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DEPARTMENT OF THE NAVY
NAVAL UNDERSEA RESEARCH AND DEVELOPMENT CENTER

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NAVAL UNDERSEA RESEARCH AND DEVELOPMENT CENTER

An Activity of the Naval Material Command

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THE PROBLEM

Determine and study the acoustic and related properties of the sea floor – specifically, sound velocity, density, porosity, grain size, and other properties. Determine differences between sediment types within various physiographic provinces and associated sedimentary environments.

RESULTS

1. Marine sediments studied were from three great physiographic provinces in the North Pacific: (1) continental terrace (shelf and slope), (2) deep-water, abyssal plain (turbidites), and (3) abyssal hill (pelagic). Samples were taken by coring and snapper-type samplers from surface ships, and *in situ* by divers and from deep-diving submersibles. Measurements of velocity were made *in situ* and in the laboratory.

2. The measurements and computations of mass physical properties are listed in tables according to sediment types within each environment; relationships between properties are shown in diagrams. The following properties are tabulated: grain size (mean and median diameter of grains, percents of sand, silt, and clay), saturated bulk density, density of mineral grains, porosity, sound velocity, velocity ratio (velocity in sediment/velocity in seawater), impedance, density $\times$ (velocity)^2, and Rayleigh reflection coefficients and bottom losses at normal incidence.

3. Significant differences in density and porosity are caused by mineralogy, size and shape of grains, rates of deposition, and sediment structure. Clay mineralogy is particularly important.

4. Earlier studies of sound velocity-porosity relationships produced generalized equations and curves over a full range of velocities and porosities. Data now available from deep-water areas indicate important environmental differences due to sediment structural rigidity. General equations and diagrams relating velocity and porosity should be abandoned in favor of entry into diagrams or equations for a single environment. When no sediment data are available, velocity should be predicted directly rather than, for example, predicting porosity and then velocity. Velocity is predictable within 1 to 2 percent in most environments.

5. Mean size of mineral grains has one of the best empirical relationships with velocity. This is important because size analyses can be made on dried sediment, and much sediment-size data on present charts can be related to velocity.

6. The most important property for predicting *in situ* sediment velocity is the ratio, velocity in sediment/velocity in seawater, because this ratio is the same in the laboratory as in the sea floor.

7. Porosity and density are the best indices to sediment impedance and to density $\times$ (velocity)^2.

8. Rayleigh reflection coefficients and bottom losses at normal incidence are easily computed from laboratory measurements. These laboratory values are very close to those actually measured at sea, apparently because sediment rigidity is so low that the Rayleigh fluid/fluid model is a close approximation to reality.

9. There is no usable relationship between velocity and shear strength (cohesion) as measured in soil-mechanics tests. This is apparently because cohesion from a static test cannot be compared with dynamic ridigity.

10. No variations of velocity were found with direction of measurement (no anisotropic velocity relationships) in cored sediments, and none is predicted for the upper few hundred meters in sea-floor sediments.

RECOMMENDATIONS

1. Continue studies of the mass physical properties of sediments from all environments to determine the parameters and statistical variations of these properties. Both laboratory and *in situ* measurements should be made to determine and refine suitable corrections (laboratory to *in situ*). Laboratory studies are needed because they are the basis of *in situ* predictions.

2. Consider measurements of the mass properties as discussed in the report to be a routine part of core laboratory procedures, especially for surveys of strategic areas.

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TABLES

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PREFACE

This report is Part I in a series of Technical Publications on sound velocity, elasticity, and related properties of marine sediments in three major environments of the North Pacific: continental terrace (shelf and slope), abyssal plain (turbidite), and abyssal hill (pelagic). It details the measurement and computation of the mass physical properties of the sediments and discusses their empirical relationships.

Part II (TP 144) will be concerned with the elastic properties of the sediments, including elastic models, and the measurement and computation of elastic constants.

Part III (TP 145) will be concerned with the prediction and computations of *in situ* physical properties.

Throughout each report, references are made to the other two studies, as appropriate.

INTRODUCTION

Before the early 1950's, the only information on compressional-wave velocity (hereinafter called "sound velocity" or "velocity") in deep-water marine sediments was derived from a very few measurements made by explosive seismology. Usually, in refraction work, no usable returns were recorded from the soft, unlithified sediments of the sea floors.

In the early 1950's, laboratory studies of sound velocity in marine sediments began at Cambridge (Laughton, 1954, 1957), at the Navy Electronics Laboratory (Hamilton, 1956; Hamilton *et al.*, 1956; Shumway, 1960), and at the Lamont Geological Observatory (Sutton, *et al.*, 1957; Nafe and Drake, 1957). Nafe and Drake (1963) have summarized this field through about 1960. Since 1960, laboratory measurements on sea-floor samples have been published by Parasnis (1960), Hamilton (1963, 1965), Ryan *et al.* (1965), Horn *et al.* (1968a), and Schreiber (1968).

Apparently the first *in situ* measurements (other than those of seismology) were made by Wood and his colleagues in the tidal mud flats at Emsworth during World War II (referred to in Wood and Weston, 1964). The advent of scuba techniques allowed measurements to be made in the sea floor to depths of about 45 meters (Hamilton *et al.*, 1956). Use of deep-diving research submersibles such as the bathyscaph TRIESTE allowed extension of these measurements to about 1200 meters (Hamilton, 1963); this report includes some results of recent *in situ* measurements from DEEPSTAR 4000 and additional scuba-diving stations.

During the period 1964 to 1968, laboratory and *in situ* measurements were made of the velocity of compressional and shear waves, density, porosity, and other mass physical properties of marine sediments from several major sedimentary environments in the North Pacific and adjacent areas. These measurements allow new insights into environmental differences, empirical relationships, and elastic properties of marine sediments. The environments include the continental shelves and slopes off North America and Asia, the deep-sea abyssal hills, and the abyssal plains in the deep basins of the Bering, Okhotsk, Japan, South China, Celebes, and Sulu Seas. Calcareous ooze is the only major sediment type not represented in the present study.

In situ measurements of shear- and compressional-wave velocities were made in continental borderland sediments off San Diego from the research submersible DEEPSTAR and, in shallow water, from a diving boat. Laboratory measurements were made on samples taken in the *in situ* program, by coring from a surface ship, and by special samplers attached to a lowered camera.

During the past few years, wide-angle reflection techniques using sonobuoys have added important new information on velocity gradients to the growing data on sound velocity in marine sediments (Houtz and Ewing, 1963, 1964; Houtz *et al.*, 1968). Projects to drill through the sediment layers in the deep sea will strongly complement this type of wide-angle reflection data, and serve to identify the layers in which velocity data have been obtained. The present deep-sea drilling project began operations in the Atlantic in the summer of 1968. Information on sound velocity from the Preliminary Mohole (Guadalupe Site) has been published (Hamilton, 1965).

A very important source of information on compressional and shear waves, and on elastic and other properties, can be found in the literature of soil mechanics – a source often overlooked by geologists and geophysicists. Papers by Hardin and Richart (1963), and by Richart and Whitman (1967) contain discussions and comprehensive bibliographies.

As a result of the work cited above, the velocity of compressional waves, and density, in the most common sediment types are reasonably well known. There is a small amount of data on the velocity of shear waves in unlithified marine sediments from refraction seismology (Nafe and Drake, 1957). *In situ* determinations of shear waves in marine sediments were made by Davies (1965), Bucker *et al.* (1964), and in measurements of Bucker reported in Hamilton *et al.* (1969); these latter measurements were made from the research submersible DEEPSTAR off southern California.

The study reported here was concerned with the mass physical properties of marine sediments from three major environments in the North Pacific and adjacent areas. Within each environment studied, data are presented for each of the major sediment types except calcareous ooze. Subsequent sections list the properties of these sediments and their interrelationships, and describe the techniques used in measuring and computing them.

METHODS

Source of Samples

The sediment samples on which this report is based are from the following general areas or environments.

Continental Terrace (Shelf and Slope) Environment:

1. Slope into Middle-America Trench (cores by D. R. Ross).

2. San Lucas Fan, Baja California (cores by W. R. Normark).

3. Continental Borderland off Southern California (nearshore to deep basins); cores and *in situ* samples from scuba diving and from deep-diving submersibles; box cores from La Jolla Fan (W. R. Normark).

4. Asian continental shelf and slope (cores; Shipek sampler)

Abyssal Hill (Pelagic) Environment:

1. Deep, North Pacific Basin (cores, including those taken off Mexico by D. R. Ross and R. L. Larson).

Deep-water, Abyssal Plain (Turbidite) Environment:

1. Middle America and Japan Trench floors (cores).

2. Deep basins peripheral to the Pacific Basin, including the Aleutian, Okhotsk, Japan, South China, Celebes, and Sulu Basins (cores).

Early in the study it was determined that for this suite of samples no distinctions could be made between the continental-shelf and continental-slope sediments; consequently, these were combined under "Continental Terrace." After comparison, "red clay" samples from distal ends of aprons around ancient and modern islands ("archipelagic aprons") were included in the abyssal-hill (pelagic) environment. The few island shelf and slope sediments were placed in the "terrace" category.

The category "deep-water, abyssal-plain (turbidite) environment," as discussed here, requires some explanation. All of the samples in this environment are either from deep trench floors or from the central parts of the deep, flat basins of the seas peripheral to the Pacific Basin; for example, the Aleutian Basin (Bering Sea), and the Japan Basin (Japan Sea). In these areas, the sediment surface is usually a high-porosity, fine-grained silt-clay, overlying buried layers of sand-silt (including volcanic ash). Because the samples herein reported are from the upper 30 cm, these coarser-grained, higher-velocity layers are not adequately represented. The properties of these layers will be the subject of a special section in Part III (Prediction). However, the number of high-porosity samples allows comparisons with the high-porosity sediments from the other environments.

The exact locations of some of the samples are classified by the Navy, but these locations are not essential to basic understanding of the interrelationships of the properties.

The necessity for precise data in computing elastic constants (Part II) precluded the use of the basic files and data on which Shumway (1960) based his report. These contained numerous although small errors in density and porosity, as well as uncertainties in velocity values which may have been caused by sediment-structural disturbance due to coring, transportation, and preservation of samples, and by sample extrusion into the resonance chamber used to make the measurements. However, to supply some data on medium sands (not represented in the author's measurements), 11 samples from Shumway (1960) are included in table 1, but are not used in any computations of elastic constants. With a few exceptions (Hamilton, 1963, 1965), all other measurements have not been previously published.

Sediment-Sampling Methods

In general, the following methods were used to obtain sediment samples.

1. Coring from a surface ship. Cores were taken mostly by specially designed, thin-walled gravity corers which took samples of the upper 6 feet of sediment.

2. Samplers attached to a lowered camera (Shipek, 1965).
   a. Two plastic liners, 1 foot long and 2-3/4 inch ID, took excellent samples of the upper 1 foot of sediment.
   b. "Cores" 2 to 3 inches long were taken by hand from the bucket of the Shipek Sampler.

3. *In situ.*
   a. Cores 1 foot long were taken by scuba divers, using a plastic tube as a core barrel.
   b. Cores 1 to 3 feet long were taken from a submersible (using only a plastic liner, or tube, as a core barrel) during sound-velocity measurements.

Sample Selection

This report is concerned with measurements from the upper 30 cm, only, of the sea floor. Because of uncertainties concerning disturbance to the sediment structural strength caused by the coring process (a critical matter in computations of elastic properties), several hundred measurements deeper than 30 cm were not used. This selection does not imply that measurements from deeper depths in cores, reported by others, are in doubt; a well-designed piston corer, used by experienced personnel aboard ship, can take good samples at greater depths, as can a properly designed, thin-walled gravity corer. Another reason for using the upper 30 cm of sediment from cores was that many of the samples from the Shipek equipment, and collected *in situ* by divers and from submersibles, were from this depth interval. Thus, all samples reported are from the same interval, and all samples within this interval are reported.

Laboratory and *In Situ* Methods

Sound velocity was measured by two methods. The pulse technique (operating at about 200 kHz) was used in the laboratory; estimated margins of error in the pulse measurements were ±3 m/sec in clays, and 5 m/sec in sands. In the *in situ*, submersible/scuba-diver program, sound velocity was measured over a 1-meter path between probes inserted into the sediment to depths of 10 to 90 cm. Without disturbing the probes, measurements were then made at 14, 7, and 3.5 kHz. During this same *in situ* program, Bucker, Keir, and Whitney measured Stoneley-wave velocities (from which shear-wave velocities can be computed; see Bucker *et al.*, 1964; Hamilton *et al.*, 1969).

All sound velocities measured in the laboratory were corrected to 23°C. The *in situ* measurements were corrected to 23°C and 1 atmosphere pressure from the actual temperatures and pressures taken at the time of measurements. These corrections were made by using Tables of Sound Speed in Sea Water (NAVOCEANO, 1962), a technique shown to be valid in the TRIESTE program (Hamilton, 1963). Measurements of sediment temperatures (when sound velocities are measured), and correction to a common value, are critically important in comparisons of differences in velocities between sediment types and environments. This is because variations in "room temperature" can cause changes in sediment velocity on the order of 20 to 30 m/sec, which can be more than some variations between sediment types or environments. Even greater variations may be caused by measuring sediment just after coring the cold sea floor, or after removal of samples from a refrigerator.

Bulk, saturated densities were determined by the weight-volume method. Porosities were determined after oven-drying at 105°C, and corrections to allow for the amounts of dried salts in the dried mineral residues. For fine distinctions between sediment types, the salt correction should be made; it amounts to about 1 percent (additional) porosity at porosities around 80 percent. The bulk densities of the mineral solids were determined by the pycnometer method.

Size analyses were made on clays using the pipette method, and on sands by sieving and use of the Emery Settling Tube. The results, when plotted, were used to determine the median and mean diameters of grains. The mean diameters were determined by averaging the 16th, 50th, and 84th percentiles (Folk and Ward, 1957).

The ratio of sound velocity in sediment to sound velocity in water was determined by dividing the sediment velocity by the velocity of sound in seawater at 23°C, 1 atmosphere pressure, and of the same salinity as that of the bottom water at the sampling site. As discussed below, this ratio is of considerable importance in predicting *in situ* velocity values, because it is the same in the laboratory as it is *in situ*.

Sediment nomenclature followed that of Shepard (1954), except that, within the sand sizes, the names for the various grades of sand followed the Wentworth Scale, as follows:

Sediment Name	Median Diameter, mm	ϕ Scale
<u>Sand</u>	2.0 to 0.0625	-1.0 to 4.0
Very coarse	2.0 to 1.0	-1.0 to 0.0
Coarse	1.0 to 0.5	0.0 to 1.0
Medium	0.5 to 0.25	1.0 to 2.0
Fine	0.25 to 0.125	2.0 to 3.0
Very fine	0.125 to 0.0625	3.0 to 4.0
<u>Silt</u>	0.0625 to 0.004	4.0 to 8.0
<u>Clay</u>	less than 0.004	greater than 8.0

Both median and mean grain diameters are tabulated in millimeters. In the scatter diagrams the grain sizes are shown in logarithmic phi-scale (Inman, 1952).

