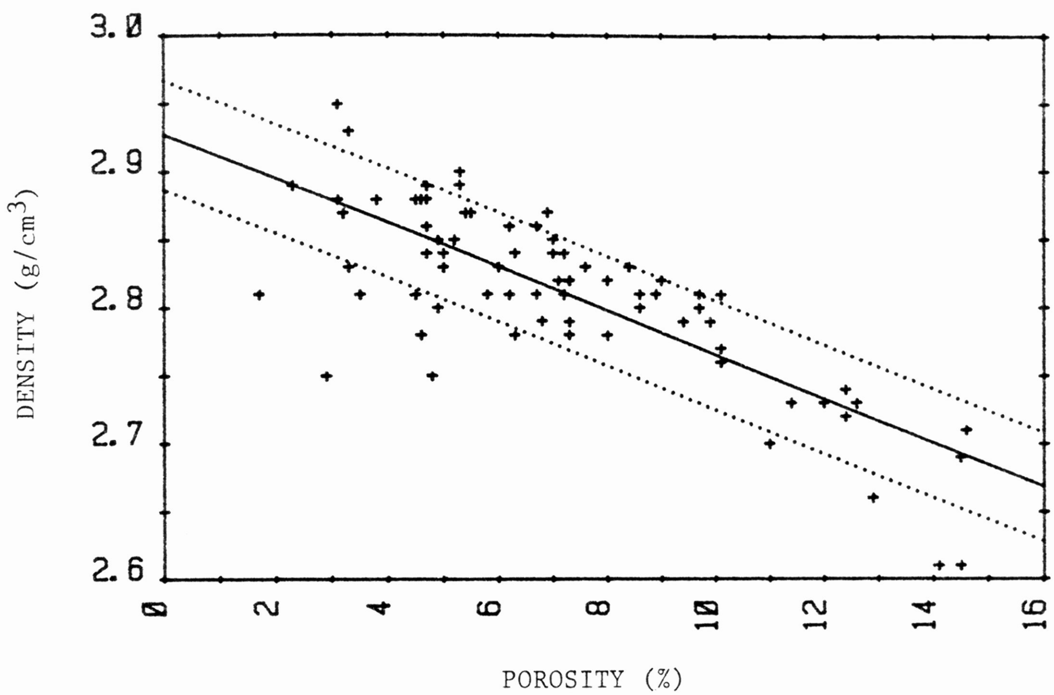

Fig. 7. Linear regression fit to Leg 37 samples: $\rho = A + B \cdot \phi$. Dotted lines indicate rms error envelope. 73 samples total. $A = 2.93 \pm 0.01$ and $B = -0.016 \pm 0.002$. Rms error = 0.04 g/cm^3 and $r^2 = 0.618$. All data from Aumento et. al. [1977] and Hyndman [1977].

yields

$$\rho_F = (3.50\pm0.01) + (-3.79\pm0.03) \div V_F \quad g/cm^3 \qquad (7)$$

where $V_F \leq 6.65$ km/s [Figure 8].

Equation (7) represents the variations of velocity and density of basalt as functions of a changing porosity. This porosity is the sum of both grain boundary porosity, that which can be measured in a hand sample, and open cracks filled with sea water or formation porosity. Since hand samples essentially contain only a grain boundary porosity, none of the previous data suggest the relationship in Figure 8. However, downhole logging results for velocities and densities from sites 396B, 417D, and 504B [Christensen et al., 1979; Salisbury et al., 1980; Anderson et al., 1982; Becker et al., 1982] do correlate and thereby confirm the proposed velocity-density relationship [Figure 9].

There now exist two equations to represent the variation of density with velocity in ocean crustal rocks: for velocities greater than 6.65 km/s

$$\rho_F = (3.81\pm0.02) + (-5.99\pm0.11) \div V_F \quad g/cm^3 \qquad (5)$$

and for velocities less than 6.65 km/s

$$\rho_F = (3.50\pm0.01) + (-3.79\pm0.03) \div V_F \quad g/cm^3. \qquad (7)$$

To calculate the uncertainty in either evaluation of ρ_F, it is necessary to recall

$$(\sigma_f)^2 = \sum_{i=1}^{n} (\sigma_{xi} \frac{\partial f}{\partial xi})^2 \qquad (6)$$

