2. *Elements of marine geology*

2.1 The topography of the seafloor

More than 70 % of the Earth is covered by water and most of the water is deep, i.e. approximately 4 km. An overview of the Earth's bathymetry is given in figure 1.

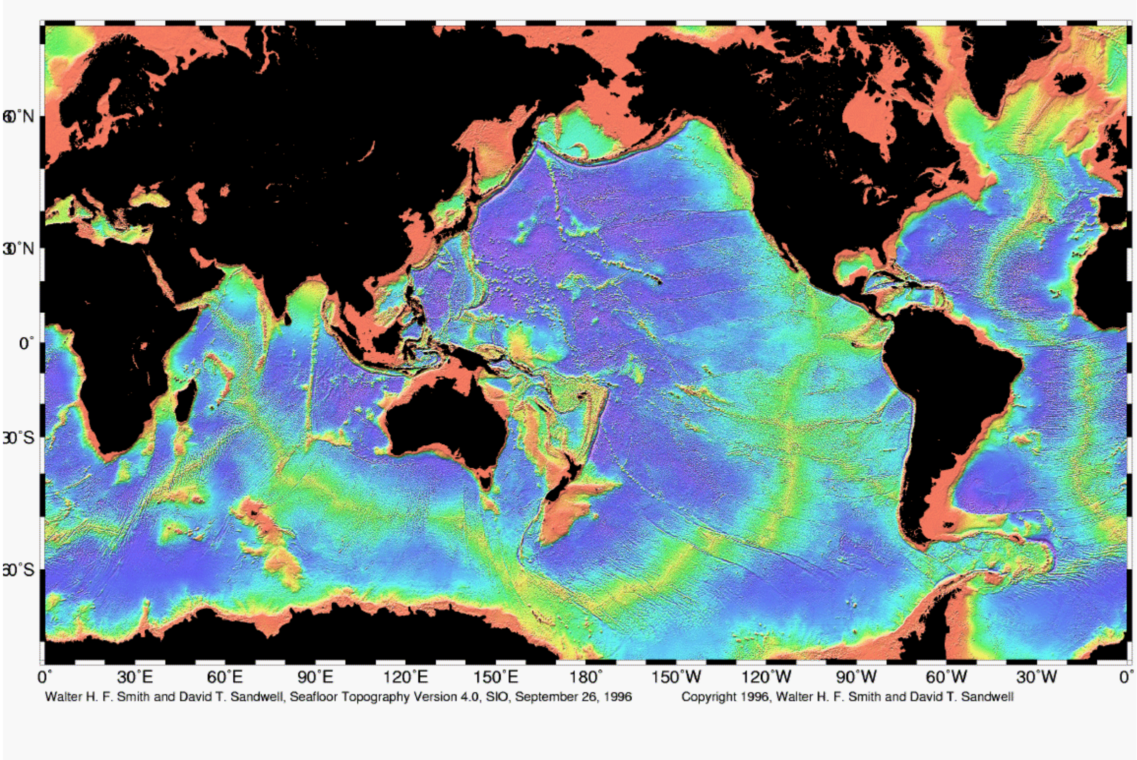

Figure 1

The topography of the seafloor exhibits a diversity of features not unlike those of the continental landmasses. The seabed is in general quite flat, even close to seamounts, ridges and the continental slope, with a slope seldomly exceeding 10°.
Figure 2 presents an artist's impression of common ocean basis features.

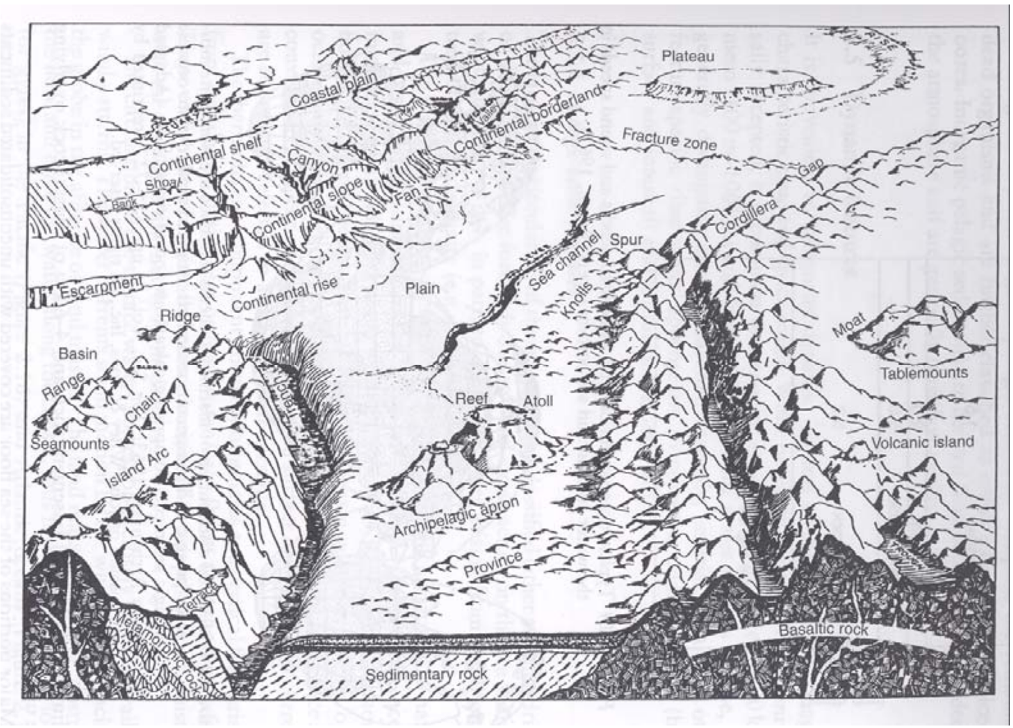

Figure 2

The features that are noted in figure 2 are defined in the glossary. Figure 3 below provides some further explanations.

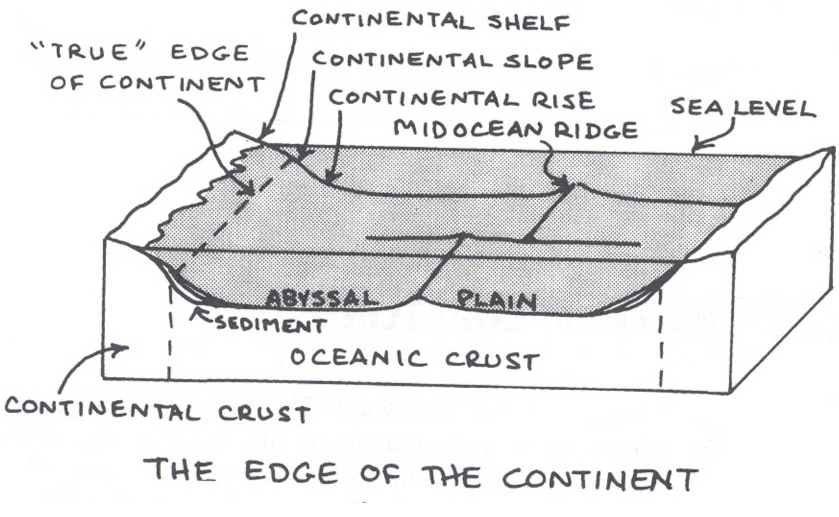

Figure 3

2.2 The interaction of sound with the seafloor

The seafloor is a reflecting and scattering boundary. It is often layered with a density and sound speed that may change gradually or abruptly with depth or even over short ranges. The seafloor is highly variable in its acoustic properties since its composition may vary from hard rock to soft mud. Because of the variable stratification of the

bottom sediments in many areas, sound is often transmitted into the bottom where it is refracted and internally reflected further down.
Further, the small-scale roughness of the seafloor causes scattering and attenuation of sound.
Thus the ocean bottom is a complicated propagating medium.

Marine seismic studies have greatly improved our understanding of the crustal structure of that part Earth covered by the ocean. A typical bottom-structure section is presented in figure 4. This section consists of 5 km of water, about 0.5 km of unconsolidated sediments, 1-2 km of basement rock and 4-6 km of crustal rock overlying the upper mantle. Also, typical sound speed values are indicated for each layer.

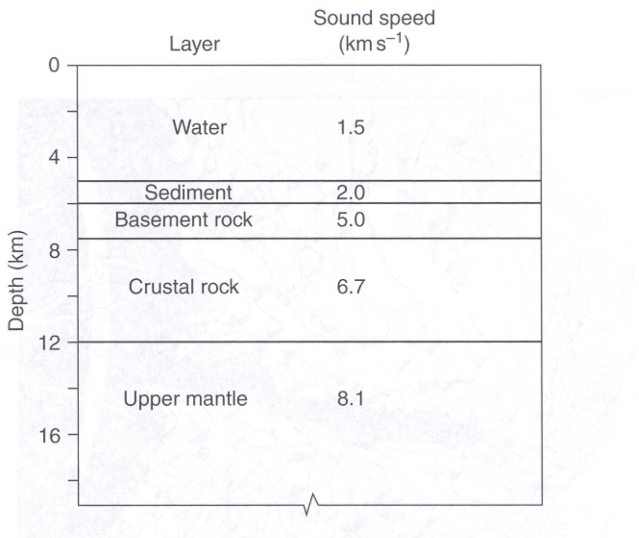

Figure 4

The crustal rock, or oceanic crust, consists largely of basalt, and is usually covered by sedimentary rock (denoted ‘basement rock’ in figure 4).
The continental crust largely consists of granite and is much thicker than the oceanic crust. It is covered by a wider range of sediments, but usually coarser than those in the deep ocean.