The data were examined statistically in computer programs as follows: (1) the arithmetic mean (average), standard deviation, and standard error were computed for each individual property within each environment and for each sediment type; (2) regression lines and their equations and errors were computed for the illustrated diagrams; and (3) various groups of data were examined to determine any significant differences. The formulas used in these computations are listed and discussed by Arkin and Colton (1965) and Griffiths (1967); they were

$$\text{Arithmetic mean (or average), } \bar{X} = \frac{\Sigma(X)}{n} \tag{1}$$

$$\text{Standard deviation (of the sample), } \hat{\sigma} = \left(\frac{\Sigma(X - \bar{X})^2}{n - 1}\right)^{1/2} \tag{2}$$

$$\text{Standard error (of the mean), } \hat{\sigma}_{\bar{X}} = \frac{\hat{\sigma}}{(n)^{1/2}} \tag{3}$$

$$\text{Least Significant Difference, LSD} = t\left(\hat{\sigma}^2_{\bar{X}_1} + \hat{\sigma}^2_{\bar{X}_2}\right)^{1/2} \tag{4}$$

where

X = individual item in data

n = number of items

LSD = the minimum difference (Least Significant Difference) necessary to demonstrate a statistically significant difference between two samples

t = the number of standard errors required to reach a given confidence level [*e.g.*, 3 (standard error) = confidence level of 99.7 percent]

$\hat{\sigma}_{\bar{X}_1}$ and $\hat{\sigma}_{\bar{X}_2}$ = standard errors of the two samples

RESULTS

In this report the results of the study are reported in scatter diagrams and tables. Table 1 lists the averaged (arithmetic mean) properties of sediments from the continental terrace (shelf and slope) environment and table 2 lists the averaged properties of sediments from the abyssal-hill (pelagic) and abyssal-plain (turbidite) environments. Statistical "standard errors" are listed for some of the more important properties.

Scatter diagrams which illustrate the more useful relationships between sediment mass physical properties were selected from the large number constructed for study; many of these are illustrated and discussed in the following section. The best empirical entries into the data to obtain a desired property are discussed in Appendix A. Regression equations for some of the more important, illustrated data are in Appendix B.

TABLE 1: SEDIMENT PROPERTIES, CONTINENTAL TERRACE (SHELF

Sediment Type	No. Samples	Grain Diameter Mean (mm)	Grain Diameter Median (mm)	Sand (%)	Silt (%)	Clay (%)	Bulk Grain Density (g/cc)	Density (g/cc) Avg.	Density (g/cc) SE	Porosity (%) Avg.	Porosity (%) SE	Veloc Avg.
Sand												
Coarse	2	0.530	0.520	100.0	---	---	2.71	2.03	---	38.6	---	1836
Medium	12	0.376	0.356	99.8	0.2	---	2.70	2.01	0.009	39.7	0.46	1749
Fine	9	0.153	0.171	88.1	6.3	7.1	2.70	1.98	0.024	43.9	1.29	1742
Very fine	3	0.090	0.094	83.9	13.0	2.9	2.74	1.91	---	47.4	---	1711
Silty sand	11	0.073	0.126	65.0	21.6	13.4	2.71	1.83	0.025	52.8	1.55	1677
Sandy silt	6	0.036	0.051	34.5	51.2	14.3	2.75	1.56	---	68.3	---	1552
Sand-silt-clay	17	0.018	0.041	32.6	41.2	26.1	2.71	1.58	0.030	67.5	1.66	1578
Clayey silt	40	0.006	0.011	6.1	59.2	34.8	2.71*	1.43	0.016	75.0	0.87	1535
Silty clay	17	0.003	0.004	5.3	41.5	53.6	2.69	1.42	0.013	76.0	0.74	1519

Notes: Laboratory values: 23°C, 1 atmosphere.

Density: saturated, bulk density; porosity: salt-free; ratio: velocity in sediment/velocity in seawater at 23°C, 1 atmosphere, and salinity of se

SE: standard error of the mean.

$\rho_2 V_2$ = sediment impedance, g/cm^2 sec $\times 10^5$.

$\rho_2 (V_2)^2$ = sediment density $\times$ (velocity)^2, g/cm sec^2, or dynes/cm^2 $\times 10^{10}$.

$$R = \text{Rayleigh reflection coefficient at normal incidence} = \frac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1}$$

BL = 20 log R, bottom loss, dB.

ρ_1, V_1: seawater density, velocity; ρ_2, V_2 : sediment density, velocity.

A

NENTAL TERRACE (SHELF AND SLOPE) ENVIRONMENT

cc)	Porosity (%)		Velocity (m/sec)		Ratio		$\rho_2 V_2$		$\rho_2 (V_2)^2$		R		BL	
SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE
--	38.6	---	1836	---	1.201	---	3.7347	---	6.8577	---	0.4098	---	7.8	---
.009	39.7	0.46	1749	6	1.144	0.004	3.5087	0.020	6.1380	0.050	0.3835	0.002	8.3	0.05
.024	43.9	1.29	1742	10	1.139	0.006	3.4433	0.040	6.0888	0.117	0.3749	0.005	8.6	0.12
--	47.4	---	1711	---	1.121	---	3.2645	---	5.5946	---	0.3517	---	9.1	---
.025	52.8	1.55	1677	9	1.096	0.006	3.0633	0.050	5.1387	0.102	0.3228	0.008	9.9	0.20
--	68.3	---	1552	---	1.015	---	2.4201	---	3.7587	---	0.2136	---	13.5	---
.030	67.5	1.66	1578	9	1.032	0.006	2.4939	0.059	3.9420	0.113	0.2504	0.010	12.1	0.36
.016	75.0	0.87	1535	3	1.004	0.002	2.1989	0.026	3.3782	0.045	0.1767	0.012	15.2	0.66
.013	76.0	0.74	1519	3	0.994	0.002	2.1571	0.024	3.2804	0.042	0.1586	0.005	16.1	0.29

°C, 1 atmosphere, and salinity of sediment pore-water.

B

TABLE 2: SEDIMENT PROPERTIES, ABYSSAL PLAIN (TURBIDITE) AND ABYSSA

Environment Sediment Type	No. Samples	Grain Diameter Mean (mm)	Grain Diameter Median (mm)	Sand (%)	Silt (%)	Clay (%)	Bulk Grain Density (g/cc)	Density (g/cc) Avg.	Density (g/cc) SE	Porosity (%) Avg.	Porosity (%) SE	Velocity (Avg.
Abyssal Plain (Turbidite)												
Sandy silt	1	0.017	0.017	19.4	65.0	15.6	2.46	1.65	---	56.6	---	1622
Silt	1	0.016	0.018	7.2	79.5	13.3	2.47	1.60	---	60.6	---	1634
Clayey silt	15	0.005	0.006	7.6	50.3	42.1	2.61	1.38	0.029	78.6	1.53	1535
Silty sand	35	0.002	0.003	2.9	36.1	61.3	2.55	1.24	0.010	85.8	0.49	1521
Clay	2	0.001	0.001	0.1	20.3	79.6	2.67	1.26	---	85.8	---	1505
Abyssal Hill (Pelagic)												
Clayey silt	3	0.0035	0.0053	3.3	50.0	46.7	2.58	1.41	---	76.4	---	1531
Silty clay	32	0.0026	0.0023	2.6	32.9	65.2	2.71	1.37	0.014	79.4	0.77	1507
Clay	6	0.0015	0.0013	0.6	20.7	78.9	2.76	1.42	0.023	77.5	1.35	1491

Notes: Laboratory values: 23°C, 1 atmosphere.

Density: saturated, bulk density; porosity: salt-free; ratio: velocity in sediment/velocity in seawater at 23°C, 1 atmosphere, and salinity of sed

SE: standard error of the mean.

$\rho_2 V_2$ = sediment impedance, g/cm^2 sec X 10^5.

$\rho_2 (V_2)^2$ = sediment density X (velocity)^2, g/cm sec^2, or dynes/cm^2 X 10^{10}

R = Rayleigh reflection coefficient at normal incidence = $\dfrac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1}$

BL = 20 log R, bottom loss, dB.

ρ_1, V_1: seawater density, velocity; ρ_2, V_2: sediment density, velocity.

 A

AIN (TURBIDITE) AND ABYSSAL HILL (PELAGIC) ENVIRONMENTS

/cc)	Porosity (%)		Velocity (m/sec)		Ratio		$\rho_2 V_2$		$\rho_2 (V_2)^2$		*R*		*BL*	
SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE
·	56.6	---	1622	---	1.061	---	2.6795	---	4.3462	---	0.2627	---	16.6	---
·	60.6	---	1634	---	1.069	---	2.6111	---	4.2666	---	0.2208	---	12.0	---
029	78.6	1.53	1535	2	1.003	0.001	2.1154	0.048	3.2475	0.076	0.1506	0.011	16.7	0.76
010	85.8	0.49	1521	2	0.994	0.001	1.8919	0.014	2.8769	0.021	0.0944	0.004	20.7	0.32
··	85.8	---	1505	---	0.985	---	1.8911	---	2.8449	---	0.0941	---	20.6	---
··	76.4	---	1531	---	1.000	---	2.1615	---	3.3091	---	0.1596	---	15.9	---
.014	79.4	0.77	1507	2	0.985	0.001	2.0674	0.021	3.1155	0.032	0.1412	0.005	17.2	0.31
.023	77.5	1.35	1491	1.4	0.975	0.001	2.1118	0.035	3.1491	0.050	0.1477	0.008	16.7	0.54

23°C, 1 atmosphere, and salinity of sediment pore-water.

B

DISCUSSION AND CONCLUSIONS

General

Some of the empirical relationships discussed below are of considerable practical use in predicting sound velocity and other properties, but it should be emphasized that compressional- and shear-wave velocities are elastic properties of the sediment body. They are transmitted because of the elasticity of the sediment, and expressions such as "the dependence of velocity on porosity" are not literally true in a physical sense. The "nonelastic" properties are effective in determining sound velocity only in the effect that they have on elasticity of the sediment. For this reason, discussions of the underlying causes of most of the empirical relationships will be reserved until the discussion of the elastic properties (Part II).

Porosity and density will be discussed first because of their importance in empirical and elastic relationships with other mass physical properties. When both density and porosity data are available, and it is desired to enter diagrams of the two properties vs. sound velocity (or other properties), density should be preferred because of the laboratory procedures used in determining the two values. Density is usually the first property determined for a saturated sediment (simply, the weight/volume relationship). Following a density measurement, the sediment is oven-dried until there is no more weight loss due to further drying at a stated temperature (usually 105° or 110°C). The porosity is then computed by using values for the weight of the evaporated water per unit volume (assuming that 1 gram equals 1 cc of water). To determine a truer value, the weight of dried salts (weighed with the dried minerals) must be considered. When sediments contain appreciable amounts of clay minerals, the amount of evaporated water is a function of drying temperature; thus, both bulk grain density (dry density) and porosity are functions of drying temperatures in clayey sediments (see, for example, Igelman and Hamilton, 1963). For these several reasons, density values are apt to be more accurate than porosity values, and there is less chance of error in using density as an index property to associated properties.

Porosity

In a saturated sediment, the volume of voids (or pore space) occupied by water is expressed as porosity, n, or void ratio, e (more common in the literature of soil mechanics):

$$n = \frac{\text{Volume of voids}}{\text{Total volume}} = \frac{e}{1+e} \tag{5}$$

$$e = \frac{\text{Volume of voids}}{\text{Volume of solids}} = \frac{n}{1-n} \tag{6}$$

Porosity is usually expressed in percent; void ratio, as a decimal fraction.

The amount of pore space in a sediment is the result of a number of complex interrelated factors; most important are the size, shape, distribution, and mineralogy of the solid grains.

Equal-sized spheres, if regularly packed, have porosities ranging from 47.6 percent in the loosest arrangement to 26 percent in the densest arrangement; in these arrangements, porosity is independent of sphere diameter (Graton and Fraser, 1935). Although, in modern marine sediments, sands are not equal-sized spheres, there are differences in porosity due to loose or dense packing.

In natural sedimentary processes, sand-sized grains in suspension sink rapidly to the bottom (or are carried along the bottom) and assume positions among other grains under the influence of gravity and water motions. When finer grains are not present, a single-grained structure is formed (fig. 1A). When finer-sized grains (silt and clay sizes) are also present, they occupy spaces (pores) between the larger grains, the porosity is decreased, and a mixed-grained structure is formed (fig. 1B). In these structures the larger grains are usually in direct contact. Marine sands will normally vary in porosity between about 35 to about 50 percent, with averages around 40 percent for medium sands and 44 percent for fine sands. These sands, with their grain-to-grain contacts, have distinctly different skeletal or mineral structures from those of the high-porosity, silt-clay sediments, and should be studied separately.

Grain shape is an important factor in porosity. When platy minerals such as biotite are present in sands, they are apt to bridge between grains of quartz and feldspar, and cause increases in porosity having little to do with grain size (fig. 1C); this effect has been demonstrated in the laboratory (Terzaghi and Peck, 1948).

The sediment structure formed by fine silt and clay-sized particles is controlled by the adsorbed water around the grains and interparticle forces (discussed in Part II). When these particles fall to the sea floor they are not controlled by gravity (as are sand grains), but are apt to stick to the grain on which they first alight and be held there by interparticle forces to form three-dimensional structures of the honeycomb or "cardhouse" types (figs. 1D, 1E). Bowles (1968) studied ultrathin sections of "undisturbed" marine sediments from an abyssal plain, and the continental slope and shelf of the Gulf of Mexico, and concluded that the microstructures did not conform entirely to either the cardhouse or honeycomb structures. Bowles suggests that it may be more accurate to say that the microstructure is characterized by a loose, open framework of randomly oriented particles.

Most investigators agree that the clay-sized particles are not in a mineral-to mineral contact, but are in contact through their adsorbed water layers (Yong and Warkentin, 1966). Overburden pressures are transmitted through these contacts, and the cohesion between particles results in appreciable shear strengths.

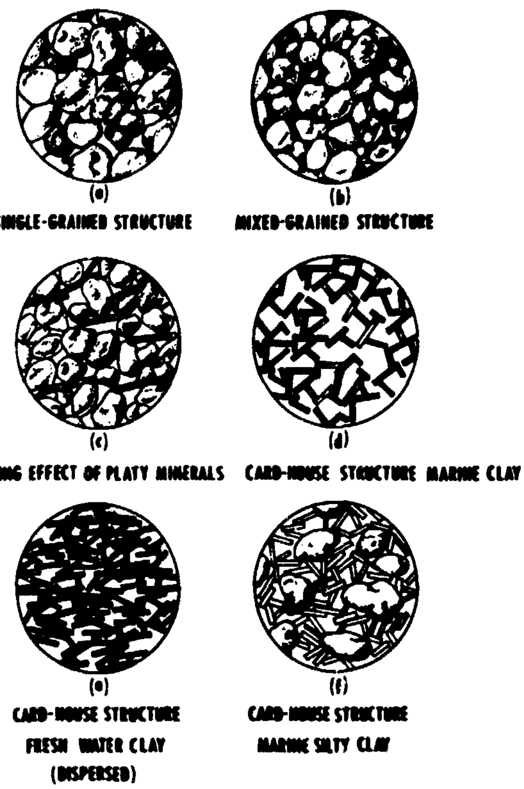

Figure 1. Common sediment structures (from soil-mechanics literature).

In summary, differences in porosity can be the results of size, size distribution (sorting), shape, and mineralogy of grains and their packing, rates of deposition, sediment structure, and other factors. The interrelated effects of these factors usually result in a general decrease in porosity with increasing grain size (fig. 2).