Applying (6) to the general form $\rho_F = A + B/V_F$ of both (5) and

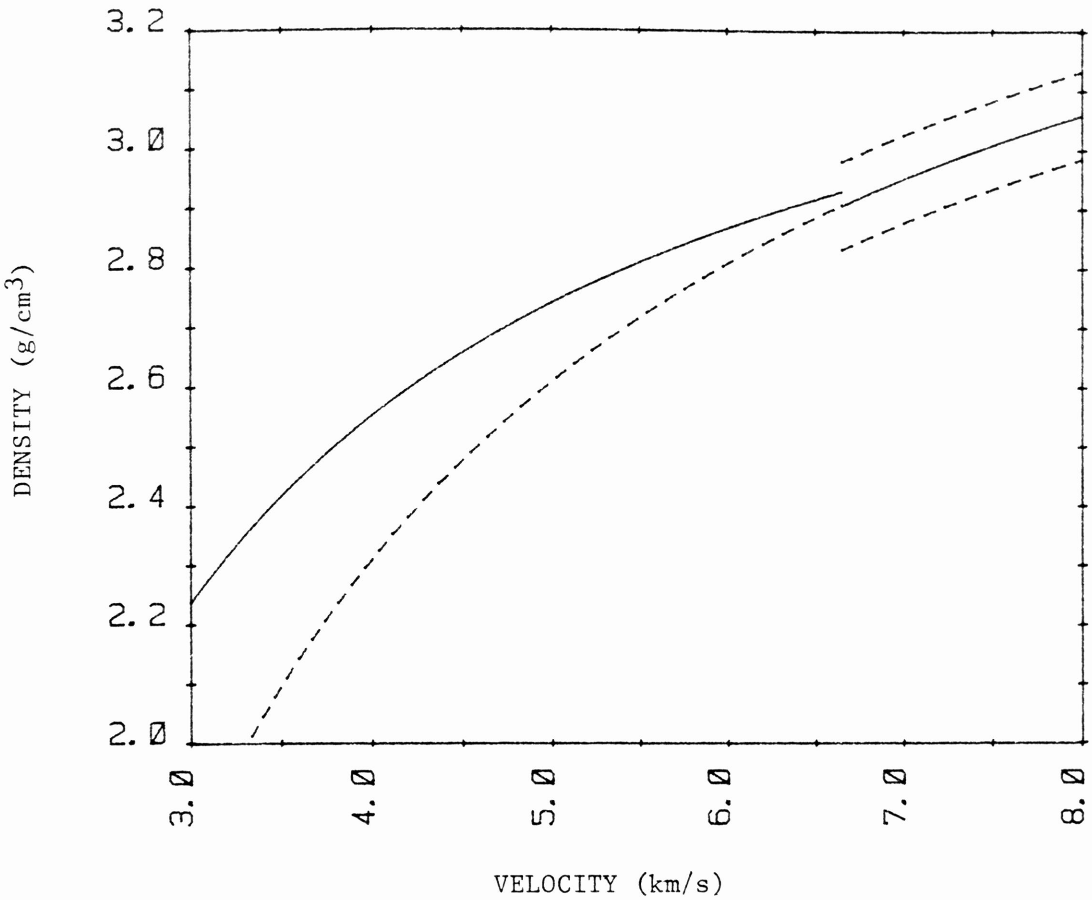

Fig. 8. Parametric velocity-density relation with regression fit to all data for comparison. Parametric function is of the form $\rho = A + B/V_p$. $A = 3.50 \pm 0.01$ g/cm^3 and $B = -3.79 \pm 0.03$ g/cm^3·km/s.