2.3 The seafloor sediments

Major portions of the seafloor are covered with unconsolidated sediments. In general, the structure of the ocean bottom consists of a thin stratification of sediments overlying the oceanic crust in the deep ocean and relatively thick stratification over continental crust on the continental shelves. Relatively recent sedimentations will be characterised by plane stratification parallel to the seabed, whereas older sediments and sediments close to the crustal plate boundaries may have undergone significant deformation, see figure 5 below.

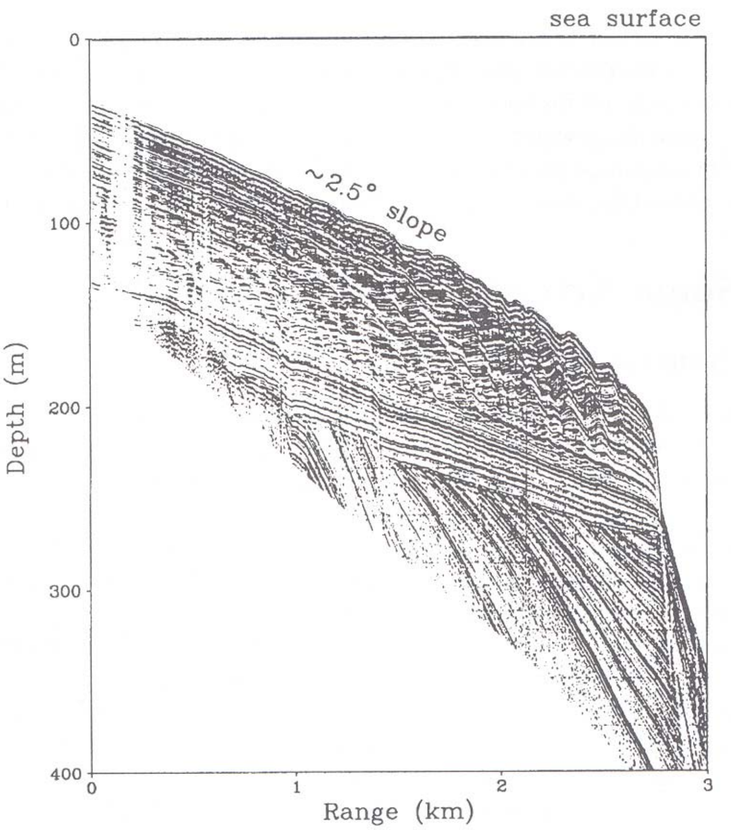

Figure 5

Sediments can be classified according to their origin as either ‘terrigeneous’ or ‘pelagic’:

- Terrigeneous sediments are derived from land and are particularly prominent near the mouths of large rivers. These sediments are generally classified as sand, silt and mud.
- Pelagic sediments are derived from either organic or inorganic sources. Organic pelagic sediments comprise the remains of dead organisms and are further classified as either ‘calcareous’ or ‘siliceous’ oozes. Inorganic pelagic sediments are derived from materials suspended in the atmosphere and are generally classified as clay.

2.4 Geo-acoustic modelling of the seafloor

Ocean bottom sediments are often modelled as fluids which means that they support only compressional waves. The rigidity, and hence the shear speed, of the sediment is usually considerably less than that of a solid, such as rock. The latter situation applies to the ocean basement or the situation where there is no sediment overlying the basement. Then the bottom must be modelled as ‘elastic’, i.e., it supports both compressional and shear waves. In reality, the bottom is ‘viscoelastic’, i.e., it is also lossy.

The information required for a complete geo-acoustical model of the sea bottom should include the following depth-dependent material properties:

- compressional wave speed

- shear wave speed
- compressional wave attenuation
- shear wave attenuation
- density

Sound waves are compressional waves. Compressional waves can pass through solids, liquids and gases. They consist of alternating pulses of compression and expansion acting in the direction in which the wave is travelling, see figure 6. Compressional waves are also denoted P (for ‘primary’) waves.

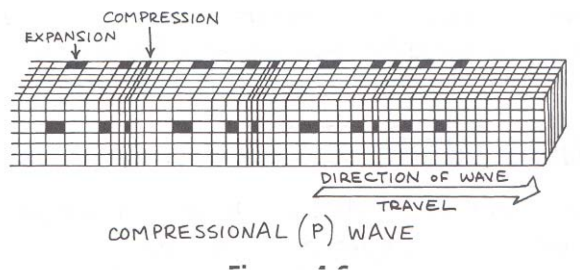

Figure 6

Shear waves travel through materials in alternating series of sidewise movements, see figure 7. Shear involves changing the shape of an object. Solids have elastic characteristics that provide a restoring force for recovery from shearing, but liquids and gases lack these elastic characteristics. Therefore, shear waves cannot be transmitted through liquids and gases. Shear waves are slower than compressional waves. Shear waves are also denoted S (for ‘secondary’) waves.

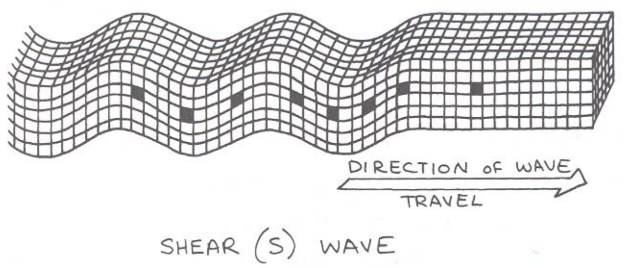

Figure 7

2.5 What the rocks and sediments are made of

Silicate minerals (i.e. minerals that contain the anion $(SiO_4)^{4-}$), a few oxide mineral (i.e. minerals that contain the anion O^{2-}) and calcium carbonate ($CaCO_3$, the mineral calcite) make up the bulk of the Earth’s crust – about 99 % by volume. These minerals are called the ‘rock-forming minerals’. Rock-forming minerals are everywhere: rocks, soils, sediments, but also roads and building are constructed of these rock-forming minerals.
Examples of silicates: Quartz (SiO_2), Feldspars (e.g. $KalSi_3O_8$)
Examples of oxides: Magnetite (Fe_3O_4), Hematite (Fe_2O_3)

A rock is a naturally formed, coherent 'aggregate' of minerals, which may be mixed with other solid materials such as organic matter. Thus rocks are collections of mineral grains stuck together. Rocks are grouped into three families, which are defined and distinguished from one another by their properties and by the processes that form them: 'igneous', 'metamorphic' and 'sedimentary rocks'.

Igneous rocks are formed by the cooling and consolidation of 'magma' (molten rock that is under the ground). If the magma reaches the surface while it is still in a molten state, it is called 'lava'. Examples of an igneous rock are 'granite' (consisting largely of feldspar and quartz) and 'basalt'.
Unconsolidated rock and mineral particles that are transported by water, wind, or ice and then deposited are called 'sediment'. A special kind of sediment is 'soil', which consists of loose particles that have been altered by biological processes to form a material that can support rooted plants. Sediments eventually become sedimentary rock, either by chemical precipitation from water at the Earth's surface or by the cementation of sediment.
The original form and mineralogy of metamorphic rock have been altered as a result of high temperature, high pressure, or both.

There are three groups of sediment: 'clastic sediment,' 'chemical sediment' and 'biogenic sediment'.
Clastic sediment is formed from loose, fragmental rock and mineral debris produced during weathering. Each individual particle in a clastic sediment is a 'clast' (being grains or rock fragments). Clast shapes vary from angular to rounded, and they range in size from the largest boulders down to the submicroscopic clay particles. Clast size is the primary basis for classifying both clastic sediments and clastic sedimentary rock, see figure 8.

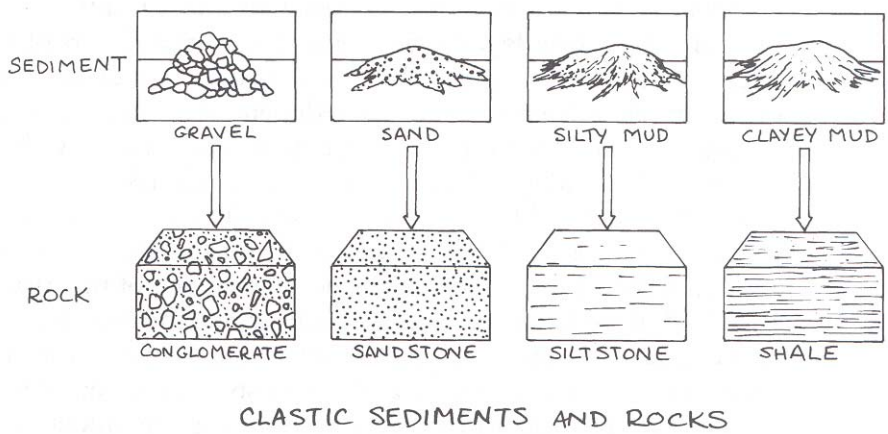

Figure 8

The four basic types of clastic sediment, based on the diameter of the particles they contain, are:

Gravel (pebbles, cobbles, boulders)	> 2 mm
Sand	0.06 mm – 2 mm
Silt	0.004 – 0.06 mm
Clay	< 0.004 mm

('mud' = silt + clay)

The range of clast sizes within a given sediment reflects a characteristic called ‘sorting’. A poorly sorted sediment has a wide range of clast sizes. Clast sorting and clast shape reflect the mechanisms of sediment transport and deposition. Mass-wasted sediment and ice-transported sediment tend to be poorly sorted, with angular clasts. (Poorly sorted sediments that are ice-transported are called ‘tills’). Water- and wind-transported sediments tend to be well sorted, with rounded clasts of uniform size.