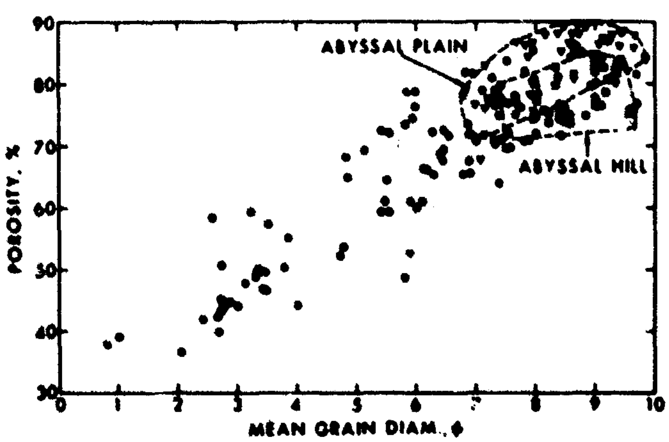

Figure 2. Mean diameter of mineral grains vs. porosity. Round dots are continental terrace (shelf and slope); squares, abyssal hill (pelagic); and triangles, abyssal plain (turbidite) environments.

In the sediments of this study, there is a significant difference between the porosities of abyssal-plain (turbidite) sediments and sediments from the abyssal-hill (pelagic) environment. Silty clay has an average porosity of 85.8 percent in abyssal-plain sediments, and 78.8 percent porosity in abyssal-hill red clay (table 2). At the same mean grain diameter and the same percentage of grains less than 2 or 4 microns, the turbidites still have greater porosities (figs. 2, 3); sorting can also be eliminated as a factor in explaining the difference. The basic difference in porosity of sediments between the two environments appears to be in mineralogy.

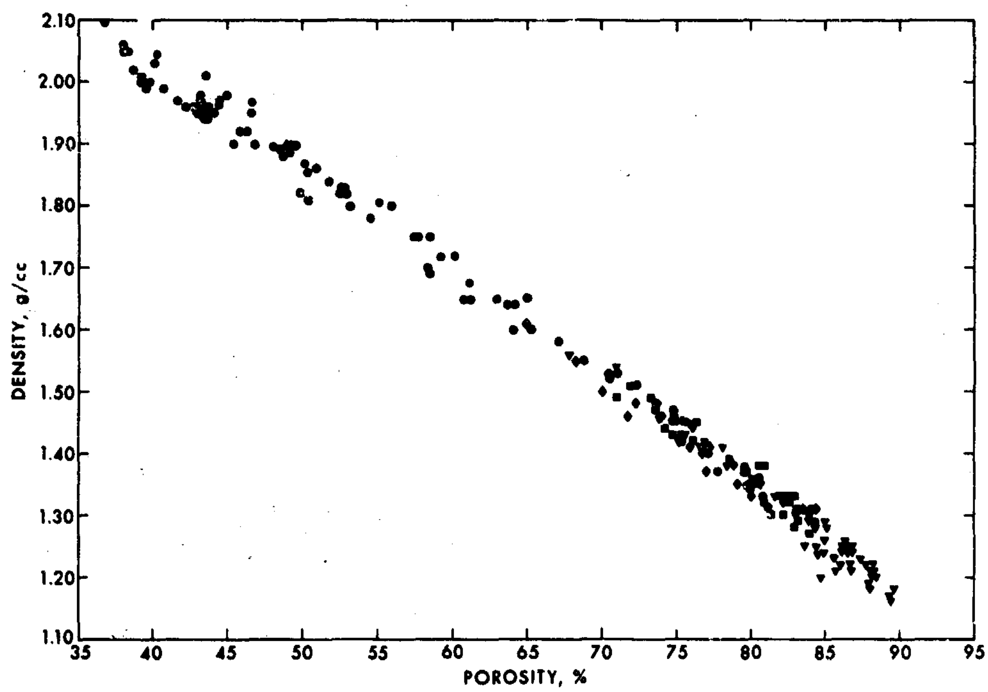

Figure 3. Porosity vs. saturated bulk density, all environments.

The clay minerals were identified by X-ray diffraction in 75 of the samples (analyses by W. R. Bryant Associates). These measurements will be presented in another report, but some of the general results are pertinent to this study. Griffin and Goldberg (1963) noted that montmorillonite is concentrated near the land masses in the North Pacific, and is apt to be associated with volcanics. Rateev *et al.* (1968) have recently summarized clay mineral distributions in the world's oceans, and noted that montmorillonite, kaolinite, and gibbsite occur in highest percentages in the tropical humid zone, with greater development in the zone of tropical lateritic soils. Although this group has an equatorial type of distribution, montmorillonite may form aclimatic and azonal concentrations where it is influenced by the volcanic materials from which it originates. The clay-mineral distributions of this study are in accord with these general statements. Analyses of the Central Pacific red clays indicated an average of 2.1 percent montmorillonite; the peripheral basins, 2.6 to 12.8 percent with the heaviest concentrations near the Philippine Islands. The abyssal-plain sediments averaged more than twice the amount of montmorillonite (about 6 percent of the total sample) than did the abyssal-hill sediments.

The clay mineral montmorillonite has more specific surface area per unit mass, by more than an order of magnitude, than either illite or kaolinite, and thus adsorbs more water, relative to its mass, than do the other common clay minerals. The result is that montmorillonite has a higher water content, or amount of pore water, and thus higher porosities than either illite or kaolinite (Grim, 1961; Meade, 1964). Grim (1962) has shown that 5 to 10 percent montmorillonite in illite or kaolinite samples causes substantial increases in liquid limit, or water content; computations indicate the increases in porosity would be of the order of 5 to 13 percent. These properties of montmorillonite, and its concentration in the abyssal plain sediments, are apparently important factors in the porosity difference.

Saturated Bulk Density

The saturated bulk density of a unit volume of gas-free sediment has two components: mineral grains, and water within the pore spaces. The relationship between these factors and porosity is

$$\rho_{sat} = n\rho_w + (1\text{-}n)\,\rho_s \qquad (7)$$

where

ρ_{sat} = saturated bulk density

n = porosity (fraction); volume of voids/total sample volume

ρ_w = density of pore water

ρ_s = bulk density of mineral solids

Averaged bulk densities of mineral solids, saturated bulk densities, and porosities of the sediments are listed in tables 1 and 2 for the sediment types within each environment.

The relationships of saturated bulk density to porosity for the sediments reported here are shown in figure 3. This type of diagram is very useful in predicting either density or porosity, given either value. It is also useful in cross-checking laboratory or literature data for accuracy of reported or computed densities or porosities; for example, a line drawn from the density value for seawater (at 100 percent porosity) through the plotted point of density vs. porosity for a sample, will cross the zero porosity line at the bulk density of the mineral grains, thus graphically solving for bulk density of minerals. Density-porosity values which are probably in error can be spotted at a glance in that they indicate impossible, or improbable, mineral-grain densities.

DENSITY OF PORE WATER

Sigma-T tables (*e.g.*, NAVOCEANO SP-68, 1966) list "laboratory" values for the density of seawater at various temperatures, salinities, and 1 atmosphere of pressure. For example, in the Pacific Ocean, below water depths of 1500 m, the salinity varies between 34.65 and 34.68 ppt (Defant, 1961). From the listed values for Sigma-T, assuming that the bottom-water salinity is the same as that in the upper 30 cm of sediment, the density of pore water in the laboratory sediment samples from the deep Pacific would be about 1.0237 g/cc, at 23°C.

BULK DENSITY OF MINERAL GRAINS

Most of the mineral grains found on continental shelves and slopes, and in adjacent abyssal plains, have been transported through air-water paths or along the sea floor from adjacent islands and continental areas. Most of these mineral grains will be products of the rocks and sediments of the continental or island source areas. Although pelagic particles (those deposited from the water mass) may be evenly deposited over some of these areas, these particles are masked (near the source areas) by the large volume of terrigenous minerals; further away, pelagic components may be important.

Because of the geographic variations in pelagic organisms such as diatoms and radiolaria (silica) and Foraminifera (calcium carbonate), and variations in island and continental rocks and minerals of sediments, the bulk grain densities of one basin or area near sediment sources cannot be safely used in computations involving grain densities for another basin or area.

In far northern areas, where diatoms flourish, the deposition of low-density silica markedly affects the average grain densities. For example, the deep Bering Sea sediment (Aleutian Basin) has a relatively low, average grain density of 2.44 g/cc. The table below lists averages for the deep, central portions of the various basins. These averages are merely indicative; there are too few samples to be definitive.

Basin	Avg. Bulk Density of Minerals (g/cc)	No. of Samples in Average
Aleutian	2.44	8
Okhotsk	2.41	3
Japan	2.60	14
South China	2.70	5
Celebes	2.63	3
Sulu	2.68	3
Japan Trench (floor)	2.58	1

In abyssal-hill, deep-sea clay areas, there is less variation in grain densities. In the deep North Pacific Basin, well away from land areas, an average grain density for 21 samples was 2.735 g/cc; the average for the total environment (table 2) is 2.70 g/cc.

For all samples, the least grain density was from the deep Bering Sea (2.31 g/cc); the highest grain density was 2.80 g/cc from abyssal-hill red clay. The significant difference in average grain densities of the two deep-water environments (abyssal plain, 2.56 g/cc; abyssal hill, 2.70 g/cc) is a reflection of environmental control. Because many of the abyssal-plain sediments were from northern areas (where diatoms flourish), the average grain density of this general environment is skewed to the low side for the sediments reported here.

Density-Porosity Relationships

Among all of the deep-water samples from the abyssal hills and plains, the lowest saturated bulk density was 1.16 g/cc from the Okhotsk Basin; the highest was 1.65 g/cc in a silty layer in the turbidites of the Japan Basin. The average saturated densities of these two environments show a significant difference: 1.30 g/cc for the abyssal plains, and 1.39 g/cc for the abyssal hills. This difference is due to the previously discussed mineralogy and porosities of the sediments, and differences in water densities (in the laboratory) are not significant.

As previously discussed, at the same grain size, abyssal-plain sediments are apt to have higher porosities than abyssal-hill sediments (fig. 2; table 2). If mineral-grain densities of both environments were the same, this would result in lower saturated bulk densities in abyssal-plain sediments. Because the abyssal-plain sediments have lower grain densities, the difference is even more marked: at the same grain size, in this suite of samples, the abyssal-plain sediment densities are distinctly lower than the abyssal-hill red clay densities (fig. 4). These interrelationships have important consequences in values of sound velocity and other elastic constants.

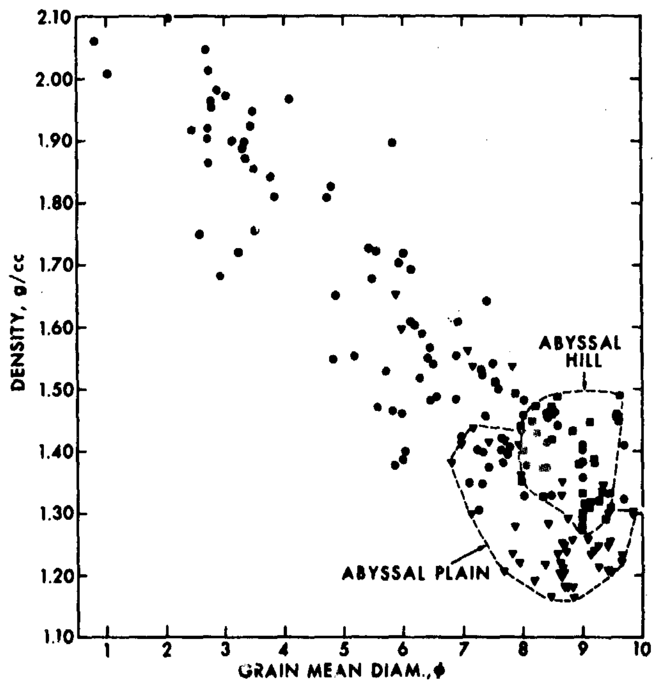

Figure 4. Mean diameter of mineral grains vs. saturated bulk density, all environments.

Sound Velocity

SOUND VELOCITY/POROSITY RELATIONSHIPS

The sound velocity/porosity relationship has received much attention because porosity is an easily measured property likely to yield predictable relationships with sound velocity. This is because porosity is the volume of water-filled pore space within a unit volume of sediment, and, in the sediment elastic system, the compressional-wave speed is largely determined by the dominant effect of the compressibility of pore water, rather than that of the mineral solids. This matter is discussed in a basic way in Part II (Elastic Properties).

Earlier studies have illustrated the general relationships between sound velocity and porosity over the full range of sediment porosities (Hamilton, 1956; Hamilton *et al.*, 1956; Sutton *et al.*, 1957; Nafe and Drake, 1963; Laughton, 1957; Horn *et al.*, 1968; Schreiber, 1968). In general, there is little change of velocity between 90 percent and about 75 percent porosity; in fact, in some environments, velocity may slightly decrease with porosity in this range. From about 75 to 80 percent, to the low porosities of sand, there is a rapid increase of sound velocity with decreasing porosity (figs. 5, 6).

Much of the earlier work was done on continental shelf and slope deposits, and data from the vast areas of the deep sea and deep peripheral basins were scarce. At present, however, data are becoming available from these deeper areas and it is now becoming possible to make important environmental distinctions. For the Atlantic and adjacent areas, some of these differences between environments and sediment types have been noted by Schreiber (1968) and by Horn *et al.* (1968). This report will show some of the environmental differences for the North Pacific and adjacent areas.

Figure 5 shows sound velocity vs. porosity for some higher-porosity sediments. The label "seawater" represents the velocity in seawater at 23°C, a salinity of 34.6 ppt, and 1 atmosphere pressure. It can be seen that most of these high-porosity sediments have sound velocities less than that in seawater. This has been shown to be true by all investigators who have studied high-porosity sediments, and substantiated by *in situ* measurements by divers, from submersibles, and by seismic work at sea.

An important observation, made possible by the accumulation of data from deep-sea sediments, is that many of the highest-porosity, deep-water sediments have sound velocities higher than most continental-shelf sediments at the same porosities. Figure 5 shows two curves often used to illustrate porosity-velocity relationships: Shumway (1960), and Wood (1941, as applied to marine sediments; Hamilton, 1956). Over the full range of porosities and velocities, various investigators (cited above) have demonstrated the valid, general relationships between sound velocity and porosity, but it can be readily seen that at this scale, without regard to environments, there is little usable relation between porosity and sound velocity in these higher-porosity sediments. However, areas in the diagram (fig. 5) including two of the major environments show definite environmental effects. Shumway's curve shows porosity-velocity relationships in a third environment. Because most of Shumway's samples were from the continental shelf and slope off southern California, his curve adequately defines only these sediments. The Wood equation is even farther away from the deep-water data points.

Most of the early studies (including my own) produced empirical equations and diagrams relating sound velocity and porosity (over the full range of porosities and sound velocities), which it was hoped could be used to predict sound velocity *in situ* in sea-floor sediments. These equations and diagrams have been of importance in narrowing the range of values for sound velocity, but none of them can be used, in general, to accurately predict sound velocity *in situ*, in a given locality, with the precision necessary for many studies in the fields of geophysics and military oceanography. For example, if one entered figure 5 at 80 percent porosity, without regard to environment, one would get a range of velocity values of about 50 m/sec (from 1490 to about 1540 m/sec), which is too great an error for use in underwater acoustics and for many geophysical problems. Entry with regard to environment reduces the error; for example, the range of velocity values in abyssal-plain turbidites at 80 percent porosity is about 30 m/sec. For these and other reasons, general diagrams and equations which relate laboratory values of porosity and sound velocity (over the full range of both properties in all environments) should be abandoned for use in predicting velocity, especially *in situ*. As noted below, some indices may be better than porosity, and diagrams or equations for particular environments and sediment types should be used. Direct prediction of velocity from tabulated averages is the best method when no sediment data are available. In addition, as fully discussed in Part III ("Prediction of *In Situ* Properties"), laboratory values of velocity require correction to *in situ* values.