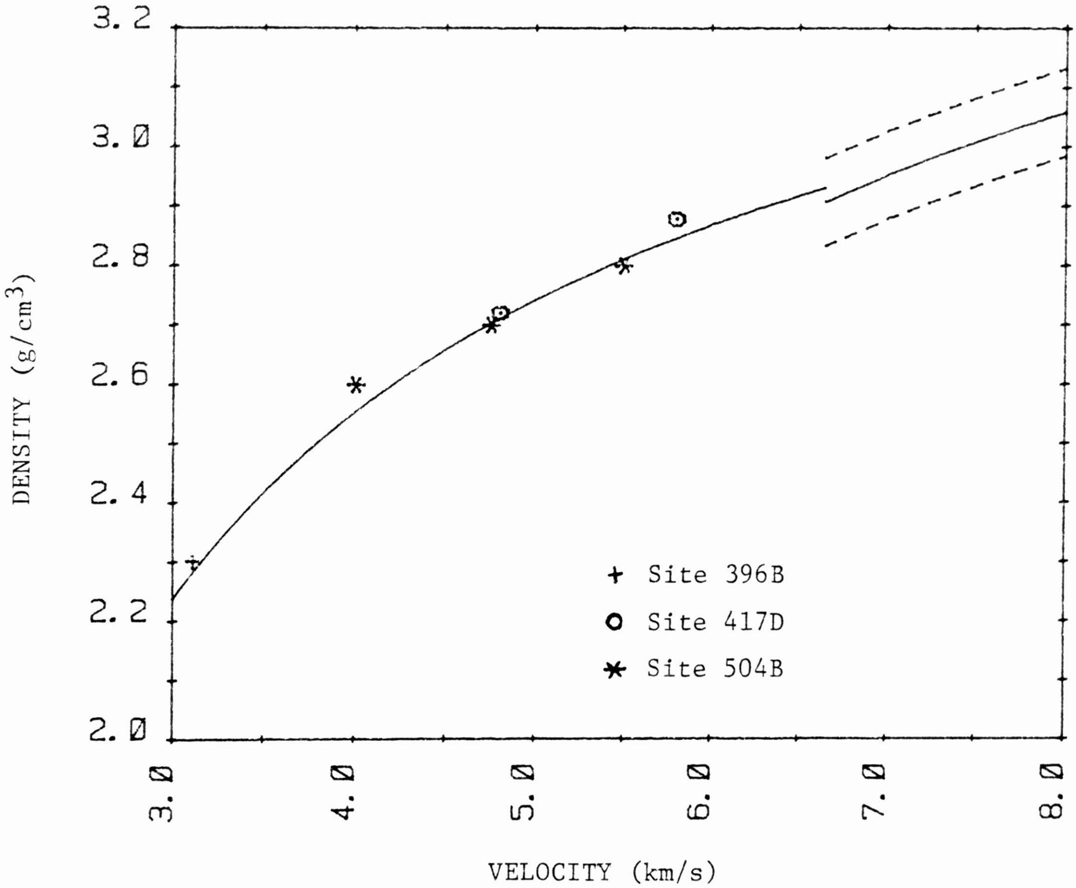

Fig. 9. Downhole logging results with velocity-density relations for comparison. Data from Kirkpatrick[1979], Salisbury et. al.[1980], and Anderson et. al.[1982].

(7) yields

$$\sigma\rho_F = \left[\sigma_A{}^2 + (\sigma_B/V_F)^2 + (B \cdot \sigma V_F/V_F{}^2)^2\right]^{-\frac{1}{2}} \tag{8}$$

Sample calculations for V_F = 5.0±0.2 km/s and 7.0±0.2 km/s yield ρ_F = 2.74±0.03 g/cm^3 and 2.95±0.04 g/cm^3 respectively.

Thus, it is assumed that a velocity less than 6.65 km/s is due to a porosity effect. Deep drilling results indicate that alteration is mainly confined to the surface. In addition, downhole logging data correlate well with equation (7) and thereby validate its use. For velocities higher than 6.65 km/s, equation (5) is used to describe essentially zero porosity mafic rocks of varying grain sizes and compositional forms.

DETERMINATION OF OCEAN MODEL DENSITIES

Two methods have been proposed through which the mean density of the ocean crust can be determined, lithology assignments and velocity-density relationships. The mean densities of the major constituents of the crust have been determined [Table 2]. In addition, a set of velocity-density relationships have been derived both theoretically and empirically. Hence, there now exist both the method and means to determine layer densities within the crust. To average the individual layer densities into a single crustal density, a weighted average method is employed [Arkin and Colter, 1970] in which

$$D = \sum_{i=1}^{n} \frac{\rho_i t_i}{\Sigma t_J} \tag{9}$$

for n layers, ρ_i and t_i the density and thickness of the i^{th} layer, and D the average density of the ocean crust. To calculate the error in an evalution of (9), again equation (6) is recalled.

$$(\sigma_f)^2 = \sum_{i=1}^{n} (\sigma_{x_i} \frac{\partial f}{\partial xi})^2 \tag{6}$$

Expanding (9) for clarity

$$D = (\rho_1 t_1 + \rho_2 t_2 + \rho_3 t_3 + \ldots + \rho_n t_n) \div (t_1 + t_2 + t_3 + \ldots + t_n).$$

Calculating the partial derivatives yields

$$\frac{\partial D}{\partial \rho_i} = \frac{ti}{T} \tag{10a}$$

and

$$\frac{\partial D}{\partial t_i} = \frac{(\rho_i T - M)}{T^2} \tag{10b}$$

where $T = \Sigma t_i$ and $M = \Sigma \rho_i t_i$.