No water on or in the Earth is completely pure and free from dissolved matter. When this dissolved matter is precipitated from seawater, chemical sediment is the result. This can happen through the biochemical reactions resulting from the activities from plants and animals in the water. For example, tiny plants living in seawater can decrease the acidity of the water, causing calcium carbonate to precipitate. Many ‘limestones’ form in this manner.
Chemical sediments can also be precipitated through inorganic reactions, such as the evaporation of seawater. Any salts that were dissolved in the seawater will be left behind as a residue from which minerals can precipitate.

Biogenic sediment is composed of the remains of organisms, i.e., the fossil remains of animals and plants. Solid fossils, such as shells and bones, often end up as broken fragments (clasts). A sediment largely composed of biogenic clasts is called a ‘bioclastic sediment’.
An important type of biogenic sediment is ‘ooze’, a fine muddy sediment that accumulates on the deep ocean floor. Ooze may be calcareous (carbonate-bearing remains of marine organisms such as corals) or siliceous (silica-rich remains of tiny floating marine organisms).

Clastic sedimentary rocks are classified on the basis of particle size, just as sediments are. The four basic classes are ‘conglomerate’,’sandstone’, ‘siltstone’ and ‘mudstone’ (or ‘shale’), i.e., the rock equivalents of gravel, sand, silt and clay, see figure 8. Most of the clasts in sandstones are quartz.
Chemical sedimentary rocks result from the lithification of chemical sediments. Most chemical sedimentary rocks contain only one important mineral. Chemical sedimentary rocks formed by evaporation of sea water are called ‘evaporites’, examples of which are rock salt (halite, NaCl) and gypsum ($CaSO_4.2H_2O$). Most of the evaporites are mined because they have industrial or human consumption uses.
Limestone is the most important of the biogenic sedimentary rocks. The majority of all limestones are bioclastic in origin consisting of the fossilised shells, or shell fragments, of marine organisms. These organisms built their shells of carbonate, so limestones are formed chiefly of the carbonate mineral calcite.

2.6 Geo-acoustic parameters of sediments and rocks

2.6.1 Sediments

Sound velocity and related properties of marine sediments

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New laboratory measurements of sediment properties in cores from the Bering Sea, North Sea, Mediterranean Sea, equatorial Pacific, and other areas, have been combined with older measurements and the results, with statistical analyses, are presented (for various sediment types in three general environments) in tables, diagrams, and regression equations. The measured properties are sound velocity, density, porosity, grain density, and grain size; computed properties are velocity ratios (sediment velocity/water velocity) and impedance. Mineral-grain microstructures of sediments are critical in determining density, porosity, and sound velocity; compressibility of pore water is the critical factor in determining sound velocity. New regression equations are provided for important empirical relationships between properties. Corrections of laboratory values to sea-floor values are discussed. It is concluded that sound velocity and density are about the same for a given sediment type in the same environment in any ocean if porosity is about the same. Given the mean size of mineral grains, or average porosity, of a sediment, the average sound velocity can be predicted within 1% or 2% in most environments. Comparisons with recent *in situ* measurements validate the laboratory measurements.

PACS numbers: 43.30.Bp, 43.30.Dr, 92.10.Vz, 43.35.Bf

INTRODUCTION

This report (the latest in a series) presents a summary and discussion of old and new measurements of sound velocity and related properties of marine sediments that have been made in the authors' laboratory. The measured properties are sound velocity, density, porosity, grain density, and grain sizes; computed properties are the velocity ratio and impedance. Included are measurements on sediments from the three general environments: continental terrace (shelf and slope), abyssal hill, and abyssal plain. The data included measurements on sediments from the main ocean basins, plus new measurements on samples from the Bering Sea, the North Sea, the Mediterranean Sea, and other areas.

In previous studies the only general sediment type not included was calcareous sediment (that containing more than 30% $CaCO_3$). To make this report more complete, and for the convenience of the reader, some of the results of a separate paper (Hamilton *et al.*[1]) on sound velocity and related properties of calcareous deep-sea sediments are included. These calcareous samples were taken in the equatorial Pacific during three Scripps Institution of Oceanography expeditions.

The purposes of this paper are to present averaged values and statistical analyses of laboratory-determined properties of the various sediment types in the three general environments, and to present and discuss important empirical relationships between properties; regression equations are provided for these relationships. Aside from use in some theoretical studies, the only utility in knowing the laboratory properties of sea-floor sediments is to be able to predict them for the *in situ* conditions. Predictions of the *in situ* values of various properties, have been the subject of previous reports (Hamilton[2,3]). In this report, procedures for correcting laboratory to *in situ* values will be noted in connection with some specific physical properties. In the last section values of sediment properties from the regression equations and tables are compared with recent *in situ* and other laboratory measurements.

I. METHODS AND RESULTS

A. Methods

Most of the samples of this report came from the upper 30 cm of the sea floor. This restriction was because of uncertainties concerning sediment disturbance in some gravity cores at deeper depths, and many samples came from this interval from subsamples from the Shipek grab-sampler, from box cores, and from sediments collected *in situ* by divers, and from submersibles. Exceptions to this restriction were made for those cores taken in soft, high-porosity sediments by the hydroplastic gravity corer (Richards and Keller[4]), and by piston corers when it could be shown that the tops of the cores were not missing (e.g., Mayer[5]). In this connection it is considered that the box corer takes the least disturbed samples, but only to shallow depths; the hydroplastic gravity corer also takes excellent samples as indicated by open worm burrows in sediment cores (e.g., Keller *et al.*[6]).

Compressional wave velocity (hereafter also called sound velocity or velocity) was measured in the laboratory by a pulse technique (operating at about 200 kHz); estimated margins of error were ±3 m/s in clays and ±5 m/s in sands. *In situ* measurements were made at 14, 7, and 3.5 kHz using probes inserted into the sea floor by divers or a submersible. All sound velocities were corrected to 23 C° and 1-atm pressure using tables for the speed of sound in sea water.[7] The velocity ratio (sediment velocity/sea water velocity) was determined by dividing the sediment velocity at 23 °C by the speed of sound in sea water at 23 °C, 1-atm pressure, and of the same salinity as the bottom water at the sampling site.

Saturated bulk densities were determined by the weight–volume method. The bulk densities of the mineral solids were determined by a pycnometer method. Porosities

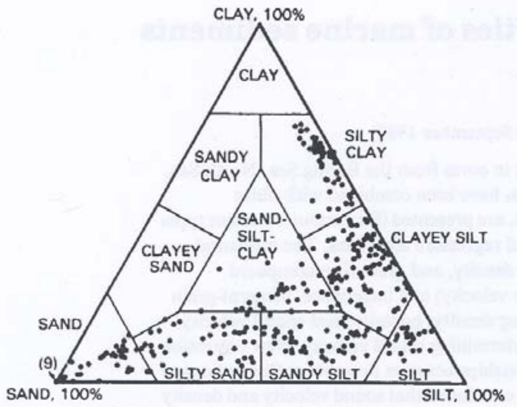

FIG. 1. Continental terrace (shelf and slope) samples plotted on the Shepard (1954) nomenclature diagram.

were determined after drying sediment samples in an oven at 110 °C ± 5 °C and corrections were made to allow for the amount of dried salts in the dried mineral residues.

Size analyses were made by wet sieving to separate the sand sizes (greater than 0.062 mm) which were further analyzed in the Emery settling tube. The pipette method was used for size analyses of the silt and clay fraction (less than 0.004 mm). The results of these analyses were plotted and the mean grain sizes were computed by averaging the 16th, 50th, and 84th percentiles (Folk and Ward[8]); when the 84th percentile would have involved undue extrapolation, the 25th, 50th, and 75th percentiles were used. In the scatter diagrams and tables the mean grain sizes are shown in the logarithmic phi scale ($\phi = -\log_2$ of grain diameter in mm); as emphasized by Pierce and Graus[9] a mean (arithmetic average) in the logarithmic phi scale is the geometric mean in the metric scale.

Sediment nomenclature followed that of Shepard[10] (Fig. 1), except that within the sand sizes, nomenclature followed the Wentworth scale.[11]

The data were examined statistically to determine: (1) the arithmetic mean (average), standard deviation, median, and standard error of the mean for the most important properties (density, porosity, sound velocity, and the velocity

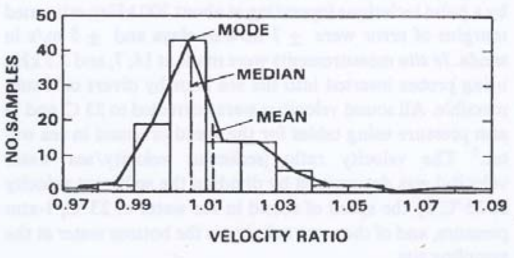

FIG. 2. Frequency distribution diagram for the velocity ratio (sediment velocity/sea water velocity) for clayey silt samples from the continental terrace (shelf and slope); see text for discussion.

ratio) for each sediment type within each environment, (2) frequency distribution diagrams were plotted for these data and the mean, median, and mode were indicated; see Fig. 2 for an example, and (3) regression equations and their standard errors of estimate were computed for most of the illustrated relationships between properties.