The higher sound-velocity values for the deep-water silts and clays will be discussed in detail in Part II (Elastic Properties), but the main cause is apparently the more rigid sediment structure of deep-sea sediments due to interparticle bonding. This bonding involves a complex of relationships between van der Waal's and Coulombic

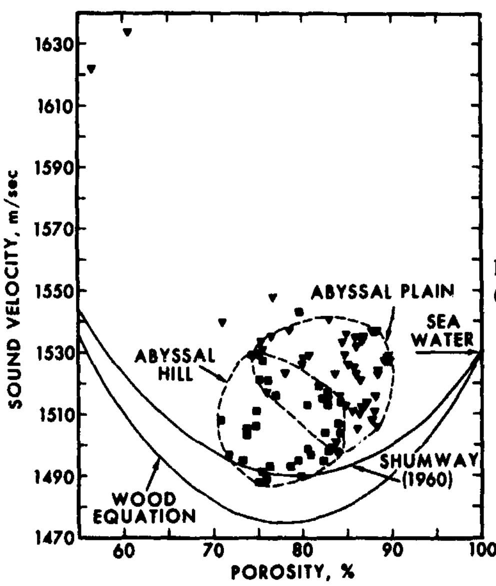

Figure 5. Porosity vs. sound velocity, abyssal hill (pelagic) and abyssal plain (turbidite) environments.

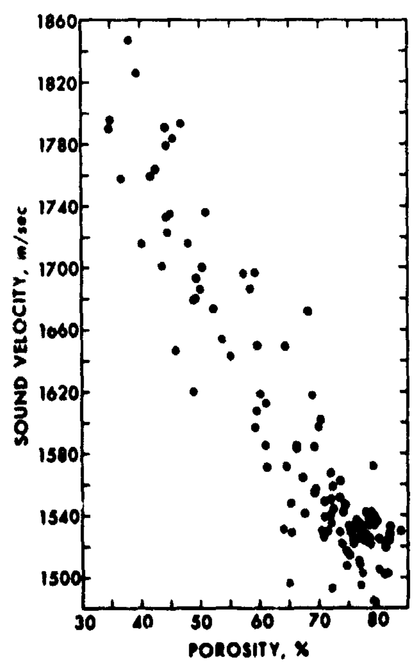

Figure 6. Porosity vs. sound velocity, continental terrace.

forces, geochemical alterations (such as cementation between grains from solution and redeposition of minerals within the sediments, and from deposition of authigenic minerals from seawater such as iron, manganese, phillipsite, and other species), rates of deposition, chemistry of interstitial waters, mineralogy, and other factors. This increase in structural strength of deep-water sediments has also been noted in soil-mechanics tests (Moore, 1964; Hamilton, 1964; Richards and Hamilton, 1967).

SOUND VELOCITY-DENSITY RELATIONSHIPS

The empirical relationships between sound velocity and saturated bulk density (figs. 7, 8) are similar to those for sound velocity and porosity because of the linear relationship between density and porosity. Figure 7 illustrates, as in the case for velocity-porosity, the importance of environmental differentiation; at any given density, the abyssal-plain sediments are apt to have higher velocities. Figure 7 also illustrates

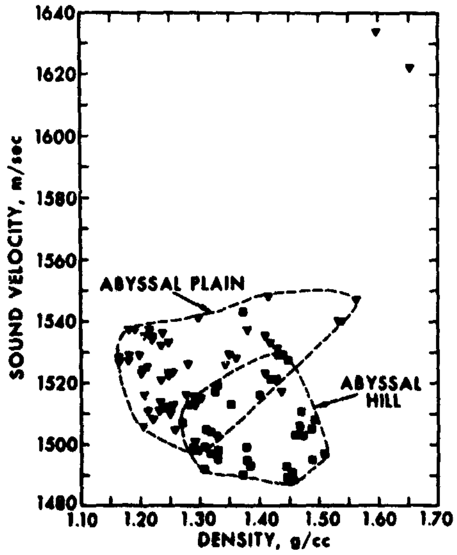

Figure 7. Density vs. sound velocity, abyssal hill and abyssal plain.

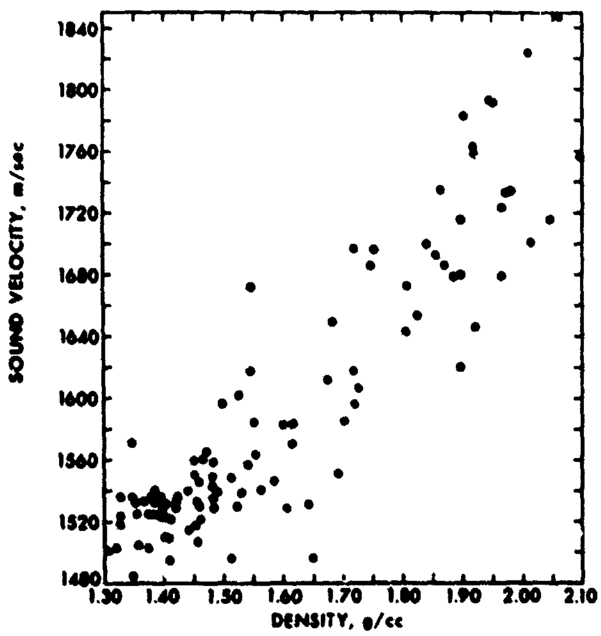

Figure 8. Density vs. sound velocity, continental terrace.

the averaged data of table 2: the abyssal plains, in general, have lower densities and higher velocities than the abyssal-hill sediments.

Density is one of the critical constants of elasticity in determining sound velocity; consequently, further discussion of the velocity-density relationships will be deferred to Part II.

SOUND-VELOCITY/SIZE-ANALYSIS RELATIONSHIPS

Data derived from laboratory size-analyses of sediments usually include mean and median mineral-grain diameters, the percentages of sand, silt, and clay, and various statistical parameters. An important finding is that some of these textural properties are among the best indices to empirical derivations of sound velocity. The relationships between sound velocity and the textural properties are of considerable importance for other reasons: much of the sediment data on charts, and in the literature of oceanography, include size analyses only, and size analyses can be made on dried sediment. The latter fact is especially important because density, porosity, and sound-velocity measurements are valid only on fully saturated sediments, but size analyses are the same whether one starts with wet or dry sediment. Thus, much of the size data in the literature can be used for estimating or predicting acoustic properties such as density, porosity, and sound velocity; many old cores, now partly or wholly dessicated, can be analyzed to obtain meaningful acoustic data.

As recently emphasized by Horn *et al.* (1968), mean grain size is a better index to other properties than median diameter because it is a better measure of the distribution of sizes. The mean grain size (usually an average of the 16th, 50th, and 84th percentiles) is not always obtainable in the usual laboratory procedures because the higher percentile may require an undue amount of extrapolation beyond 10 phi for the finer-grained, deep-water sediments (Schreiber, 1968); in this case, other percentiles may be used (*e.g.*, the 25th, 50th, and 75th).

In the samples of this report, the relationships of mean grain size to porosity, density, and velocity are in accord with previous work in Pacific sediments (Hamilton *et al.*, 1956; Shumway, 1960), and in sediments from the Atlantic and adjacent area (Sutton *et al.*, 1957; Schreiber, 1968; Horn *et al.*, 1968): with increasing grain size, porosity decreases and density and velocity increase. Empirically, mean grain size is an important index to porosity, density, and velocity (figs. 2, 4, 9). In many cases, data from sediment size analyses list median rather than mean grain size. In such a case, one might be forced to use the median as the mean. The averages of both median and mean grain size are listed in table 2, which serves to illustrate the differences between the two.

The effect of grain size on velocity is particularly empirical because its true effect on velocity is through certain elastic properties and through porosity and density; even in these latter properties it is only one of several significant factors (as previously discussed). Schon (1963) and Hardin and Richart (1963) studied sound velocity and other physical properties in controlled, laboratory studies and concluded that grain size has only a porosity-dependent influence on velocity.

When using mean grain size as an index property, it is important to consider the environment for which data are sought. This is illustrated by the diagrams of mean grain size vs. density, porosity, and velocity. At the same mean grain size, abyssal-plain turbidites have higher porosities and velocities and lower densities than abyssal-hill sediments (figs. 2, 4, 9). Statistical analyses (discussed below) verify that these differences are significant.

In this report, mean grain size is considered the best index property to sound velocity for continental-terrace sediments, and is about equal to percent clay size (fig. 10) as the best index in the two deep-sea environments. Horn *et al.* (1968) found that mean grain size was the best index to velocity in cores from the Norwegian and Mediterranean Seas, and Schreiber (1968) found median grain size and porosity of most importance in cores from the North Atlantic and Caribbean.

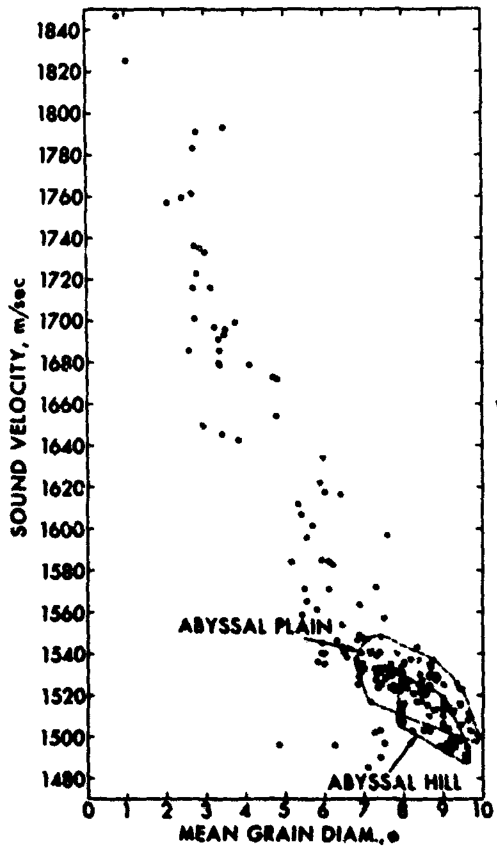

Figure 9. Mean diameter of mineral grains vs. sound velocity, all environments.

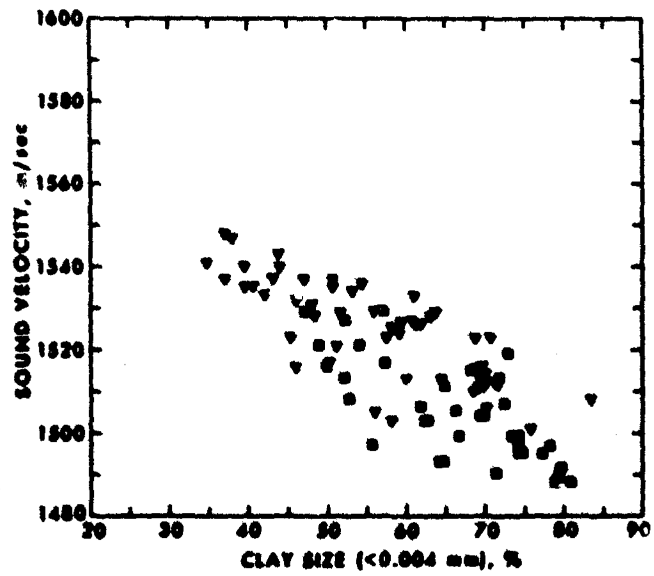

Figure 10. Percent clay size vs. sound velocity, abyssal hill and abyssal plain.

The usual laboratory size-analysis procedures divide a sample into three grain-size groups which are expressed as weight percentages of a unit volume: sand (2.0 to 0.0625 mm), silt (0.0625 to 0.004 mm), and clay (less than 0.004 mm).

Grain sizes affect velocity through their effects on porosity, density, and other factors. As index properties to sound velocity, percentages of sand, silt, and clay are as important as mean grain size, and better than porosity and density in all of the environments. In continental-terrace sediments, percent sand (fig. 11) and mean grain size (fig. 9) are of about equal value in deriving velocity. In abyssal-hill and abyssal-plain sediments, percent clay (or its complement, percent silt plus percent sand) may be slightly better than mean grain size in deriving velocity (fig. 10). Sorting has no usable relationship with velocity in these sediments.

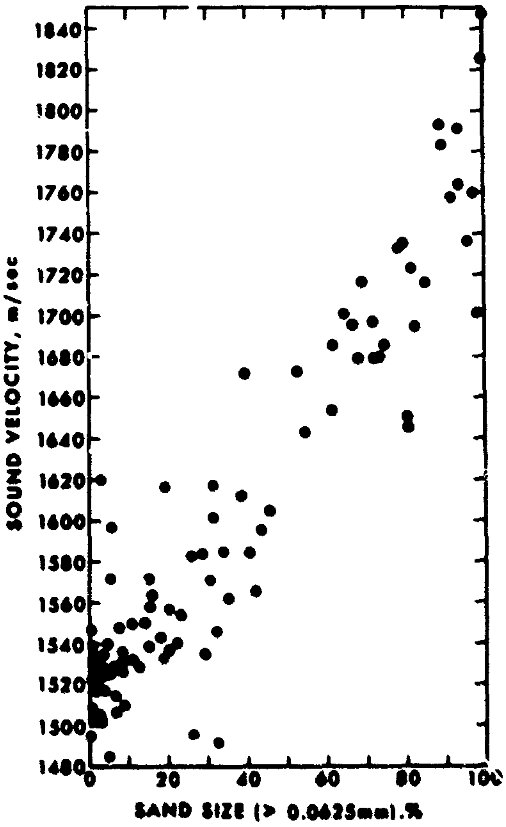

Figure 11. Percent sand size vs. sound velocity, continental terrace.

Impedance (Density x Velocity)

The characteristic impedance of a medium is the product of density, ρ, and velocity V_p (impedance = ρV_p, g/cm^2 sec); it is an important property of any material. The amount of energy reflected (or lost) when sound passes from one medium into another of greater impedance is largely determined by impedance differences (or "mismatches"). When the impedances of any two media are the same, sound travels through their boundary without reflection; this phenomenon led to the development of "rho-C rubber" which has about the same characteristic impedance as that of seawater and has been used to coat and protect underwater sound transducers without energy losses (Kinsler and Frey, 1950). In the field of marine geophysics, echo-sounding and continuous-reflection-profiling records indicate the travel-time of sound between impedance mismatches at the particular power and frequencies involved in the sound source and in amplifying and filter systems. Most surficial sea-floor sediments have sound velocities less than that in the overlying bottom water, but the echo-sounder records strong reflections in these areas because sediment densities are so much greater than water densities that a sufficient impedance mismatch is created.

The use of impedances in computation of reflection coefficients and bottom losses is discussed in a section below. Impedances were computed for the sediments of this study, using measured values of density and velocity (tables 1, 2).

Impedance increases with decreasing porosity in an almost linear relationship in the two deep-water environments and for the higher-porosity sediments of the continental terrace (figs. 12, 13). Density increases with impedance in a gently arcuate trend for terrace sediments, but is virtually linear for the abyssal-hill and abyssal-plain sediments (figs. 14, 15). Either density or porosity is more accurate than mean grain size in determining impedances.