Therefore, the error in (9) is found to be

$$(\sigma_D)^2 = \sum_{i=1}^{n} \left[(\sigma\rho_i \frac{ti}{T})^2 + (\sigma_{t_i} \cdot \frac{\rho_i T - M}{T^2})^2 \right] \tag{11}$$

By applying the layer thicknesses, velocities, and lithologies given in Table 1 to equations (9) and (11), the mean densities of the given models can be calculated [Table 4]. The lithology assignment method indicates a ocean crustal density of 2.90±0.06 g/cm^3. However, this method does not account for porosity effects which can seriously affect both the velocity [Fountain, 1980] and density of layer 2. The velocity-density relationships do account for porosity effects and therefore indicate a slightly lower crustal density of 2.89±0.03 g/cm^3. However, the discrepancy is statistically insignificant in light of errors involved in the seismic refraction method as well as that of the velocity-density relationships.

TABLE 4. Mean Densities of Ocean Models

Reference	Thickness (km)	Density from Lithology (g/cm^3)	Density from Velocity (g/cm^3)
Raitt [1963	6.57±1.61	2.89±0.07	2.87±0.04
Shor et.al. [1971]	6.11±1.63	2.90±0.07	2.89±0.04
Christensen and Salisbury [1975]	6.36±1.35	2.90±0.07	2.88±0.04
Sonobuoy Type 1	7.60	2.90±0.06	2.89±0.02
Sonobuoy Type 2	7.20	2.89±0.05	2.89±0.01
Houtz and Ewing: Atlantic [1976]	7.27±1.29	2.89±0.06	2.88±0.03
Houtz and Ewing: Pacific [1976]	7.07±1.32	2.89±0.07	2.89±0.03
Purdy [1983]	7.01	2.89±0.05	2.92±0.01

APPLICATIONS TO FURTHER GEOPHYSICAL PROBLEMS

The results of this study indicate a mean density of the oceanic crust at 2.89 g/cm^3. This value is significant in that now there can be standardization within models and calculations concerning the crust. However, perhaps the most important consequence of this study lies in the application of derived techniques to additional problems and areas of study.

The vertical velocity gradient model of the ocean crust has attracted considerable interest in recent years [Kennett and Orcutt, 1976; Spudich and Orcutt, 1980]. When one considers that pressure increases with depth and that seismic velocities in rocks increase with pressure [Birch, 1960; Christensen, 1965], it becomes apparent that even a homogeneous layer would exhibit a vertical velocity gradient. The gradient model is therefore considered more realistic than a simple layered model. The velocity-density relationships of equations (5) and (7) can be applied to gradient models by either integrating the gradient function if it is known, or by simply approximating the gradient as several small linear gradients. Applying the latter technique to gradient model CH-10 A from Kennett and Orcutt [1976] yields a density of 2.95±0.05 g/cm^3 and can be seen graphically in Figure 10. However, there do exist several problems with the analysis of gradient models. Aside form uncertainties in depths and thicknesses of layers [Christensen and Salisbury, 1975], the largest problem lies in that the MOHO seems to