B. Results

The results of this study are reported in three forms: diagrams, regression equations, and tables. The diagrams serve to illustrate the ranges and scatter in selected properties, as well as the relationships between properties. Regression equations and their standard errors of estimate for most of the illustrated data are in the Appendix. Tables I and II list averages of properties of the terrigenous sediments (derived from land) of the continental terrace (shelf and slope). Tables III and IV list averages of properties of sediments of abyssal hills (pelagic clay), and sediments of abyssal plains (terrigenous and siliceous). Tables V and VI list properties of calcareous sediments from the equatorial Pacific. In these tables, the number of samples in each sediment type and the standard error of the mean are listed for density, porosity, sound, velocity, and the velocity ratio.

All of the data (in diagrams, regression equations, and tables) are for laboratory conditions at 1-atm pressure and 23 °C. For studies of sound propagation, some properties must be corrected to *in situ* values in the sea floor (as discussed in appropriate sections).

The last publications of the tables (Hamilton[12]) included measurements through 1979; the last set of illustrations and regression equations were published in 1974.[3]

II. DISCUSSIONS AND CONCLUSIONS

A. Introduction

Enough information is now at hand to be able to predict, with some confidence, certain properties of the sediment surface in the general environments. If sediment type and environment (as from a sediment chart) are the only information available, then data from the tables can be used in predictions. The probable errors in a property to be expected when predicting mean (average) values in the sediment are indicated by the standard error of the mean (SE) listed with each important property. The overall dispersion can be determined by computing the standard deviation (SD):SD = (SE)(no. samples)$^{1/2}$. When sediment information such as mean grain size, density, or porosity are available, then one can enter the regression equations for the particular environment; in which case the accuracy of the determination of Y (given X) is indicated in each case by the standard error of estimate listed with each equation. This standard error may be used in the same manner as the standard deviation.[13] The expectable scatter and range of properties can be seen in the figures.

As noted in Sec. I, frequency diagrams were plotted for each property in each sediment type in each environment, and the mean and median values were indicated. These mean and median values were insignificantly different save in one case: clayey silt in the continental terrace. In this case the

TABLE I. Continental terrace (shelf and slope) environment; average sediment size analyses and bulk grain densities.

Sediment type	No. samples	Mean grain size mm	Mean grain size ϕ	Sand, %	Silt, %	Clay, %	Bulk grain density, g/cm^3
Sand							
Coarse	2	0.5285	0.92	100.0	0.0	0.0	2.710
Fine	28	0.1638	2.61	92.2	4.1	3.7	2.709
Very fine	16	0.0988	3.34	81.0	12.5	6.5	2.680
Silty sand	40	0.0529	4.24	57.0	30.9	12.1	2.677
Sandy silt	47	0.0340	4.88	30.3	57.8	11.9	2.664
Silt	19	0.0237	5.40	7.8	80.1	12.1	2.661
Sand–silt–clay	29	0.0177	5.82	31.7	42.9	25.4	2.689
Clayey silt	105	0.0071	7.13	7.4	58.3	34.3	2.656
Silty clay	54	0.0022	8.80	3.9	34.8	61.3	2.715

TABLE II. Continental terrace (shelf and slope) environment; sediment densities, porosities, sound velocities, and velocity ratios.[a]

Sediment type	Density, g/cm^3 Av.	SE	Porosity, % Av.	SE	Velocity, m/s Av.	SE	Velocity ratio Av.	SE
Sand								
Coarse	2.034	...	38.6	...	1836	...	1.201	...
Fine	1.962	0.017	44.5	1.0	1759	9	1.152	0.006
Very fine	1.878	0.017	48.5	1.0	1709	14	1.120	0.009
Silty sand	1.783	0.014	54.2	0.8	1658	7	1.086	0.005
Sandy silt	1.769	0.018	54.7	1.1	1644	7	1.076	0.004
Silt	1.740	0.027	56.2	1.6	1615	6	1.057	0.004
Sand–silt–clay	1.575	0.021	66.3	1.4	1582	7	1.036	0.005
Clayey silt	1.489	0.014	71.6	0.7	1546	3	1.012	0.002
Silty clay	1.480	0.010	73.0	0.5	1517	2	0.990	0.001

[a] Laboratory values: 23 °C, 1 atm; density: saturated bulk density; porosity: salt free; velocity ratio: velocity in sediment/velocity in sea water at 23 °C, 1 atm, and salinity of sediment pore water. SE: standard error of the mean. Standard deviation, SD, can be computed with: SD = (SE) (no. samples)$^{1/2}$. Median values [rather than mean (av.) values] are recommended for predicting values for clayey silt; these are: density: 1.484 g/cm^3; porosity: 72.5%; velocity: 1534 m/s; velocity ratio: 1.006 (see text for discussion).

TABLE III. Abyssal plain and abyssal hill environments; average noncalcareous sediment size analyses and bulk grain densities.

Enviroment Sediment type	No. samples	Mean grain size mm	Mean grain size ϕ	Sand, %	Silt, %	Clay, %	Bulk grain density, g/cm^3
Abyssal plain							
Clayey silt	24	0.0052	7.59	4.2	55.7	40.1	2.655
Silty clay	57	0.0022	8.81	4.0	34.5	61.5	2.672
Clay	6	0.0014	9.53	0.0	22.2	77.8	2.663
Bering Sea and Okhotsk Sea (siliceous–diatomaceous)							
Silt	1	0.0179	5.80	6.5	76.3	17.2	2.474
Clayey silt	5	0.0049	7.68	8.1	49.1	42.8	2.466
Silty clay	23	0.0024	8.71	3.0	37.4	59.6	2.454
Abyssal hill							
Deep-sea ("red") pelagic clay							
Clayey silt	17	0.0056	7.49	3.9	58.7	37.4	2.678
Silty clay	60	0.0023	8.76	2.1	32.2	65.7	2.717
Clay	45	0.0015	9.43	0.1	19.0	80.9	2.781

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TABLE IV. Abyssal plain and abyssal hill environments; noncalcareous sediment densities, porosities, sound velocities, and velocity ratios.[a]

Environment / Sediment type	Density, g/cm³ Av.	Density SE	Porosity, % Av.	Porosity SE	Velocity, m/s Av.	Velocity SE	Velocity ratio Av.	Velocity ratio SE
Abyssal plain[b]								
Clayey silt	1.454	0.024	74.2	1.4	1528	3	0.999	0.002
Silty clay	1.356	0.015	80.1	0.8	1515	2	0.991	0.001
Clay	1.352	...	80.0	...	1503	...	0.983	...
Bering Sea and Okhotsk Sea (siliceous–diatomaceous)								
Silt	1.447	...	70.8	...	1546	...	1.011	...
Clayey silt	1.228	0.019	85.8	0.9	1534	2	1.003	0.001
Silty clay	1.214	0.008	86.8	0.4	1525	2	0.997	0.001
Abyssal hill								
Deep-sea ("red") pelagic clay								
Clayey silt	1.347	0.020	81.3	0.9	1522	3	0.995	0.002
Silty clay	1.344	0.011	81.2	0.6	1508	2	0.986	0.001
Clay	1.414	0.012	77.7	0.6	1493	1	0.976	0.001

[a] Laboratory values: 23 °C, 1 atm; density: saturated bulk density; porosity: salt free; velocity ratio: velocity in sediment/velocity in sea water at 23 °C, 1 atm, and salinity of sediment pore water. SE: standard error of the mean. Standard deviation, SD, can be computed with: SD = (SE) (no. samples)$^{1/2}$.
[b] For approximate properties of thinner, coarse-grained layers in abyssal plain turbidites: see continental terrace Tables I and II in the fine sand to sand–silt–clay sizes (silt is most common).

TABLE V. Properties of calcareous sediments from the equatorial Pacific: average size analyses, bulk grain densities, and calcium carbonate contents.[a]

Sediment type / Area	No. samples	Mean grain size mm	Mean grain size ϕ	Sand, %	Silt, %	Clay, %	Bulk grain density, g/cm³	$CaCO_3$, %
Clayey sand								
Ontong-Java Plat.	24	0.0296	5.08	51.4	17.4	31.2	2.688(12)	85.5(11)
Sand–silt–clay								
Ontong-Java Plat.	34	0.0154	6.02	37.3	22.3	40.4	2.703(16)	83.6(17)
East Pacific	44	0.0091	6.78	23.5	28.9	47.6	2.654(28)	79.7(33)
Clayey silt								
East Pacific	29	0.0078	7.00	15.1	45.9	39.0	2.671(9)	55.7(17)
Sandy clay								
Ontong-Java Plat.	24	0.0076	7.04	26.7	18.3	55.0	2.701(14)	80.0(11)
Silty clay								
Ontong-Java Plat.	17	0.0042	7.89	16.6	21.3	62.1	2.701(5)	77.8(11)
East Pacific	151	0.0056	7.48	14.1	33.3	52.6	2.653(51)	66.2(93)

[a] Numbers in parentheses after grain densities and $CaCO_3$ percents indicate number of samples for these tests.