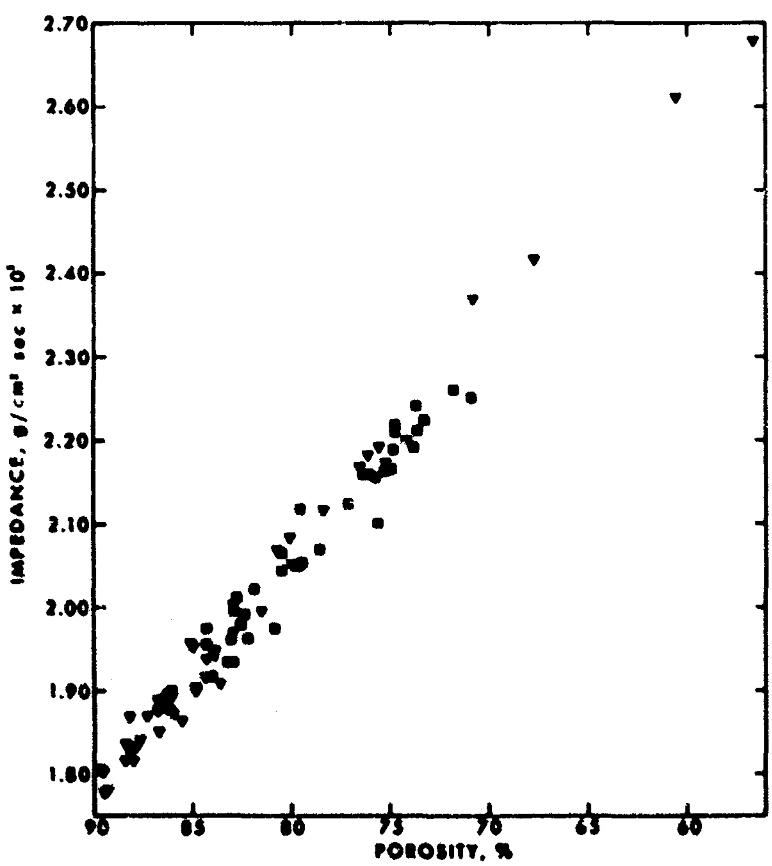

Figure 12. Porosity vs. impedance, abyssal hill and abyssal plain.

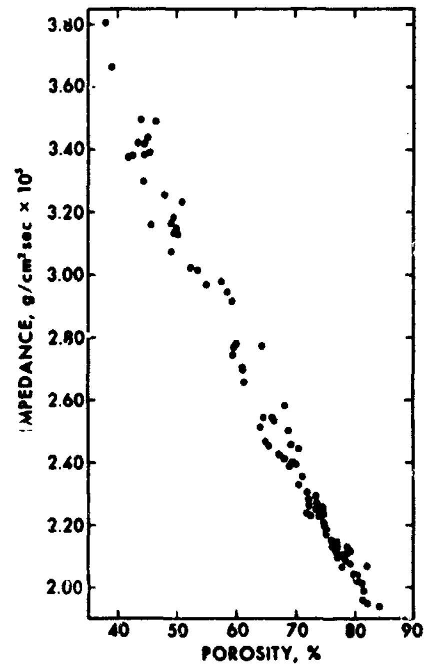

Figure 13. Porosity vs. impedance, continental terrace.

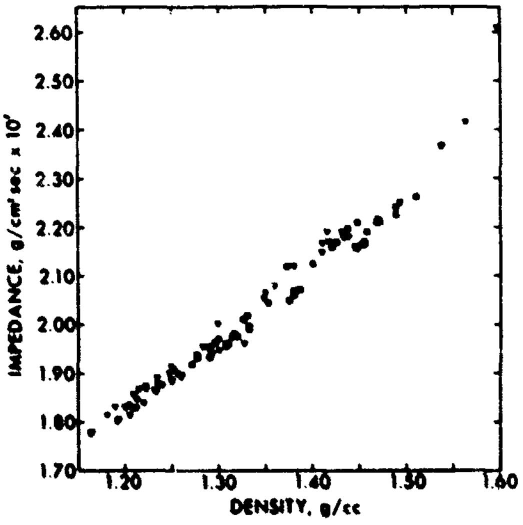

Figure 14. Density vs. impedance, abyssal hill and abyssal plain.

The excellent relationships between impedance and other properties are useful in predicting impedance for use in reflection studies, and to determine values of sound velocity. In the continental-terrace sediments (figs. 13, 15), density is slightly better than porosity as an index to impedance; in the two deep-water environments, density is slightly better than porosity (figs. 12, 14). Given porosity or density, the appropriate diagram can be entered to obtain impedance, which, when divided by the appropriate density, yields a value of sound velocity.

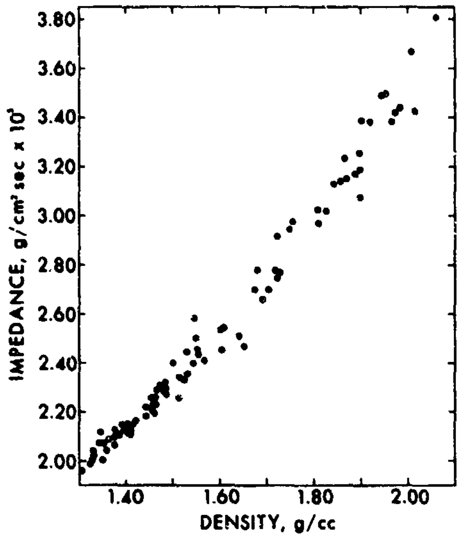

Figure 15. Density vs. impedance, continental terrace.

Density x (Velocity)^2

The product of density and the square of velocity is particularly significant in the relationshps between constants in either elastic or viscoelastic media. It is linked to other constants by $\rho V_p^2 = \kappa + 4/3\mu = \lambda + 2\mu$ (where κ = the bulk modulus, μ = the rigidity modulus, and λ = Lame's constant). Discussions of this topic will be deferred to Part II (Elasticity), but the empirical relationships between ρV_p^2 and other properties are of interest here. Using measured values of density and velocity, ρV_p^2 was computed for the sediments of the present study (tables 1, 2).

The best index property to ρV_p^2 is porosity in all three environments (figs. 16, 17); density is also good (figs. 18, 19). Mean grain size has the third best empirical relationship to ρV_p^2, but the scatter around the regression lines is about two to four times greater than for porosity or density. In the abyssal-hill, clay environment, both porosity and density have, practically, a linear relationship with ρV_p^2; in the abyssal-plain (turbidite) environment the trend with density is gently arcuate.

As to be expected, considering the relationships between porosity and density (and both with sound velocity), ρV_p^2 increases with decreasing porosity and increasing density. Either porosity or density can be used to enter the diagrams (or equations) to get ρV_p^2; dividing the result by the appropriate density results in a value for V_p^2; the square root yields a value for sound velocity.

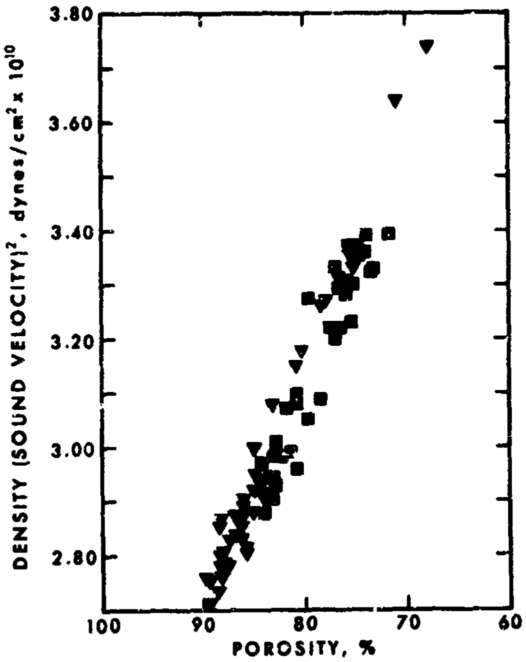

Figure 16. Porosity vs. density X (velocity)^2, abyssal hill and abyssal plain.

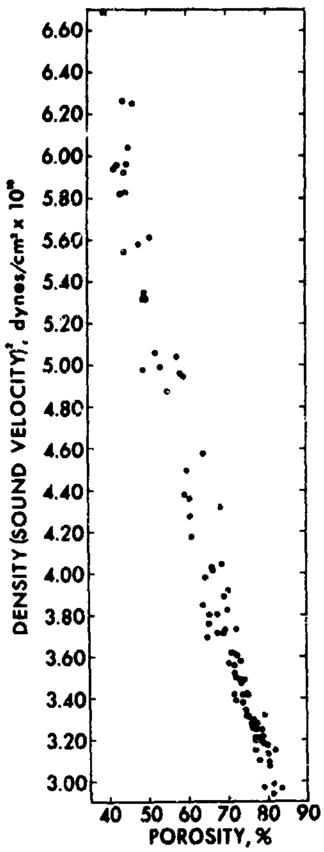

Figure 17. Porosity vs. density X (velocity)^2, continental terrace.

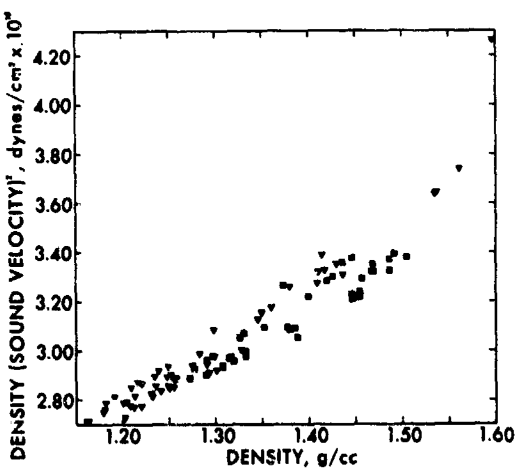

Figure 18. Density vs. density X (velocity)^2, abyssal hill and abyssal plain.

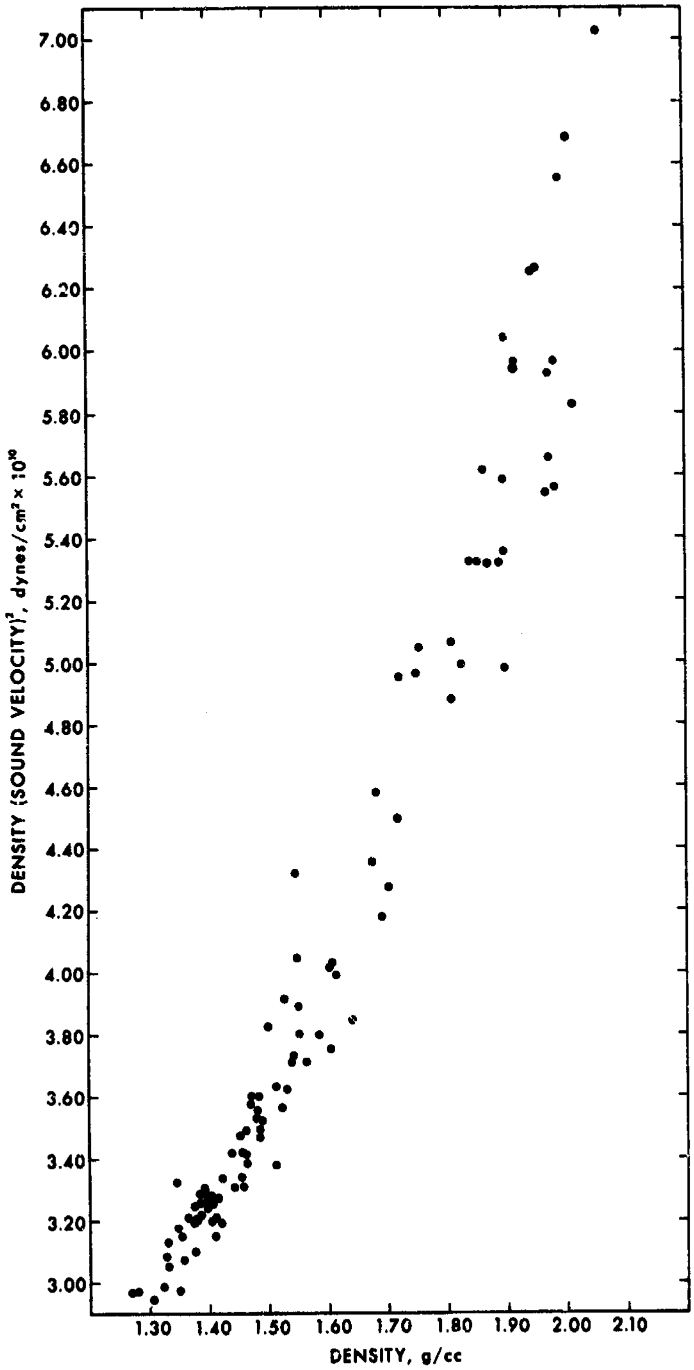

Figure 19. Density vs. density X (velocity)^2, continental terrace.

Ratio Of Velocity In Sediment To Velocity In Water

In the fields of marine geophysics and underwater acoustics, the ratio of sound velocity in surface sediment to sound velocity in bottom water is important. Whether this ratio is less than or greater than 1 is important in studying the reflection and refraction of sound waves incident on the sea floor. It is also important in predicting the *in situ* velocity of sound in sediments at the water-sediment interface (see Part III, Prediction) because this ratio is the same in the laboratory as it is *in situ*. Thus, all that is needed to determine *in situ* sediment velocity is the ratio and bottom-water velocity.

Scatter diagrams can be used to determine the ratio, but the best way to obtain this useful property is to determine the sediment sound velocity in the laboratory at a known temperature by measurement or entry into diagrams or tables (as in the present report), and divide by the appropriate speed for seawater. The error due to salinity is so small that one could use the speed for 34.5 ppt (1529.4 m/sec at 23°C, 1 atmosphere) with negligible error. Average values for this ratio are listed in tables 1 and 2.

Reflection Coefficient And Bottom Loss At Normal Incidence

GENERAL

The subject of reflection, refraction, and energy losses of sound incident on the sea floor is too complex for simple statements, and is not within the scope of this report. There is voluminous literature on this and closely-related subjects; for basic discussions, the reader is referred to textbooks by Ewing *et al.* (1957), Officer (1958), Kinsler and Frey (2nd ed., 1962), White (1965), and Tolstoy and Clay (1966); acoustic models and equations, and experimental work at sea, are discussed in recent papers by Bucker (1964), Bucker *et al.* (1965), Barnard *et al.* (1964), Marsh *et al.* (1965), and Cole (1965); a recent, annotated bibliography is useful (Frey, 1967).

The real sea floor cannot be included in any single *geoacoustic model* (a "model" of the real sea floor with emphasis on measured or extrapolated values for those properties of importance in acoustic problems). In shallow water, the two most common geoacoustic models are (1) low-impedance silt-clay over higher impedance sand-silt, and (2) a single layer of high-impedance sand. In the deep sea, the two most common geoacoustic models are (1) a fairly homogeneous, thick, clay layer over rock (with or without volcanic ash layers), and (2) the thick turbidite sections of abyssal plains in which there are multiple, alternating layers of low-impedance mud, and higher-impedance sand-silt over rock. As discussed in the various sections of this report (with appropriate references) and by Hamilton *et al.* (1969), the sediments of the real sea floor have, in general, the following properties (at frequencies of interest in underwater acoustics): (1) sound velocities range from about 3 percent below to about 22 percent above the velocity in the bottom water, (2) there is no dependence of velocity on frequency, (3) the sediment body, or layers, absorb sound (attenuation has, probably, a linear dependence on frequency), (4) density, velocity, and other elastic-property gradients are present, and (5) almost all open-ocean sediments have a finite rigidity and transmit shear waves. Any rigorous acoustic model must include these properties of the real sea floor and other properties, such as roughness and slope.