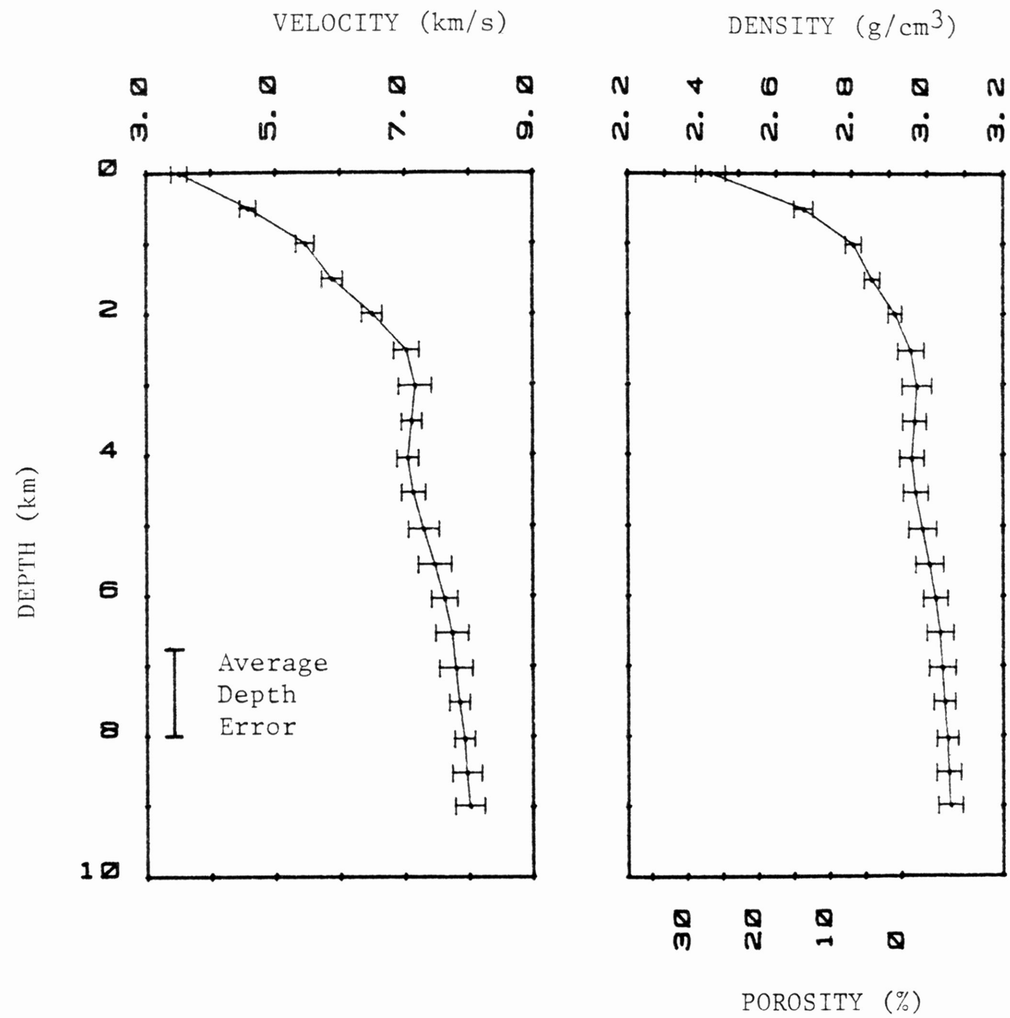

Fig. 10. Conversion from a velocity gradient to a density gradient. Porosity is that of basalt assuming normal sea water as pore fluid.

disappear. This makes it extremely difficult to determine the depth at which the crust gives way to mantle. Suffice it to say that the analysis of gradient models in terms of crustal structure and density is beyond the scope of this study. However, the method and the means are now available.

The remote detection of formation porosity is a problem of both academic and commercial importance. Since formation porosity is essentially open cracks with standing water, a high formation porosity indicates a high permeability or potential fluid flow. The fluid could be water conducting heat and chemically altering the material, or it could be petroleum or gas. Knowing where the most porous zones are located should aid in determining well location and increasing output. Although the parametric curve relating velocity and density to porosity, equation (7), correlates well with observed formation densities and velocities [Figure 9], the porosities do not. This is due to the approximation that the interstitial fluid is sea water. It is known that there are fractions of alteration by-products as well as dissolved and suspended solids. However, with increased downhole logging data, the porosity scale can be empirically calibrated.

Another problem is that of the state of the crust as a function of age. Hydrothermal circulation is generally considered a primary agent in the alteration process of the upper crust [Turcotte and Schubert, 1982]. As the alteration continues, the fractures begin to fill with alteration by-products and the permeability drops.