TABLE VI. Properties of calcareous sediments from the equatorial Pacific: densities, porosities, sound velocities, and velocity ratios.[a]

Sediment type / Area	Density, g/cm³ Av.	Density SE	Porosity, % Av.	Porosity SE	Velocity, m/s Av.	Velocity SE	Velocity ratio Av.	Velocity ratio SE
Clayey sand								
Ontong-Java Plat.	1.493	0.004	71.8	0.2	1596	3	1.043	0.002
Sand–silt–clay								
Ontong-Java Plat.	1.497	0.004	71.8	0.2	1577	4	1.031	0.003
East Pacific	1.404	0.011	76.9	0.6	1540	1	1.007	0.001
Clayey silt								
East Pacific	1.353	0.014	80.0	0.8	1535	1	1.004	0.001
Sandy clay								
Ontong-Java Plat.	1.490	0.004	72.2	0.3	1555	3	1.017	0.002
Silty clay								
Ontong-Java Plat.	1.481	0.007	72.7	0.4	1540	2	1.007	0.001
East Pacific	1.376	0.007	78.6	0.4	1534	1	1.003	0.0004

[a] Laboratory values: 23 °C, 1 atm; density: saturated bulk density; porosity: salt free; velocity ratio: velocity in sediment/velocity in sea water at 23 °C, 1 atm, and salinity of sediment pore water. SE: standard error of the mean. Standard deviation, SD, can be computed with: SD = (SE) (no. samples)$^{1/2}$.

frequency distribution diagrams of velocity and the velocity ratio (Fig. 2) showed a positively skewed distribution. In such distributions it is good practice to give both mean and median, and to favor the median as an "average" for predictions (e.g., Downie and Heath[14]); these values for clayey silt are less than 1% apart, but both are listed in Table II.

The properties of deep-sea calcareous sediments from two general areas of the equatorial Pacific are listed in Tables V and VI from a separate report.[1] Part of the abstract of this report is included in the next paragraph; the full report includes diagrams, regression equations, and discussions (as in the present report).

Three expeditions to the Ontong-Java Plateau and the eastern equatorial Pacific obtained unusually good samples of calcareous sediments from box and piston cores; water depths ranged from 1600 to 4900 m. Two sedimentary environments are represented; a plateau top and side slope, and the deep, eastern equatorial Pacific; these are shown separately in the tables. Dissolution, dilution, and winnowing by currents cause reductions (with increasing water depth) of $CaCO_3$, percent sand, mean grain size, sediment rigidity, and sound velocity. The best indices to predict sound velocity are percent sand, mean grain size, and the velocity ratio; this ratio varies from about 1.05 on top of the Plateau to about 1.00 at 4400 m. Hollow tests (shells) of Foraminifera act as solid particles in transmitting sound. Density and porosity are good predictors of velocity in the east Pacific, but not in the Plateau area because of large amounts of hollow Foraminifera. Eastern Pacific sediment has higher porosities and lower density (than the Plateau samples) because of less $CaCO_3$ and more biogeneous silica. As water depth increases from 1600 to 4900 m, percent sand and mean grain size decrease markedly, but total porosity increases only 1% to 3%. This is due to dissolution and breakdown of hollow tests of Foraminifera and transfer of intraparticle porosity (within the tests) to interparticle porosity between the grains.

The question has been raised: are the properties of a sediment type from the Pacific about the same as in the same sediment type from the same environment in the Atlantic, Mediterranean, or Indian Ocean? The answer is yes, if porosities are about the same; examples are given in the last section.

Fortunately for predictions of sediment properties, there are a relatively few common minerals which make up the structure of most surficial sediments. In terrigenous sediments these minerals are usually quartz, the feldspars, calcite, micas, clay minerals, and very minor amounts of certain heavy minerals.[15,16] In pelagic sediments the most common minerals are the clay minerals, and biogenous calcite and silica. As a result, the bulk elastic properties of mineral grains in a given sediment type in a given environment are apt to be about the same the world over. In a two-phase system, (minerals and water), the sea water varies very little in properties given the same temperature, pressure, and various expectable salinities. If one is working with velocity or the velocity ratio, sea water is not a very variable factor. When porosity is about the same, the volume of sea water is about the same, and one would expect sound velocity to be about the same for the same sediment type in the same environment in any ocean.

In the present data set, the information of the abyssal plains requires some amplification. These plains are composed of sediments (turbidites) deposited from turbidity currents moving along the sea floor. The sediment mass is usually composed of thicker silt–clay layers alternating with thinner layers of sand and silt. Due to selectivity of samples in this study, only about the upper 30 cm of these sediments are reported. Almost all of these are higher-porosity silt–clays and the deeper, common, coarser-grained layers are not represented (the very few available samples were eliminated to avoid misleading averages). These coarser-grained layers have the same properties as found in the continental terrace (Tables I and II) in the range from fine sand to silt; silt is most common. For modeling purposes, values from the continental terrace can be used. An example of this type of modeling was discussed by Gilbert.[17] Variations in layer thicknesses and properties in turbidites are well illustrated by Horn *et al.*[18–20] and by Tucholke.[21] Tucholke made the point (with which we agree) that, at lower frequencies, the sediment surface properties may not suffice for modeling sound interactions with the sea floor, and that lower layers must be considered, and possibly averaged, for reporting the sediment "Surface." Examples of this averaging are shown by Tucholke[21] and by Hamilton[12] (Appendix A).

B. Porosity

In a gas-free sediment, the volume of voids (or pore spaces between mineral grains) occupied by water is expressed as porosity (volume of voids/total volume). In natural marine sediments, porosity usually ranges between about 35% and about 90%; average values for the various sediment types are listed in the tables. In general, porosity increases with decrease in grain size (Fig. 3). Although not easily seen in Fig. 3, there are significant differences in porosity in the various environments, given the same grain size. For example, entering the mean grain size versus porosity equations (Appendix) at a mean grain size of 8ϕ (0.004 mm),

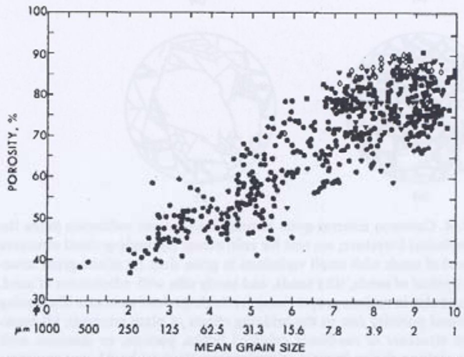

FIG. 3. Mean grain size versus porosity. Round dots represent continental terrace (shelf and slope) samples; squares: abyssal hill samples; triangles: abyssal plain samples; open diamonds: diatomaceous samples from the Bering and Okhotsk Seas.

percent porosities are: continental terrace: 72.6; abyssal plains: 76.7; and abyssal hills, 80.2. The siliceous sediments of the Bering and Okhotsk Seas (not in the regression equations) have porosities of about 86% at about 8ϕ. These differences are the results of the varying mineralogy and microfabrics, as discussed below.

There is much scatter in the relationship between mean grain size and porosity because of a number of interrelated factors (see Mitchell[22] for full discussions). The most important factors are grain size, uniformity of grain size (sorting), grain shape, packing of grains, and mineralogy.

Sand-size mineral grains are too large to be affected by interparticle forces or adsorbed water. When they fall to the sea floor, they assume positions among other grains under the influence of gravity and water motions. In a well-sorted sand a single-grained structure is formed [Fig. 4(a)]; the grains are in solid contact. The porosity of this structure depends largely on packing. It has been shown[23] that equal-sized spheres, if regularly packed, can have porosities ranging from 47.6 in the loosest possible arrangement to 26% in the densest arrangement. Natural sands, and even centrifuged or vibrated natural sands in the laboratory, rarely have porosites less than 35%. In the present study, the average porosity in fine sand is 44.5%, and in very fine sand is 48.5%; the overall range in these sands is 37% to 57%.

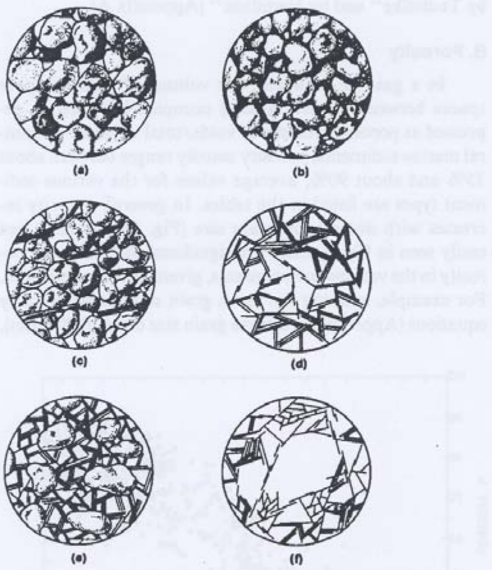

FIG. 4. Common mineral-grain structures of marine sediments (from the geotechnical literature; see text for references): (a) single-grained structure (typical of sands with small variations in grain size); (b) mixed-grain structure (typical of sands, silty sands, and sandy silts with admixtures of sand, silt, and clay in various proportions); (c) single-grained structure illustrating increased porosity due to the bridging effects of platy minerals; (d) bookhouse structure of randomly oriented books, packets, or domains with short, linking chains (typical of terrigenous clay); (e) bookhouse structure with silt- or sand-size particles suspended in a clay matrix (typical of a terrigenous silty clay or clayey silt); (f) pelagic clay structure of randomly oriented single plates, packets, and flocs with long, linking chains.