A viscoelastic solid model which includes many of the above properties has been successfully used to predict reflectivity of sound incident on the sea floor (Bucker, 1964;

Bucker *et al.*, 1965). However, much simpler models of a fluid over a fluid (with or without absorbing layers) have been successfully used in reconciling reflectivity theory with experimental data (*e.g.*, Cole, 1965). These simpler models are, apparently, successful because rigidity and shear-wave velocities are low in most sediments.

However, as Morris (1967) has shown, the introduction of shear waves into a rigorous model results in higher theoretical bottom losses (energy is lost through the conversion of compressional to shear waves at layer boundaries). This whole matter needs continuing study and refinement, which are now possible with the aid of computers and new information on the acoustic properties and layering of the sea floor.

In the following section, Rayleigh reflection coefficients and bottom losses are computed by using the measured sediment densities and velocities of this report, and equations of the fluid/fluid model. These computations are listed and discussed because they appear to be close to measured values and because they are useful in studies of reflection and refraction of sound incident on the sea floor. A case is not being made for the fluid/fluid model. The model preferred by the author is that of a viscoelastic solid in which complex Lame constants are independent of frequency, and in which there is provision for shear waves (Bucker *et al.*, 1965; Hamilton *et al.*, 1969).

THE SEA FLOOR AS A LIQUID MODEL

The simplest reflection model involves a simple, harmonic, plane wave incident on a plane boundary between two fluids across which there is a change in velocity and density. Several recent textbooks include the derivations of the appropriate equations for this model (*e.g.* Ewing *et al.*, 1957; Officer, 1958; Kinsler and Frey, 1962). The Rayleigh reflection coefficient for this model expresses the ratio of the amplitudes, or pressures, of a reflected wave to that of the incident wave; at normal incidence, the reflection coefficient, R, is expressed by

$$R = \frac{\rho_2 V_{p_2} - \rho_1 V_{p_1}}{\rho_2 V_{p_2} + \rho_1 V_{p_1}} \tag{8}$$

where

$\rho_1 V_{p_1}$ is the impedance of the first medium

$\rho_2 V_{p_2}$ is the impedance of the second medium

Bottom loss, BL, of a plane wave at normal incidence (on a peak-pressure basis), expressed in dB, is

$$BL = 20 \log R \tag{9}$$

Using the above equations, and measured densities and velocities (tables 1, 2), R and BL were computed for the sediments of the present study, and plotted against porosity and density (figs. 20-23). Seawater impedance was computed for 23°C, 1 atmosphere, and the appropriate salinity.

Density is slightly better than porosity as an index to both reflection coefficients and bottom losses. Both density and porosity are better indices of R and BL than any of the grain-size parameters.

Faas (1969) made a statistical study of reflection coefficients, R, computed with equation 8 and literature values of density, porosity, and velocity reported by Hamilton *et al.*(1956), Sutton *et al.* (1957), Shumway (1960), and Morgan (1964); most of these sediments were from shallow water, except those of Sutton (which were not all from common deep-sea environments). Faas's linear equation relating R to porosity, n, R = 0.6468 - 0.6456 (n), is close to that of this report for the shallow-water sediments.

The reflection coefficients and bottom losses of tables 1 and 2 were computed with values of sediment and water densities and velocities at 23°C, and 1 atmosphere. Computations using *in situ* values indicate the laboratory values of bottom losses are within a few tenths of a dB of the *in situ* losses. For example, *in situ* bottom loss for abyssal-hill silty clay, at a water depth of 5000 m, is only 0.3 dB greater than the laboratory value. Consequently, in the following comparison of computations vs. measured losses (Breslau, 1967), no attempt was made to compute *in situ* losses.

Comparisons between the computations of bottom losses at normal incidence, BL (tables 1, 2: figs. 22, 23), and normal-incidence measurements of bottom loss at sea (Breslau, 1965, 1967), and measurements at sea of angles of intromission and critical angles of sound incident on the sea floor (Fry and Raitt, 1961), lead to the conclusion that the fluid/fluid model (without attenuation) is a close approximation for some studies of reflectivity (given certain layer thicknesses and properties) and energy levels of incident sound. Implicit in this comparison is that at normal incidence there is little or no dependence of reflection coefficients and bottom loss on frequency. This appears to be true when any second layer of the subbottom, for various reasons, cannot reflect sound which interferes with that reflected from the water-sediment interface (see Cole, 1965, for discussion). Some experiments in the laboratory which meet one of Cole's conditions (a highly attenuating first layer) support this conclusion: Nolle *et al.* (1963, p. 1398) observed practically no variation of the reflectivity of the sand-water surface in model studies involving frequencies of 0.5 and 1.0 MHz. Experiments at sea at seismic and higher frequencies (discussed below) lead to the same conclusion.

Breslau (1965, 1967) reported the results of the reflection measurements of 12-kHz sound at normal incidence on the shallow, and the deep, Atlantic sea floor. He demonstrated that measured sound-pressure and energy losses could be theoretically explained by the fluid/fluid model and by equations 8 and 9 used with measured or assumed values for the mass physical properties of sediments. Although he did not measure sound velocity in his sediment samples, his assumed values are realistic. His measurements verify the validity of computations of bottom loss, BL, from actual measurements of density and velocity (as in this report). The relationships between porosity and bottom loss which he established (Breslau, 1965, fig. 6) verify a similar plot by Hamilton *et al.* (1956, fig. 12). Breslau found that bottom loss, BL, had a slightly better correlation with percent silt and clay than with porosity but noted -- probably correctly -- that his problems in sediment sampling may have caused this, and that porosity should be a better index property to bottom loss (fig. 22). The average, computed bottom losses and porosities for the various sediment types of the present

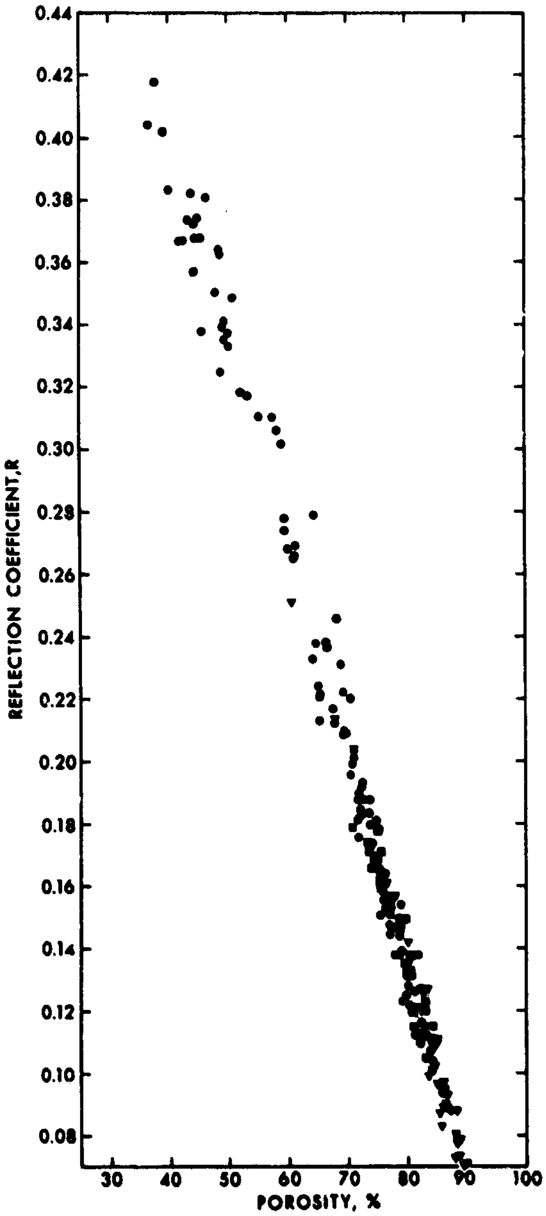

Figure 20. Porosity vs. Rayleigh reflection coefficient (R) at normal incidence, all environments.

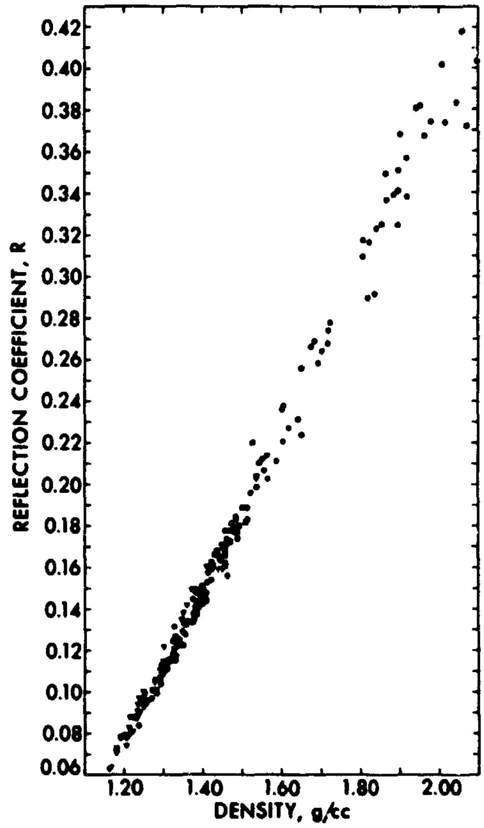

Figure 21. Density vs. Rayleigh reflection coefficient (R) at normal incidence, all environments.

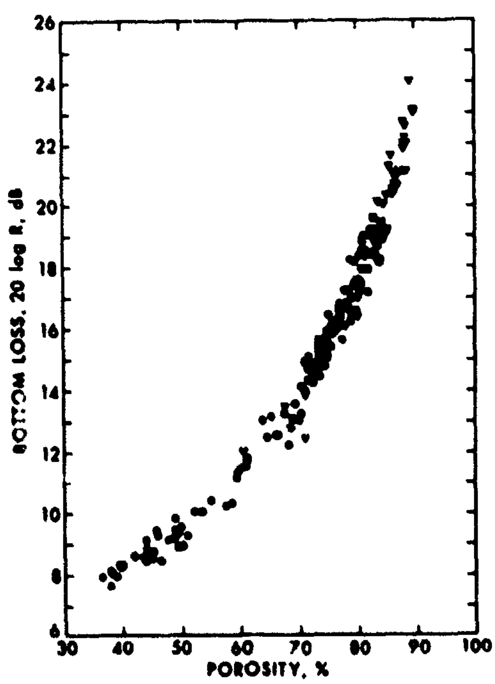

Figure 22. Porosity vs. bottom loss at normal incidence, 20 log R, all environments.

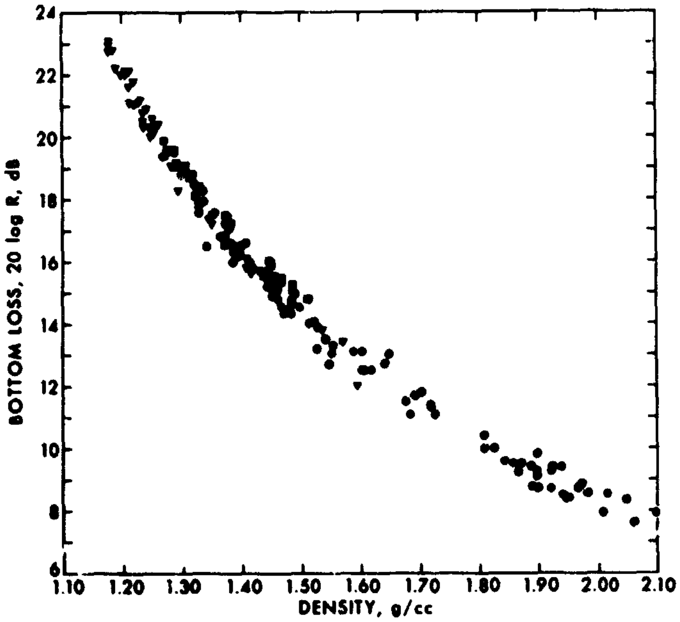

Figure 23. Density vs. bottom loss at normal incidence, 20 log R, all environments.

report (tables 1, 2) fall well within Breslau's measured values for shallow water (fig. 24). The inset in figure 24 shows Breslau's (1967, fig. 47) measured bottom losses (peak-pressure basis) in pelagic sediments northwest of Bermuda, and the close agreement with the average bottom loss in pelagic, abyssal-hill sediments of the North Pacific as computed with data from this report (table 2).

In the fluid/fluid model (as applied to the sea floor), when sound velocity in the bottom water, V_1, is greater than in the surficial sediment, V_2, and the angle of incidence varies from 90°, there is a decrease in reflection coefficients, and an angle of incidence is reached at which intromission occurs (all energy enters the lower medium and none is reflected). At the angle of intromission, θ_I, there is a 180° phase change in the reflected wave; expressed in the form used by Kinsler and Frey (1962, p. 145),

$$\cot^2 \theta_I = \frac{(V_1/V_2)^2 - 1}{[(\rho_2/\rho_1)^2 - (V_1/V_2)^2]} \tag{10}$$

Fry and Raitt (1961) examined long-range reflection records from the Pacific, and noted that as the range increased between sending and receiving ships, and the angle of incidence of sound reflected from the sea floor decreased from 90°, a point was reached over many areas of the sea floor where a phase change of 180° took place in the reflected wave. The decrease in reflection coefficients, and the phase change, indicated an insonified area in which the bottom-water velocity was greater than the sediment-surface velocity. The angle of incidence was then computed from the geometry of the experiment and water-velocity data. The ratio, V_1/V_2 (the reciprocal of the ratio of this report, V_2/V_1), and *in situ* sediment velocities were then computed with an equation of the fluid/fluid model (a form of equation 10) and with assumptions for the densities of the bottom water and sediment.

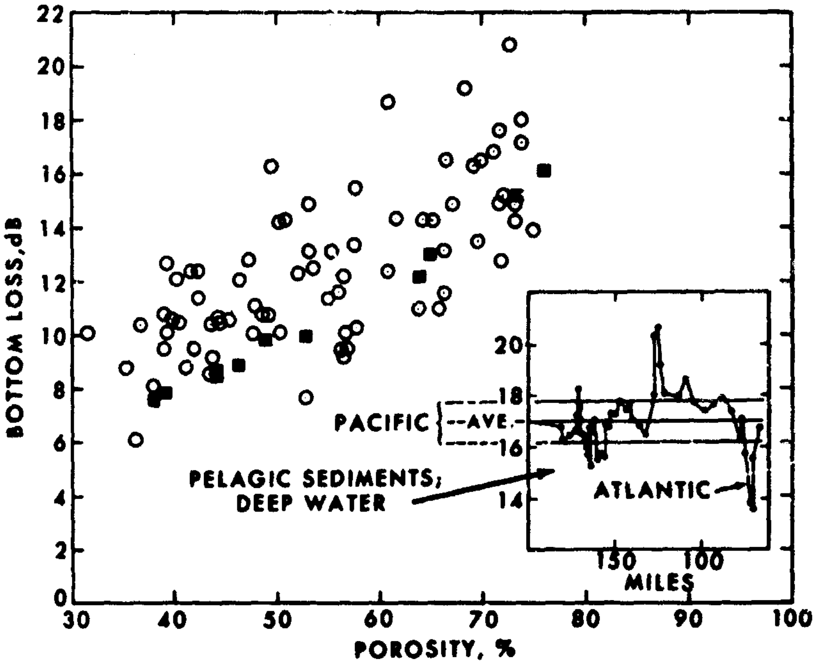

Figure 24. Porosity vs. bottom loss, dB (20 log R), at normal incidence. Upper left: open circles, measurements at 12 kHz, shallow water (Breslau, 1967) compared with average, computed values (solid squares) from this report. Lower right: deep-sea measurements in pelagic sediments northwest of Bermuda (Breslau, 1967) compared with computed mean and 3 standard errors, abyssal-hill silty clay, Pacific (from this report).