This phenomenon is measured by heat flow as a function of age [Sclater, Jaupart, and Galson, 1980]. To try to characterize the state of alteration of Layer 2 as a function of age by examining samples would be difficult in that alteration is highly site specific. However, the seismic record tends to average an entire layer and thereby removes this site specific problem. Applying the velocity-density relationship of equation (7) to layer 2 velocities [Houtz and Ewing, 1976; Purdy, 1983] and plotting the resultant densities as a function of age demonstrates a systematic increase in density [Figure 11]. Fitting the data to a linear relationship of the type $\rho = A + B \text{ Log (age)}$ yields

$$\rho = (2.65 \pm 0.02) + (0.09 \pm 0.01) \text{ Log (age)} \quad g/cm^3 \tag{12}$$

where the age is in Myr. It must be noted that all points, except Purdy [1983], represent averages for a specific age over an entire ocean. For this reason, there was no error analysis performed. The suggestion is that now there exists a method to determine the density of the crust site by site. This could be utilized in two manners. First, a more in-depth study could be undertaken to produce a better relationship between crustal density and age. And second, the density of the ocean crust can now be mapped in an areal manner. Both of these suggested investigations would prove invaluable to the studies of both geodynamics and global gravity.

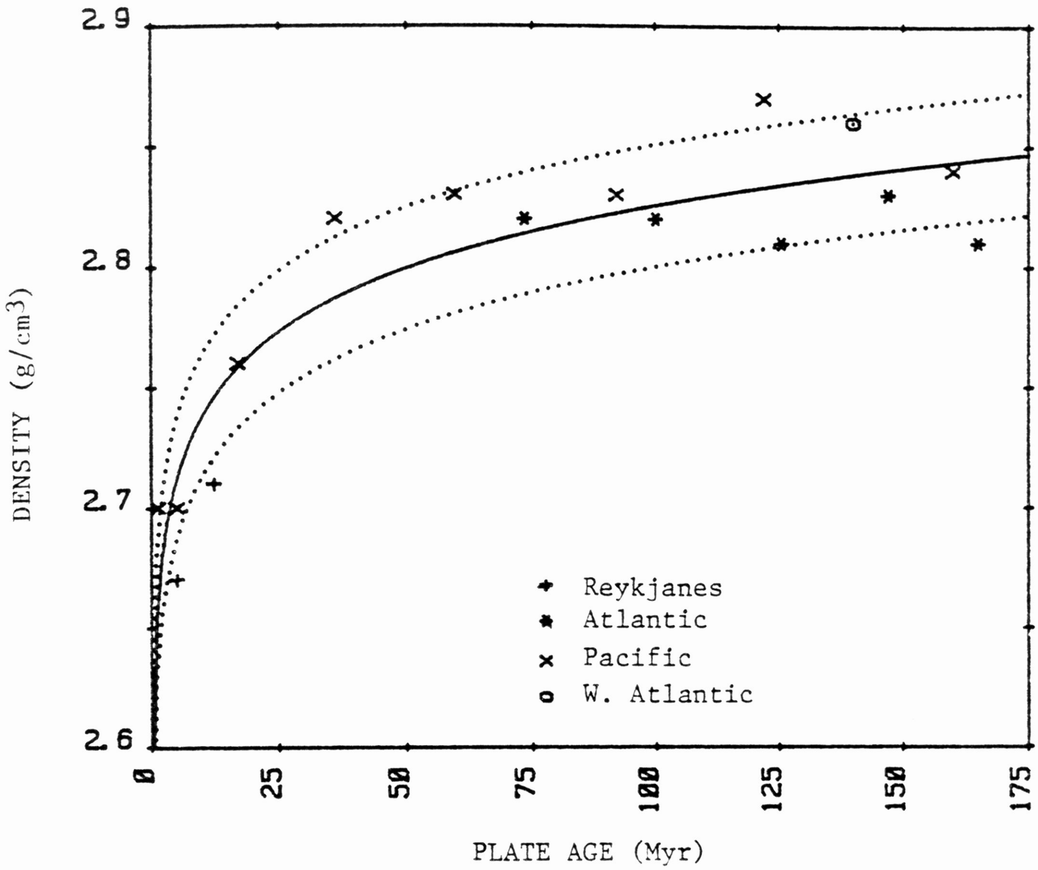

Fig. 11. Density of layer 2 as a function of plate age. All data from Houtz and Ewing [1976], except W. Atlantic from Purdy [1983].