When finer-size sand, silt, and clay particles are present in a sand, a mixed-grained structure is formed wherein the finer particles fill many interstices between the larger sand grains [Fig. 4(b)]. The presence of the finer material in the pore spaces tends to reduce porosity, but some of the finer material also is more irregular in shape and lies between the larger grains causing a net tendency toward increasing porosity (Mitchell[22], p. 143). In mixed-grain materials such as silty sand, the average porosity in the present data set (Table II) is 54%. When platy minerals such as biotite are present there can be a bridging effect [Fig. 4(c)] which causes porosity to be greater than might be expected from the grain size.

Fine silt and clay particles form higher-porosity structures which are distinctly different from the single- and mixed-grain structures. These fine particles have adsorbed water layers on their surfaces and form structures controlled by interparticle forces. When these particles arrive on the sea floor they are apt to adhere to the first particles they contact, and to be held there by interparticle forces to form three-dimensional structures of the types illustrated in Figs. 4(d) to 4(f).

The subject of the microstructure or fabric of clays has received increasing interest during the past two decades because clay fabric determines the physical properties of clays. Progress in this field accelerated when the electron microscope became available to clay-fabric researchers. There are several recent, excellent reviews of this subject.[22,24–29] The following information is drawn from these and other referenced sources.

The first widely accepted diagrammatic representation of clay structure was the "honeycomb" structure of Terzaghi[30] and Casagrande.[31] Later, Goldschmidt[32] and Lambe[33] proposed a structure in which single clay platelets were arranged in mostly edge to face contacts similar to a cardhouse. The honeycomb and cardhouse structures are illustrated in the reviews noted above and in Hamilton *et al.*[34] and Hamilton.[35]

Early use of the electron microscope appeared to confirm the cardhouse, single-platelet structure. Later studies, however, have concluded that structures composed of single platelets of clay minerals are probably rare. The favored concepts at present are that the clay platelets adhere in face-to-face arrangements to form books or packets (also called domains). These packets may be in perfect stacks or in face-to-face arrangements similar to fallen stacks of books. The packets form randomly oriented structures with linking chains. In terrigenous sediments (such as in the Gulf of Mexico) the linking chains are short [Fig. 4(d)].[24,25,28,29] This type of structure is probable for the silt–clays for the abyssal plains where the average porosity in the present data set (87 samples) is 79%; or in the continental terrace silt–clays (159 samples) in which the average porosity is 72%.

An important concept in microfabrics is illustrated in Fig. 4(e). When silt- or sand-size grains are present in a dominantly clay fabric they are apt to be suspended in the clay matrix and do not touch.[36,37] This is important because the elastic properties of such structures are largely determined by the clay structure.

In deep-sea pelagic clays the clay packets may form flocs and some single clay platelets may form fabrics similar to the cardhouse structure [Fig. 4(f)]. These flocs, packets, and cardhouse structures appear to be connected with long, linking chains (Bennett *et al.*[28,29]). These authors note that varying porosities appear to be related to the lengths of the connecting chains; thus forming larger and smaller inter-voids between the fabric units. In deep-sea pelagic clay in this data set, porosities range from 71% to 91%, with an average (122 samples) of 80%.

A possible exception to the multiplate or packet fabrics has been noted by Krinsley and Smalley.[38] They report that when quartz particles are in the clay- and fine-silt sizes, they are apt to be flat plates which may form open structures of the cardhouse type. It seems possible that this may also be true for fine platelets of calcium carbonate resulting from the presence of nannoplankton and fragments of tests of Foraminifera which have been broken down by dissolution or crushing.

Pelagic sediment composed of biogenous silica (Radiolaria and diatoms) forms strong, highly porous structures (not illustrated). In the siliceous sediments from the Bering and Okhotsk Seas (Table IV), the average porosity (29 samples) is 86%.

The later reviews[25,28,29] show convincing evidence (through electron microscopy) of the orientation of clay platelets, flocs, and particles under overburden pressure (long axes normal to the pressure) and the reduction of porosity under pressure. Pusch[39] and Mitchell[22] emphasized that failure of the clay fabric under normal shear pressure was probably mostly through the links or chains between packets. This reorientation of mineral particles under overburden pressure is important in sound propagation in the sea floor because it may cause anisotropic sound velocity relationships (velocity greater parallel to the sea floor).[35,40]

C. Density

The saturated bulk density, ρ_{sat}, of a gas-free sediment can be computed with the equation: $\rho_{sat} = n\rho_w + (1-n)\rho_s$; where n is fractional porosity, ρ_w is density of pore water, and ρ_s is the bulk density of the mineral grains. Regression equations (Appendix) are given for density versus porosity (not illustrated) in the three environments.

It can be assumed that pore water in surficial sediment has about the same salinity as the bottom water (e.g., Siever *et al.*[41]). In the laboratory, sea-water density can be obtained from Sigma-T tables (e.g., NAVOCEANO[42,7]) for given values of salinity and temperature at one atmosphere pressure. In the present data set these values, at 23 °C, ranged from 1.022 g/cm³ for the shallow Bering Sea to 1.027 g/cm³ for the deep Mediterranean (where salinities are greater than 38 ppt); a value of 1.024 g/cm³ can be used for most deep-water sediments.

Tables I-VI list average values of bulk mineral-grain densities for the sediment types in the various environments. The overall average values in these environments are found in Table VII.

These values are close to previous reports: Keller and Bennett[43] reported an average of 2.67 g/cm³ for terrigenous materials, and 2.71 g/cm³ for calcareous materials in the North Pacific and North Atlantic; Keller *et al.*[6] reported 2.72 g/cm³ for the U. S. Atlantic continental slope; and Akal[44] computed a general value of 2.66 g/cm³.

TABLE VII. The overall average values in various environments.

Environment Sediment type	No. samples	Av. bulk grain density, g/cm³
Continental terrace		
Terrigenous	340	2.678
Abyssal plain		
Terrigenous	87	2.667
Siliceous (diatomaceous)	29	2.457
Abyssal hill		
Pelagic clay	122	2.735
Ontong-Java plateau		
Calcareous	47	2.698

It can be seen that there is little variation in averages of bulk grain densities except when biogenous silica is present. This is because there are relatively few common minerals in most sea floor sediments and their mineral densities and percentages present usually balance to aggregate values between 2.6 and 2.8 g/cm³, with most between 2.65 and 2.75 g/cm³. Least values of density in the common minerals of marine sediments are found in biogenous silica (e.g., diatoms, Radiolaria). Hurd and Theyer[45] measured densities in Radiolaria from the equatorial Pacific; the range for 15 measurements was 1.67 to 2.06 g/cm³, with an average of 1.92 g/cm³. Baas Becking and Moore[46] measured values of 2.1 and 2.3 in diatomite.

Biogenous silica is mixed with pelagic clay or calcareous materials in large areas of the equatorial Pacific. These mixtures result in lower grain densities than expectable in pelagic clay or calcareous sediment. In this area, Keller and Bennett[47] reported averages of 2.40 g/cm³ for Radiolarite, and 2.30 for Radiolarian brown clay; Richards *et al.*[48] reported 2.31 g/cm³ for 11 samples of siliceous pelagic clay. In the present data set there is a value of 2.53 g/cm³ for the grain density of a siliceous calcareous sediment. In the deep Bering Sea, where the sediment is highly siliceous, there are enough terrigenous minerals to bring the average grain density up to 2.46 g/cm³.

Figure 5 illustrates the relationships between mean grain size and saturated bulk density in the three environments. The large scatter in the relationships is due to the same causes as for the scatter in the porosity versus mean grain size diagram (Fig. 3), previously discussed. This is because there is a linear relationship between density and porosity at any given grain density (see equation in the first paragraph of this section). As is the case for porosity, there are significant differences in saturated bulk density in the three environments at any given grain size. For example, entering the mean grain size versus density equations (Appendix) at a mean grain size of 8 ϕ (0.004 mm), densities are: continental terrace, 1.49 g/cm³; abyssal plains, 1.41 g/cm³; abyssal hill, 1.37 g/cm³. The siliceous sediments of the Bering and Okhotsk Seas have average saturated bulk densities of 1.22 g/cm³ at about 8.5 ϕ (0.003 mm).

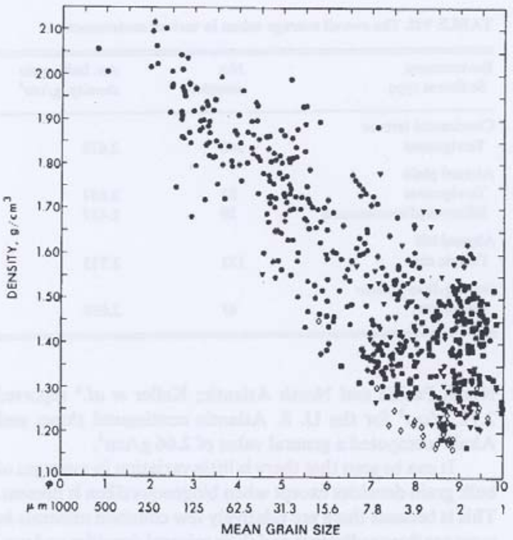

FIG. 5. Mean grain size versus saturated bulk density, all environments; symbols as in Fig. 3.

In predicting *in situ* properties of the sea floor surface, it can be assumed that porosity is the same *in situ* as in the laboratory. The laboratory value of saturated bulk density can also be used as the *in situ* density. The correction for oceanic depths can be easily made (using the equation in the first paragraph of this section), but the increased pore-water density at a water depth of 6000 m, for example, would increase laboratory to *in situ* saturated bulk density by no more than about 0.03 g/cm^3.