Computation of an average angle of intromission, 78.3°, for abyssal-hill pelagic sediments, using data from this report (corrected to *in situ* conditions), compares favorably with an average of those measured in similar sediments by Fry and Raitt, 75.8° (1961, table 1, Stas. F1-4, MK1 and 2, A7, DW 29, M5).

The average of velocity ratios in abyssal-hill sediments reported by Fry and Raitt, V_1/V_2 = 1.026 (V_2/V_1 = 0.977), was within 1 percent of the average for the sediments reported here (0.984, table 2), and within the maximum and minimum ratios (1.006 to 0.973). The density of bottom water used by Fry and Raitt in their computations (1.0277 g/cc, p. 592) was apparently the density of water of given salinity and temperature at surface pressure ("Sigma-T" of oceanographic tables), and not that of seawater at the indicated depths (about 1.0475 g/cc); their value for sediment density, 1.40 g/cc, was excellent. The use of 1.0277 g/cc for water density makes a small difference of about 0.002 less than they reported for V_1/V_2. For example, the average of all values in their table 1 is 1.024 (where water velocity was greater than sediment velocity); using a water density of 1.0475 g/cc, the average ratio is 1.022 (or 0.978, the reciprocal).

Sound Velocity vs. Shear Strength

GENERAL

Sound velocity in all materials increases with increasing resistance to shearing stresses (*i.e.*, dynamic rigidity of elasticity). For example, after lithification of a mud to a mudstone or shale, there is a marked increase in sound velocity in the material. This fact has led investigators of sound velocity in soft sediments (including the author) to consider the probability that there is a useful relationship between sound velocity and shear strength (cohesion) as measured in standard, soil-mechanics testing procedures. However, plots of sound velocity vs. shear strength (cohesion) in modern sea-floor sediments show that there is no relationship which would allow the use of cohesion as an index property for sound velocity (and *vice versa*). Recent studies by Horn *et al.* (1968) and Schreiber (1968a, b), as well as this study (fig. 25), indicate that over a considerable range of sediment shear strengths (cohesion) there are no measurable fluctuations in sound velocity which can be ascribed to shear strength. The reasons for this apparent anomaly are probably in the various methods of static tests of cohesion as compared with dynamic tests of rigidity.

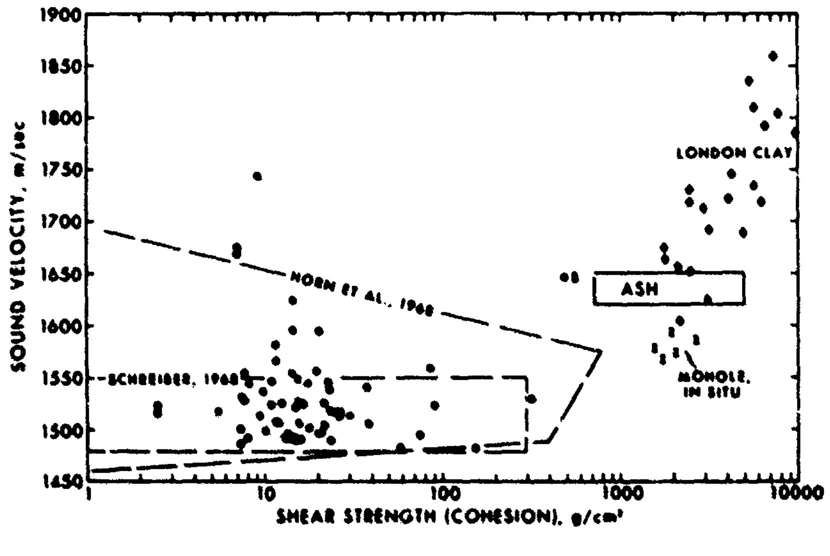

Figure 25. Sediment shear strength (cohesion) measured in soil-mechanics tests vs. sound velocity. Round dots: samples from North Pacific; x's: Mohole *in situ* (Hamilton, 1965; Moore, 1964); diamonds: London Clay (Ward *et al.*, 195[illegible]); dot labelled "B": Birch Bay Clay (Wilson and Dietrich, 1960); area labelled "ASH", and larger dashed area (Horn *et al.*, 1968); dashed rectangle: pelagic clays off Hawaii (Schreiber, 1968b).

The subject of shear strength in soils (sediments) forms a major part of any text in soil mechanics, so it will not be elaborated here. Aspects of shear strength important to dynamic rigidity will be discussed in Part II (Elasticity). In general, there are two components of shear strength, cohesion and friction. The shear strengths reported by Horn *et al.* (1968) and Schreiber (1968a, b) were measured by the fall-cone penetrometer; those of this report (fig. 25) were from vane-shear and triaxial-shear tests without normal stress; thus all are tests of cohesion only.

SHEAR STRENGTH (COHESION) AND DYNAMIC RIGIDITY

In testing clayey sediments in the laboratory at atmospheric pressure, it has been demonstrated by numerous investigators that the values of cohesion vary with the testing methods. Mitchell (1964) has a good summary of the factors involved in this phenomenon. In general, the tested cohesion at any given void ratio is a function of intrinsic cohesion (because of the factors noted above), and of the amount and rate of stress application. The rate at which stress is applied is particularly important because the measured "cohesion" increases with increasing stress rates.

Several decades of experience in foundation engineering have demonstrated that laboratory and *in situ* static tests of shear strength in clayey sediments can be used for most construction purposes where dynamic loads are not involved. In recent years there has been increasing interest in dynamic stresses and strengths in soils because of their importance in the design of foundations under airports, missile sites, radar towers, and similar installations; recent summaries and investigations, with many references to work in this field, are by Barkan (1962), Converse (1962), Richart and Whitman (1967), and Whitman and Richart (1967). In dynamic testing, vibrations are created in various ways; of interest in the present study are those created by transmitting compressional and shear waves through earth materials. In dynamic tests (in clayey sediments) it has been determined that repetitive stresses of low magnitude result in values of rigidity (shear modulus) well above the shear strengths (cohesion) measured in static tests; in fact, these values may be so far apart that cohesion from static tests cannot be safely used when dynamic shear is involved. In cohesionless sands, on the other hand, static and dynamic tests may be in reasonable agreement (Hardin and Richart, 1963; Whitman and Richart, 1967).

The subject of dynamic rigidity will be discussed in Part II, but it is instructive at this point to note some differences between static and dynamic tests in clayey sediments from recent local studies. In the San Diego Trough, *in situ* tests of compressional and shear-wave velocities in sea-floor sediments (Hamilton *et al.*, 1969) indicate values of dynamic rigidity of the order of 1.4 dynes/cm^2 $\times$ 10^8; *in situ* and laboratory static tests (Moore, in Buffington *et al.*, 1967) indicate that cohesion in these clayey silts is about 4 orders of magnitude less (1.6 dynes/cm^2 $\times$ 10^4).

In studies of sound velocity vs. shear strength (cohesion), the tacit assumption is that cohesion is a measure of dynamic rigidity, and that sound velocity increases with increasing rigidity (which is true). The lack of correlation between sound velocity and cohesion (fig. 25) is caused by the fact that static tests of cohesion cannot be compared to dynamic rigidity (as discussed above), and by the expectable, small effect of the rigidity (shear) modulus on the velocity of the compressional wave in present-day marine sediments (Part II, Elasticity).

In the San Diego Trough, if the value of cohesion (1.6 dynes/cm^2 $\times$ 10^4) is used in place of the known dynamic rigidity (1.4 dynes/cm^2 $\times$ 10^8), the increase in sound velocity exceeds that for no rigidity at all by less than 0.1 m/sec. Even the larger value of dynamic rigidity increases the velocity of the compressional wave only 5 m/sec.

To illustrate the effects of overburden pressures on both cohesion and sound velocity, data from deep-sea drilling (Preliminary Mohole, Guadalupe Site; Moore, 1964) and from London Clay (Ward *et al.*, 1959) are included in figure 25. The cohesion of the Mohole sediment samples was measured by Moore in triaxial tests in which the samples were placed under *in situ* pressures; *in situ* sound velocities were adjusted by Hamilton (1964). Additional information on London Clay is included in the next section. The value labelled "B" in figure 25 is for Birch Bay Clay (Wilson and Dietrich, 1960). The relatively high sound velocities in London Clay are due to several causes other than increased cohesion (or rigidity); for example, the porosity of this London Clay is about 38 percent.

The rectangular area labelled "ash" in figure 25 illustrates velocity-cohesion relationships measured by Horn *et al.* (1968) in volcanic ash found in cores from the Norwegian and Mediterranean Seas. The dashed lines include areas covering most of the data reported by Horn *et al.* (1968) and Schreiber (1968b; deep sea off Hawaii).

Anisotropic Velocity Relationships

Sediments or rocks which have velocities parallel to bedding planes which differ from those normal to bedding planes are termed anisotropic (or "transversely isotropic") in terms of velocity. This anisotropy parallel to bedding planes (which are usually parallel to the ground surface unless tectonically disturbed) is well known in geophysical prospecting in rocks. Velocities parallel to bedding planes in shale may be as much as 40 percent higher than velocities normal to bedding planes (LeRoy, 1950; Uhrig and Van Melle, 1955); an average figure for shale is about 10 percent. A question, then, arises concerning possible anisotropic velocity relationships in marine sediments.

When the flocculated, or "cardhouse," structure in clay (figure 1D) is placed under sufficient pressure, the sediment structure breaks down and the clay platelets assume a parallel, face-to-face, oriented structure. This orientation of clay-mineral particles under pressure causes anisotropic relationships in velocity and other properties (Ward *et al.*, 1959). The key words in the above statements are "sufficient pressure."

In laboratory compression and consolidation tests, reorientation of clay particles has been conclusively demonstrated (see recent resumé by Yong and Warkentin, 1966). In these tests, loads are applied rapidly (relative to the slow deposition of sediments forming overburden pressures in the sea floor), and when the sediment structure is broken down, there is little resistance to pressure-induced particle orientation and porosity reduction. Numerous laboratory tests have also indicated that slowly applied, very small loads result in little reduction of porosity at pressures much higher than those causing structural breakdown in more rapid loading. This has led soil-mechanics researchers to conclude that such tests are unrealistic when applied to slowly-deposited natural sediments which may compact under overburden pressures (Terzaghi, 1941; Leonards and Ramiah, 1960); Lambe, 1960; Bjerrum and Wu, 1960; Bjerrum and Lo, 1963; Crawford, 1964; Leonards and Altschaeffl, 1964).

Evidence on velocity anisotropy in marine sediments comes from both laboratory and field measurements.

From the laboratory, the following evidence can be presented:

1. From consolidation studies of marine sediments, Hamilton (1964) and Richards and Hamilton (1967) concluded (for most sediments) that there was little reduction of porosity with depth in the upper levels of the sea floor because of slow rates of deposition and sediment strength; specifically, no pressure-induced reduction of porosity in cores (0 to about 20 meters, and only about 5 percent (to 170 m) at the

Preliminary Mohole (Guadalupe Site). If there is little significant pressure-induced reduction in porosity, then there would be little particle orientation of the type necessary to induce anisotropic velocity relationships. There are, however, other causes of "velocity anisotropy" in some sediments because of alternating, thin layers of differing mineralogy or mass properties.

2. The numerous measurements of velocity in cores reported by Horn *et al.* (1968) and Schreiber (1968a, b) were all made through the core liner, parallel to the sediment surface; those made by the author were almost all normal to the sediment surface. The *in situ* measurements by divers and from submersibles were parallel to the sea floor. All of these measurements are in agreement when all variable factors of mass properties and environmental differences are taken into consideration.

3. Three sets of measurements in clayey silts off San Diego (one in a box core, and two in partially indurated mudstones sampled by divers) showed no velocity anisotropy.

4. In a laboratory consolidation-velocity tests, Laughton (1957) recorded a 13 percent increase of velocity normal to pressures of 256 kg/cm^2 (*i.e.*, parallel to the "sediment surface") in a deep-water silt-clay from the Atlantic. There are few sediment sections in the sea floor in which overburden pressures would be this high.

In the field, the following evidence on velocity anisotropy and particle orientation in sediments is known to the author:

1. Meade (1964) found no particle orientation in 600 meters of nonmarine sediments in the San Joaquin Valley, California.

2. Velocity measurements were made by the author in a laminated siltstone dredged from the north wall of the Puerto Rico Trench at an uncorrected depth of 6584 m (3600 fathoms) by Woods Hole personnel (Bowin *et al.*, 1966; Bunce and Hersey, 1966). This siltstone was probably from the lower part of the first layer of "unlithified sediment"; it was probably formed under an overburden of about 300 meters of sediment. Velocities parallel to the laminations were about 5 percent greater than normal to the laminations.

3. In Eocene time, a thick, marine clay layer (the London Clay) was deposited under the present site of London. Since the Eocene, about 400 meters of overburden has been removed by erosion (Ward *et al.*, 1959, 1965; Bishop *et al.*, 1965). The London Clay is still unlithified, but porosities have been reduced to about 38 percent. Ward *et al.* (1959) determined that particle reorientation had taken place, and that velocities parallel to ground surface were about an average of 5 percent greater than normal to the ground surface. Further tests (Ward *et al.*, 1965) indicated an increase in velocity from 4 percent at 50 feet (below ground level) to 8 percent at 138 feet.

In summary, the evidence indicates that there is no significant velocity anisotropy to corable depths in the sea floor and to 170 meters at the Preliminary Mohole (Guadalupe Site) which can be ascribed to pressure effects of overburden. The evidence, however, indicates that at sufficient depths (pressures) in the sea floor there should be a significant increase in velocity parallel to the sea floor because of pressure effects causing orientation of mineral particles. A 5 to 10 percent velocity increase parallel to the sea floor can probably be expected in a fairly homogeneous, silt-clay layer at depths on the order of 400 to 600 meters below the sea floor. Sediment thicknesses in Pacific abyssal-hill areas are usually less than these values; therefore, significant, pressure-induced velocity anisotropy in most of the Pacific may be rare. The Deep-Sea Drilling Project, now at sea, should furnish conclusive evidence in this matter.

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APPENDIX A: BEST EMPIRICAL RELATIONSHIPS, AND ENTRIES INTO SEDIMENT PROPERTY DATA

The interrelationships between mass physical properties of sediments in this report have been presented in three forms: (1) scatter diagrams illustrating the relationships between two given properties, (2) regression equations (Appendix B) for some of the illustrated data in the diagrams, and (3) tables which list the arithmetic mean (or average) and standard errors (for most sediment types) for the properties of each sediment type within the three large environments. With this much data and with several ways to derive unknown properties, it is advisable to discuss selection of the best entries into the data. The main utility of the present report is intended to be in derivations of density, porosity, sound velocity, impedance, and density X $(\text{velocity})^2$. The discussion in this appendix will be confined to the best ways to determine these properties for laboratory conditions (23°C, and 1 atmosphere pressure) with or without given data. The general subject of prediction of *in situ* properties will be reserved for Part III.