SUMMARY AND CONCLUSIONS

Two techniques have been utilized to determine the mean density of the oceanic igneous crust. The first method entails assigning an average lithology to a layer as determined by studies of ophiolites. The mean densities and velocities of these lithologies have been determined by studying both marine and ophiolitic samples. Applying this method to eight different layered models indicates a mean density of 2.90 g/cm^2. The second method involves the use of velocity-density relationships. Two functions were derived for different ranges. The first describes the variation of velocity and density of basalt as functions of porosity. The second describes density as a function of velocity only for all rock types and indicates changes in grain size and composition of mafic rocks. This second method indicates a crustal density of 2.89 g/cm^3.

The velocity-density technique is considered the more accurate since it can account for large formation porosities. The applications of this technique are widespread and of importance. The relationships can be used to analyze vertical velocity gradient models. Of commercial as well as academic importance, porosities of layers can be calculated, from which permeability can be estimated. In addition, the time dependent nature of the crust can now be examined. Layer 2 has been shown to increase in density with increasing age. The velocity-density relationships can be applied to individual seismic surveys and determine the density of the crust on site. The collection of such data can be used to map the density of the entire worlds ocean crust laterally. Such a map would prove invaluable to geodynamicists.

REFERENCES

Anderson, R.N., et al., DSDP Hole 504B, the first reference section over 1 km through layer 2 of the oceanic crust, Nature, 300, 589-594, 1982.

Aldiss, D.T., Plageogranites from the ocean crust and ophiolites, Nature, 289, 577-578, 1981.

Arkin, M., R.R. Colton, in Statistical Methods, pp. 11-20, Barnes and Noble, New York, fifth edition, 1970.

Aumento, F., et al., Site reports: sites 332, 333, 334, and 335, in Initial Rep. Deep Sea Drill. Proj., 37, 15-326, 1977.

Becker, K., et al., In situ electrical resistivity and bulk porosity of the oceanic crust, Costa Rica Rift, Nature, 300, 594-598, 1982.

Bevington, P.R., in Data Reduction and Error Analysis for the Physical Sciences, pp. 56-64, McGraw-Hill, New York, 1969.

Birch, F., The velocity of compressional waves in rocks to 10 kbar, 1, J. Geophys. Res., 65, 1083-1102, 1960.

Christensen, N.I., Compressional wave velocity in metamorphic rocks at pressures to 10 kbar, J. Geophys. Res., 70, 6147-6164, 1965.

Christensen, N.I., Ophiolites, seismic velocities and oceanic crustal structure, Tectonophysics, 47, 131-157, 1978

Christensen, N.I., Et al., Velocities of compressional and shear waves in DSDP basalts, in Initial Rep. Deep Sea Drill. Proj., 27, 445-449, 1978.

Christensen, N.I., et al., Seismic velocities, electrical resistivities, densities, and porosities of basalts from DSDP Leg 46, in Initial Rep. Deep Sea Drill. Proj., 46, 383-388, 1979.

Christensen, N.I., et al., Compressional wave velocities, densities, and porosities of basalts from Holes 417A, 417D, and 418A, Deep Sea Drilling Project legs 51 through 53, in Initial Rep. Deep Sea Drill. Proj., 51-53(2), 1467-1477, 1980.

Christensen, N.I., M. Salisbury, Structure and constitution of the lower oceanic crust, Rev. of Geop. and Space Phys., 13(1), 57-86, 1975.

Christensen, N.I., M. Salisbury, Lateral heterogeneity in the seismic structure of the oceanic crust inferred from velocity studies in the Bay of Islands ophiolite, Newfoundland, Geophys. J. Roy. Astr. Soc., 68, 675-688, 1982.

Christensen, N.I., J.D. Smewing, Geology and seismic structure of the northern section of the Oman ophiolite, J. Geophys. Res., 86(B4), 2545-2555, 1981.

Dimitriev, L., et al., Site reports: Holes 396A and 396B, in Initial Rep. Deep Sea Drill. Proj., 46, 15-49, 1979.

Donnelly, T.W., et al., Site reports: Sites 417 and 418, in Initial Rep. Deep Sea Drill. Proj., 51-3(1), 23-626, 1980.

Fountain, D.M., Influence of porosity and water saturation on the compressional-wave velocities of basalts from the North Philippine Sea,in Initial Rep. Deep Sea Drill. Proj., 58, 935-940, 1980.