D. Compressional wave (sound) velocity

Wave velocities are elastic properties of the two-phase sediment mass (sea water in pores and the mineral structure). Properties such as porosity and grain size affect sound velocity only through their effects on the elasticity of the sediment. This subject was summarized in other reports[49,50] with appropriate equations. It was concluded that the elastic properties of water-saturated sediment could be expressed through the Hookean elastic equations unless attenuation is considered; in which case linear viscoelastic equations are recommended.[12,50]

The basic equation for the velocity of a compressional wave V_p is

$$V_p = [(\kappa + 4/3\mu)/\rho]^{1/2}, \tag{1}$$

where

κ is incompressibility or the bulk modulus and equals $(1/\beta)$,
μ is the shear (rigidity) modulus,
ρ is saturated bulk density,
β is compressibility.

When a medium lacks rigidity, Eq. (1) becomes

$$V_p = (\kappa/\rho)^{1/2}, \tag{2a}$$

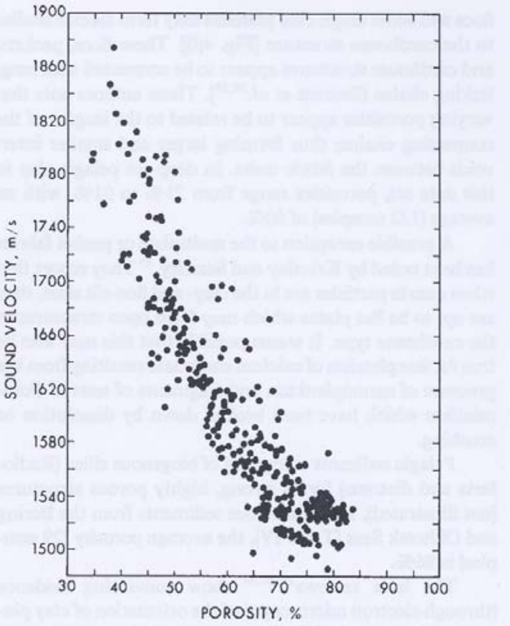

FIG. 6. Porosity versus sound velocity, continental terrace (shelf and slope).

or

$$V_p = (1/\beta\rho)^{1/2}. \tag{2b}$$

Compressibility β and density ρ in Eq. (2b) have been expanded[51–53] into

$$V_p = \left(\frac{1}{[n\beta_w + (1-n)\beta_s][n\rho_w + (1-n)\rho_s]}\right)^{1/2}, \tag{3}$$

where n is the volume of pore space occupied by water (fractional porosity), and subscripts s and w indicate mineral solids and water. Equation (3) is known as the Wood Equation (Wood[51]) and applies to a suspension (such as mineral particles in water), or to any medium lacking rigidity.

The sound velocity–porosity relationship (Figs. 6 and 7) has received much attention in past studies because porosity is an easily measured property which usually allows reasonable predictions of velocity. This is because the volume of water in the pore spaces largely determines compressional wave velocity. This is a result of the large compressibility of the water relative to the much smaller compressibility of the mineral grains. The net effect of the varying volumes of these two constituents, when used with the Wood Equation (3) is shown as the curve in Fig. 7. It can be seen that all of the abyssal hill and plain sediments have velocities higher than would have been predicted by the Wood Equation for suspensions. These higher velocities are due to the presence of rigidity and a frame bulk modulus in the mineral structure of the sediments (discussed in Hamilton[49]). In earlier work with high-porosity silt–clays of the continental terrace, which had low rigidities, the Wood Equation was successfully used

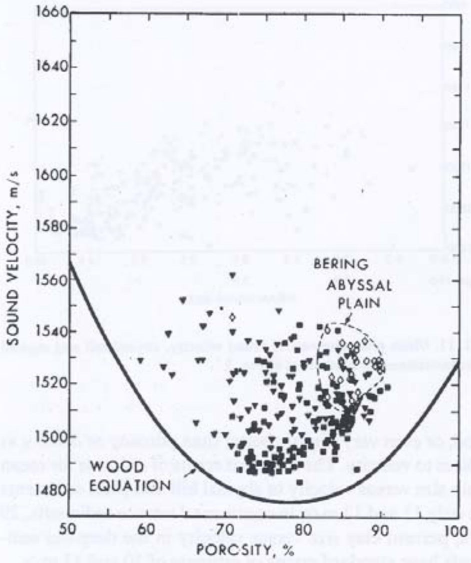

FIG. 7. Porosity versus sound velocity, abyssal hill and abyssal plain environments; see text for discussion of Wood Equation; symbols as in Fig. 3.

(Hamilton[53]), but later measurements indicated that almost all marine sediments have significant rigidities and transmit shear waves; consequently, it was recommended (Hamilton[49]) that the Wood Equation (3) be abandoned (for water-saturated sediments) in favor of Eq. (1).

In the abyssal-hill environment (Fig. 7) sound velocity decreases slightly as porosity decreases from about 80% to about 75% (a trend predicted by the Wood Equation). In the other two environments, velocity increases with decreasing porosity into the high velocities (greater than 1800 m/s) in low-porosity sands (Fig. 6). In Fig. 7, the Bering Abyssal Plain sediments have significantly higher velocities at high porosites. This is apparently due to a relatively rigid mineral structure composed of low-density silica.

It is worthwhile to consider each of the three environments separately. At 75% porosity the sediment velocities in the three environments (from regression equations, Appendix) are: abyssal hill, 1499 m/s; abyssal plain, 1520 m/s; and the continental terrace, 1531 m/s.

In Fig. 7, the point where the Wood Equation curve crosses the 100% porosity line (at about 1530 m/s) representes sound velocity in sea water at 23 °C and 1-atm pressure. It can be seen that most of the sediments have velocities less than that in sea water. This is also true of higher-porosity continental terrace sediments (Fig. 6). The sound velocity ratios (sediment velocity/water velocity) in Tables II and IV also indicate that, on the average, all of the silt–clays of the abyssal hill and plain environments, and silty clay in the continental terrace, have velocities less than one, indicating that the average velocity in these sediments would be less than velocity in sea water. This low-velocity effect is the result, at high porosities, of a balance between water and mineral compressibilities (or bulk moduli) and densities, plus low rigidities and low mineral frame bulk moduli (as discussed in a previous report[49]). This low-velocity phenomenon was noted in early laboratory studies, and has been confirmed by measurements in the sea floor (e.g., Refs. 18, 21, and 53–55). The lowest velocities measured in the higher porosity sediments are usually no more than 3% less than in sea water at the same temperature and pressure (Figs. 6 and 7; Tables II and IV). This was also a result of recent measurements *in situ* and in cores from the North Atlantic.[21,55]

The sound velocity ratio is important when studying reflection and refraction of sound waves incident on the sea floor. This ratio is also useful in predicting *in situ* sediment-surface velocities because the ratio remains the same in the laboratory or at any water depth in the sea. To correct a laboratory velocity to the sea floor, simply multiply the velocity ratio by the bottom-water velocity. This has the same result as making full temperature and pressure corrections to the laboratory measurement using the tables for the speed of sound in sea water.

The relationships between density and sound velocity (Figs. 8 and 9) are similar to those of porosity and velocity because of the linear relationship between density and porosity.

Mean grain size in all environments, and percent clay-size material in the deep-sea environments, were determined in early work to be important indices to sound velocity.[18,34,35,53,54,56] The present study verifies these earlier studies: sound velocity increases with increasing mean grain size (Figs. 10 and 11), and decreasing amounts of clay-size material (Fig. 12).

The relationships between grain-size data and velocity are especially important because grain size analyses can be made of dried sediment in which density, porosity, and ve-

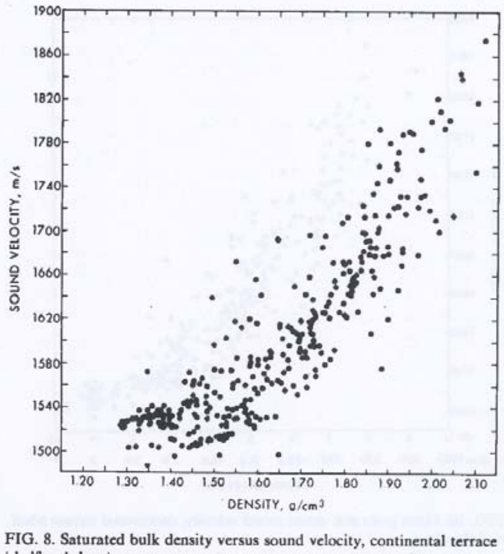

FIG. 8. Saturated bulk density versus sound velocity, continental terrace (shelf and slope).

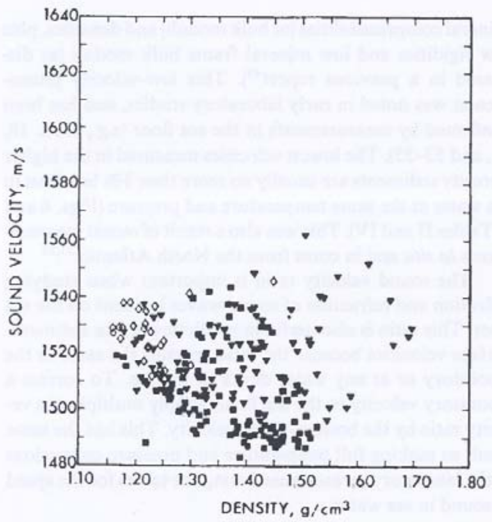

FIG. 9. Saturated bulk density versus sound velocity, abyssal hill and abyssal plain environments; symbols as in Fig. 3.

locity measurements cannot be made. Additionally, there is much data on grain sizes in geologic literature which can be used as indices to acoustic properties.