There is now enough information on sea-floor sediments from the various larger environments to allow, in the absence of data, direct prediction of sound velocity without going through the intermediate step of predicting porosity, or other properties, and then using these values as indices to velocity. When no sediment-property data are available, one should enter the appropriate table for the estimated environment and sediment type, and use the average value for the desired property. The listed standard error of the mean can be used to determine probable maximum and minimum values. The standard error, when multiplied by the appropriate value (below) indicates the confidence level; other values and confidence levels can be found in any statistics textbook (*e.g.*, Arkin and Colton, 1956, p. 113-127).

Number of Standard Errors	Confidence Level (%)*
1.0	68.3
2.0	95.5
3.0	99.7

*For 30 or more items in a sample. For smaller samples, a different "*t*" table must be used (Arkin and Colton, 1956, p. 127); for example, for 10 items in a sample, 3.17 standard errors are required for a confidence level of 99 percent.

Given mass properties of a sediment such as density, porosity, and size analyses, sound velocity and other unknown properties can be approximated by entry into the illustrated scatter diagrams, or by entry into the regression equations for the data shown in the diagrams (Appendix B). The following outline indicates best entries to get certain properties, and magnitude of "errors" which might be expected. These "errors" are the Standard Errors of Estimate of the regression equations (Appendix B).

A. To Get Sound Velocity, V_p

(Note: References are made to figures and equations in the main text of the report.)

1. Entry into diagrams: velocity vs. given property (figs. 5-11). For all three general environments, size-analysis values are slightly better indices than either density or porosity. Expectable errors in velocity for continental-shelf sediments, using mean grain

diameter are about 29 m/sec; using porosity or percent sand, about 31 m/sec; using density, about 33m/sec. In abyssal-plain (turbidite) sediments, use of grain-size parameters (percent clay, mean diameter, or percent silt) would result in errors of about 9 m/sec; using porosity and density, the errors would be about 11 and 12 m/sec, respectively.

2. Entry into diagrams: impedance (density X velocity) vs. given property (figs. 12-15).

   a. Method: divide the resulting impedance by the appropriate density to get velocity.

   b. In all three environments, density and porosity are better indices than size-analysis data. In continental-terrace sediments, porosity will yield an error in computed velocity of about 41 m/sec; density, about 38 m/sec; and mean grain diameter, about 100 m/sec. In both of the deep-water environments, use of density will result in a velocity error of about 14 m/sec; using porosity, about 18 m/sec. Use of mean grain diameter, in abyssal-hill sediments, yields velocity errors of about 55 m/sec; and in abyssal-plain (turbidite) sediments, about 86 m/sec.

3. Entry into diagrams: density X $(\text{velocity})^2$ vs. given property (figs. 16-19).

   a. Method: divide the resulting value by the appropriate density to get $(\text{velocity})^2$; take the square root to get velocity.

   b. In all three environments, porosity and density are distinctly better indices than size-analysis data. In both deep-water environments, porosity and density yield errors of about 12 to 14 m/sec in velocity. In abyssal-hill sediments, mean grain diameter yields errors of 26 m/sec; abyssal plain (turbidite), 48 m/sec. In continental-terrace sediments as a whole, there is wide scatter of data points, but values for clayey silt indicate errors of about 35 m/sec when porosity or density is used as an entry.

B. To Get Impedance (Density x Velocity) (figs. 12-15).

1. Enter the diagrams of impedance vs. density, porosity, and mean grain diameter, in that order; errors near the mean values of density and porosity are about 1 percent in the two deep-water environments, and 2 to 3 percent in continental-terrace sediments.

C. To Get Density x $(\text{Velocity})^2$ (figs. 16-19).

1. To get density X $(\text{velocity})^2$, enter the equations or diagrams with porosity, density, and mean grain diameter, in that order. Errors in density X $(\text{velocity})^2$ in the two deep-water environments will lie between 1-1/2 and 2 percent for both porosity and density, but will be between 3-1/2 and 5-1/2 percent for mean grain diameter. In continental-terrace sediments, entering with porosity or density yields errors of about 5 percent; with mean grain diameter, about 8 percent.

D. To Get Reflection Coefficients, R, And Bottom Loss, BL, (figs. 20-23).

1. In the two deep-water environments, enter both curves with density or porosity, in that order. In continental-terrace sediments, porosity is slightly better than

density as an entry. In any computations involving reflection coefficients and bottom losses where detailed accuracy is required, and density is given or can be computed from the available data, it will usually be better to use the density value, determine a value for velocity (as above) and compute the reflection coefficient and bottom loss, using equations 8 and 9.

E. To Get Density And Porosity (figs. 2, 3, 4).

1. Given either density or porosity, and desiring the other property, one may enter the density-vs.-porosity curve with an error of about 1 percent for the dependent variable. However, if values of grain density are available, it is better procedure to use tabular values for density of seawater (in pore spaces), and compute the missing property, using equation 7. If mean grain diameter is used to enter the diagrams or equations to get porosity or density, the following approximate errors would be expected: for porosity, 3 percent in abyssal-hill sediments, 4 percent in abyssal-plain sediments, and 5 percent in continental-terrace sediments; for density, errors of about 4 percent in abyssal-hill sediments, and 5 to 6 percent in continental-terrace and abyssal-plain (turbidite) sediments. In the two deep-water environments, especially, it is probably more accurate to disregard the size analyses data (if near the mean for the sediment type), and use the tabulated values of density and porosity, rather than entering the diagrams or equations.

F. Miscellaneous Notes.

1. In entering any of the diagrams or equations, it is important to use those for the particular environment in which data are desired. Both the environment and sediment type should be known or predicted before entering the tabulated data. This is especially true for the wide range of sediment types and properties of continental-terrace sediments.

2. When either density or porosity can be used as an index, and both are of equal accuracy, density should be preferred because of laboratory procedures used in obtaining these values (as discussed in the main text).

3. When size analyses, only, are available as indices, mean grain diameter and percent clay size (less than 4 microns) or its reciprocal, percent sand plus silt, should be used, in that order.

4. In the absence of data on sediment properties, the tabulated values for silty clay should be used in the abyssal-hill environment, and for surficial sediments of deep, abyssal plains, because silty clay is the most common sediment type in these areas. In areas of turbidites, silty clay or clayey silt usually alternates with layers of silty sand, sandy silt, silt, sand-silt-clay, or even fine sand. To predict these properties, the data for these sediment types from the continental-terrace tables can be used. For data on volcanic ash see Horn *et al.* (1968) and Part III (Prediction); this material often forms layers in abyssal-plain or abyssal-hill sediments adjacent to the volcanic islands.

5. Depending on the extent and reliability of available data, in many cases, instead of entering the diagrams and equations, better values of some properties in abyssal-hill sediments (and to a lesser extent in deep abyssal plains) can be obtained by using the available data for a general area to identify the sediment type and then using the tables for the appropriate environment and sediment type especially when the available data are near the mean for the corresponding tabulated properties.

APPENDIX B: EQUATIONS FOR REGRESSION LINES AND CURVES (ILLUSTRATED DATA)

Regression lines and curves were computed for those illustrated sets of (x, y) data which constitute the best indices (x) to obtain desired properties (y). Separate equations are listed, where appropriate, for each of the three general environments, as follows: continental terrace (shelf and slope), (T); abyssal hill (pelagic), (H); abyssal plain (turbidite), (P). The equations are keyed by figure numbers to the related scatter diagrams in the main text. The Standard Errors of Estimate, σ, opposite each equation, are applicable only near the mean of the (x, y) values, and accuracy of the (y) values, given (x), falls off away from this region (Griffiths, 1967, p. 448).

It is important that the regression equations be used only between the limiting values of the index property (x values), as noted below. These equations are strictly empirical and apply only to the (x, y) data points involved. There was no attempt, for example, to force the curves expressed by the equations to pass through velocity values of minerals at zero porosity, or the velocity value of seawater at 100 percent porosity.

The limiting values of (x), in the equations below, are:

1. Mean grain diameter, M_z, ϕ
   (T) 1 to 9 ϕ
   (H) and (P), 7 to 10 ϕ
2. Porosity, n, percent
   (T), 35 to 85 percent
   (H) and (P), 70 to 90 percent
3. Density, ρ, g/cc
   (T), 1.25 to 2.10 g/cc
   (H), 1.25 to 1.50 g/cc
   (P), 1.15 to 1.45 g/cc
4. Sand size grains, S, percent
   (T), 0 to 100 percent
5. Clay size grains, C, percent
   (H), 45 to 80 percent
   (P), 35 to 80 percent

Porosity, n (%) vs. Mean Grain Diameter, M_z (ϕ). Figure 2

(T) $n = 34.84 + 5.28\,(M_z)$ $\qquad \sigma = 5.8$

(H) $n = 56.31 + 2.52\,(M_z)$ $\qquad \sigma = 3.8$

(P) $n = 49.56 + 4.01\,(M_z)$ $\qquad \sigma = 3.2$

Density, ρ (g/cc) vs. Mean Grain Diameter, M_z (ϕ). Figure 4

(T) $\rho = 2.130 - 0.091\,(M_z)$ $\qquad \sigma = 0.10$

(H) $\rho = 1.705 - 0.036\,(M_z)$ $\qquad \sigma = 0.07$

(P) $\rho = 1.915 - 0.074\,(M_z)$ $\qquad \sigma = 0.08$

Sound Velocity, V_p (m/sec) vs. Porosity, n (%). Figures 5, 6.

(T) $V_p = 2475.5 - 21.764\,(n) + 0.123\,(n)^2$ $\qquad \sigma = 30.6$

(H) $V_p = 1509.3 - 0.043\,(n)$ $\sigma = 13.3$

(P) $V_p = 1602.5 - 0.937\,(n)$ $\sigma = 11.3$

Sound Velocity, V_p (m/sec) vs. Density, ρ (g/cc). Figures 7, 8

(T) $V_p = 2270.9 - 1194.4\,(\rho) + 474.6\,(\rho)^2$ $\sigma = 32.9$

(H) $V_p = 1527.8 - 15.7\,(\rho)$ $\sigma = 13.3$

(P) $V_p = 1474.4 + 38.6\,(\rho)$ $\sigma = 11.8$

Sound Velocity, V_p (m/sec) vs. Mean Grain Diameter, M_z (ϕ). Figure 9

(T) $V_p = 1936.2 - 87.33\,(M_z) + 4.45\,(M_z)^2$ $\sigma = 29.2$

(H) $V_p = 1596.4 - 10.3\,(M_z)$ $\sigma = 11.7$

(P) $V_p = 1616.3 - 10.8\,(M_z)$ $\sigma = 8.7$

Sound Velocity, V_p (m/sec) vs. Clay Size, C (%). Figure 10

(H) $V_p = 1570.3 - 0.98\,(C)$ $\sigma = 8.6$

(P) $V_p = 1568.5 - 0.79\,(C)$ $\sigma = 8.0$

Sound Velocity, V_p (m/sec) vs. Sand Size, S (%). Figure 11

(T) $V_p = 1513.7 + 2.54\,(S)$ $\sigma = 30.8$

Impedance, $\rho\,V_p$ (g/cm^{2}sec x 10^5) vs. Porosity, n (%). Figures 12, 13

(T) $\rho\,V_p = 5.8572 - 0.06408\,(n) + 0.00021\,(n)^2$ $\sigma = 0.0665$

(H) $\rho\,V_p = 4.1475 - 0.0262\,(n)$ $\sigma = 0.0261$

(P) $\rho\,V_p = 4.4431 - 0.0297\,(n)$ $\sigma = 0.0218$

Impedance, $\rho\,V_p$ (g/cm^{2}sec x 10^5) vs. Density, ρ (g/cc). Figures 14, 15

(T) $\rho\,V_p = 2.0960 - 1.5857\,(\rho) + 1.1572\,(\rho)^2$ $\sigma = 0.0621$

(H) $\rho\,V_p = 0.0321 + 1.4828\,(\rho)$ $\sigma = 0.0187$

(P) $\rho\,V_p = 1.5556\,(\rho) - 0.0414$ $\sigma = 0.0196$

Density X (Velocity)^2, $\rho V_p{}^2$ (dynes/cm^2 x 10^{10}) vs. Porosity, n (%). Figures 16, 17

(T) $\rho\,V_p = 13.0167 - 0.19858\,(n) + 0.00093\,(n)^2$ $\sigma = 0.1901$

(H) $\rho\,V_p = 6.2476 - 0.03945\,(n)$ $\sigma = 0.0548$

(P) $\rho\,V_p = 6.9496 - 0.04735\,(n)$ $\sigma = 0.0427$

Density × (Velocity)^2, ρV_p^2 (dynes/cm^2 x 10^{10}) vs. Density, ρ (g/cc). Figures 18, 19

(T) $\rho V_p^2 = 7.4685 - 8.7338\,(\rho) + 4.0934\,(\rho)^2$	$\sigma = 0.1971$
(H) $\rho V_p^2 = 0.1031 + 2.1938\,(\rho)$	$\sigma = 0.0565$
(P) $\rho V_p^2 = 2.6861 - 1.8302\,(\rho) + 1.5954\,(\rho)^2$	$\sigma = 0.0491$

Reflection Coefficient, R vs. Porosity, n (%). Figure 20

(T) $R = 0.6692 - 0.00666\,(n)$	$\sigma = 0.0131$
(H) $R = 0.6199 - 0.00607\,(n)$	$\sigma = 0.0061$
(P) $R = 0.6461 - 0.00646\,(n)$	$\sigma = 0.0257$

Reflection Coefficient, R vs. Density, ρ (g/cc). Figure 21

(T) $R = 0.3870\,(\rho) - 0.3864$	$\sigma = 0.0099$
(H) $R = 0.3435\,(\rho) - 0.3339$	$\sigma = 0.0045$
(P) $R = 0.3428\,(\rho) - 0.3358$	$\sigma = 0.0251$

Bottom Loss, BL (dB) vs. Porosity, n (%). Figure 22

(T) $BL = 14.2 - 0.33\,(n) + 0.0046\,(n)^2$	$\sigma = 0.5$
(H) $BL = 68.1 - 1.69\,(n) + 0.0132\,(n)^2$	$\sigma = 0.4$
(P) $BL = 106.2 - 2.78\,(n) + 0.0207\,(n)^2$	$\sigma = 3.8$

Bottom Loss, BL (dB) vs. Density, ρ (g/cc). Figure 23

(T) $BL = 70.7 - 57.03\,(\rho) + 12.95\,(\rho)^2$	$\sigma = 0.4$
(H) $BL = 127.4 - 137.60\,(\rho) + 41.76\,(\rho)^2$	$\sigma = 0.3$
(P) $BL = 118.6 - 123.2\,(\rho) + 35.72\,(\rho)^2$	$\sigma = 0.4$

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13. ABSTRACT

A study of marine sediments in three major environments of the North Pacific: the continental terrace (shelf and slope), abyssal plain (turbidites), and abyssal hill (pelagic). Laboratory values of the mass physical properties of the sediments (e.g., sound velocity, density, porosity, grain size, and others) are measured and computed and their empirical relationships and environmental differences are discussed.

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