Gass, I.G., D. Masson-Smith, The geology and gravity anomalies of the Troodas Massif, Cyprus, Phil. Trans. Roy. Soc. London Ser. A, 255, 417-467, 1963.

Hamilton, E.L., Sound velocity-density relations in sea-floor sediments and rocks, J. Acoust. Soc. Am., 63(2), 366-377, 1978.

Hamilton, E.L., R.T. Bachman, Sound velocity and related properties of marine sediments, J. Acoust. Soc. Am., 72(6), 1891-1904, 1982.

Houtz, R., J. Ewing, Upper crustal structure as a function of plate age, J. Geophys. Res., 81(14), 2490-2498, 1976.

Hyndman, R.D., Seismic velocity measurements of basement rocks from DSDP Leg 37, in Initial Rep. Deep Sea Drill. Proj., 37, 373-387, 1977.

Hyndman, R.D., M.J. Drury, Physical properties of basalts, gabbros, and ultramafics from DSDP Leg 37, in Initial Rep. Deep Sea Drill. Proj., 37, 395-401, 1977.

Kennett, B.L.N., J.A. Orcutt, A comparison of travel time inversions of marine refraction profiles, J. Geophys. Res., 81(23), 4061-4070, 1976.

Kirkpatrick, R.J., The physical state of the oceanic crust: results of downhole geophysical logging in the mid-Atlantic ridge at 23°N, J. Geophys. Res., 84(B1), 178-188, 1979.

Ludwig, W.J., J.E. Wafe, C.L. Drake, Seismic refraction, in The Sea, 4(1), edited by A.E. Maxwell, pp. 53-84, Wiley, New York, 1970.

Melson, W.G., et al., Site 395: 23°N mid-Atlantic ridge, in Initial Rep. Deep Sea Drill. Proj., 45, 131-168, 1979.

Moores, E.M., F.J. Vine, The Troodos Massif, cyprus, and other ophiolites as oceanic crust: evalutaion and implications, Phil. Trans. Roy. Soc. London Ser. A, 268, 443-466, 1971.

Nafe, J.E., C.L. Drake, Variation with depth in shallow and deep water marine sediments of porosity, density, and the velocities of compressional and shear waves, Geophysics, 22, 523-552, 1957.

Nafe, J.E., C.L. Drake, Physical properties of marine sediments, in The Sea, 3, edited by M.N. Hill, pp. 794-815, Wiley, New York, 1963.

Purdy, G.M., The seismic structure fo 140 Myr old crust in the western central Atlantic Ocean, Geophys. J. Roy. Astr. Soc., 72, 115-137, 1983.

Raitt, R.W., The crustal rocks, in The Sea, 3, edited by M.N. Hill, pp. 85-102, Wiley, New York, 1963.

Salisbury, M.H., et al., 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 Initial Rep. Deep Sea Drill. Proj., 51-53(2), 1579-1597, 1980.

Salisbury, M.H., N.I. Christensen, The seismic velocity structure of a traverse through the Bay of Islands ophiolite complex, Newfoundland, an exposure of oceanic crust and upper mantle, J. Geophys. Res., 83(B2), 805-817, 1978.

Sclater, J.G., C. Jaupart, D. Galson, The heat flow through oceanic and continental crust and the heat loss of the earth, Rev. of Geop. and Space Phys., 18(1), 269-311, 1980.

Shor, G.G., et al., Structure of the Pacific basin, in The Sea, 4, edited by A.E. Maxwell, pp. 3-27, Wiley, New York, 1971.

Spudich, P., J.A. Orcutt, A new look at the seismic velocity structure of the oceanic crust, Rev. of Geop. and Space Phys., 18(3), 627-645, 1980.

Turcotte, D.L., G. Schubert, in Geodynamics, pp. 7-11, p. 76, Wiley, New York, 1982.

Watts, A.B., An analysis of isostacy in the worlds oceans 1. Hawaiin-Emperor seamount chain, J. Geophys. Res., 83(B12), 5989-6004, 1978.

Weast, R.C., Velocity of sound, in Handbook of Chemistry and Physics, p. E-47, edited by R.C. Weast, CRC Press, Cleveland, 55th edition, 1974.

Wyllie, M.R.J., A.R. Gregory, L.W. Gardner, Elastic wave velocities in heterogeneous and porous media, Geophysics, 21, 41-70, 1956.