Sediment mean grain size, or clay-size percentages, affect velocity only through other elastic properties such as density and porosity.[35,57,58] However, mineral-grain shape may influence sediment structural rigidity and thereby affect sound velocity in very minor ways.[49]

Examination of the regression equations in the Appendix indicates that mean grain size and percent clay size are as

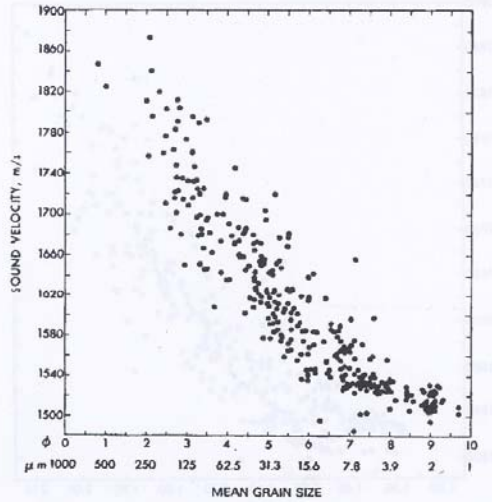

FIG. 10. Mean grain size versus sound velocity, continental terrace (shelf and slope).

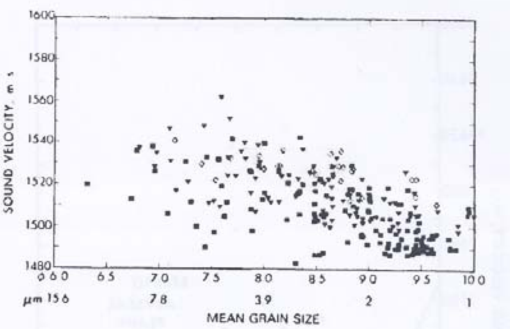

FIG. 11. Mean grain size versus sound velocity, abyssal hill and abyssal plain environments; symbols as in Fig. 3.

good, or even very slightly better than porosity or density as indices to velocity. The standard errors of estimate for mean grain size versus velocity in abyssal hill and plain sediments are only 11 and 12 m/s; in continental terrace sediments, 29 m/s; percent clay size versus velocity in the deep-sea sediments have standard errors of estimate of 10 and 11 m/s.

E. Impedance

The characteristic impedance of a medium is the product of density and velocity. This important property determines the amount of energy reflected when sound energy passes from one medium into another of different impedance. Echo-sounding and continuous-reflection records indicate the travel time of sound between impedance mismatches at the sea floor, or in the sediment or rock body. Impedances were computed for the sediments of this study and are plotted against density in Figs. 13 and 14; average values can be computed from the tables. The use of these impedances and those of sea water to compute Rayleigh reflection coefficients and bottom losses at normal incidence was discussed in an earlier report.[59] Laboratory impedances must be corrected to get *in situ* values. This can be done by first correcting sound velocity and density as previously discussed.

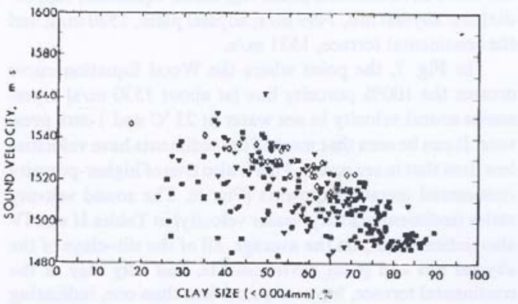

FIG. 12. Clay-size particles versus sound velocity, abyssal hill, and abyssal plain environments; symbols as in Fig. 3.

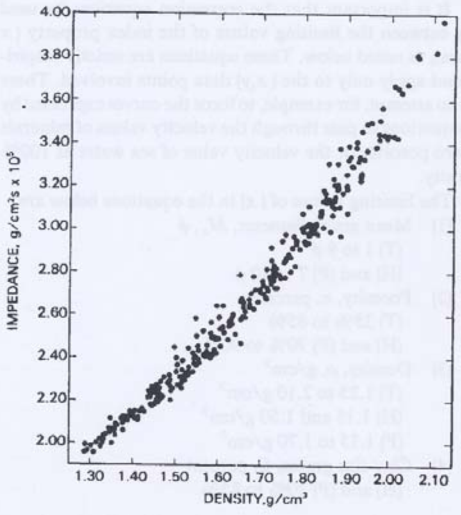

FIG. 13. Saturated bulk density versus impedance, continental terrace (shelf and slope).

F. Prediction of properties

Several questions arise when laboratory measurements are made on cored sediments and the results organized into sediment types in the principal environments, and presented for use in predictions. Are the values and relationships expressed in the tables and regression equations accurate predictors of properties in similar sediments in areas not sampled? And, are the laboratory values (after corrections) reasonably accurate predictors of *in situ values*? The answer to these questions appears to be yes.

In examining prediction capabilities, it is a valid approach to compare values of sediment properties from previously published tables (e.g., Hamilton, 1980; Ref. 12) and regression equations[2] with values measured in our laboratory since 1980. This has been done for sound velocity for a fairly full range of sediment types from the continental terrace (shelf and slope) of several widely separated regions as shown in Table VIII (the number in parentheses is the number of samples).

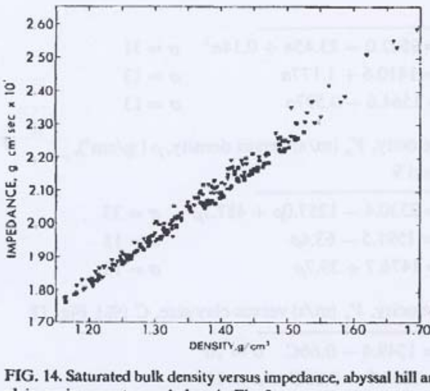

FIG. 14. Saturated bulk density versus impedance, abyssal hill and abyssal plain environments; symbols as in Fig. 3.

TABLE VIII. Comparison of values of sediment properties from previously published tables and regression equations with values measured in our laboratory since 1980.

Area / Sediment type	Percent difference in sound velocity from: Table Av.[a]	Regression Eq.[b]
Bering Sea		
Fine sand (6)	2.6	1.7
Very fine sand (4)	1.8	1.0
Silty sand (4)	2.8	0.5
North Sea		
Sandy silt (17)	0.5	0.6
Silty sand (5)	1.9	1.0
Mediterranean		
Clayey silt (21)	0.3	0.7
Silty clay (35)	0.2	0.7

[a] Hamilton[12].
[b] Mean grain size versus sound velocity (Hamilton[2]).

The conclusions are that the regression equations (mean grain size versus velocity) would have been accurate in predicting average velocities within 1% or less for most of these sediments (given the measured mean grain size), and that without any information except sediment type the averages in the table would have predicted the average velocities within 0.2% to 2.8%.

Abyssal hill pelagic clay is very widespread in the world ocean, and properties vary little compared with sediments in the other environments; thus making predictions accurate. For example, the average velocity ratio for 41 samples of pelagic clay in the 1970 table[35] was 0.985. This ratio in the 1980 table (122 samples) was 0.984.

It is important to verify measurements such as in the present tables with measurements by other organizations in different areas, and especially in the case of sound velocity, by *in situ* measurements. Some recent verifying measurements are noted below.

Bennett *et al.*[60] reported the results of a major sediment-properties program of the National Oceanic and Atmospheric Administration (NOAA) involving over 90 cores (most of which were high-quality hydroplastic gravity cores) in silty clay of the U.S. Atlantic continental slope and rise. The average value of density and porosity from the NOAA study (they didn't measure sound velocity) are compared with those for silty clay in Table II are found in Table IX.

TABLE IX. The average values of density and porosity from the NOAA study (they didn't measure sound velocity) are compared with those for silty clay in Table II.

	Density, g/cm³	Porosity, %
Continental slope	1.52	71
Continental rise	1.49	72
Table II (silty clay)	1.48	73

Measurements of sound velocity were made by Baldwin *et al.*[61] in three cores in turbidites and pelagic clay from the Atlantic using the Giant Piston Corer, and a new velocimeter. The results were averaged according to sediment type and corrected to 23 °C (as in the present and previous studies). These investigators pointed out that their velocities were within less than one-half of 1% of those published by Hamilton[35] in 1970 (or in the present tables).

For the present discussion of prediction, some very important measurements of velocity *in situ* in the sea floor have been reported by Shirley,[62] Tucholke and Shirley,[55] and Addy *et al.*[63] These measurements were made (by a velocimeter attached to piston corers) in all three of the general environments discussed in this report.

Addy *et al.*[63] reported the results of *in situ* velocity measurement in the continental terrace (shelf and slope) of the northeast Gulf of Mexico. Seventy-seven piston cores were taken. At most of the stations the *in situ* sediment velocity was measured by Shirley's velocimeter. They reported a velocity ratio of 0.982 for clay (not in Table II). Some velocities in sands are far too low (probably a result of the difficulties of coring sands). The reported averages of density and the velocity ratio in silty clay were 1.43 g/cm³ and 0.986; Table II lists 1.48 g/cm³ and 0.990 for this sediment type.

In the North Atlantic, the average *in situ* velocity ratio in four areas of pelagic clay was 0.984, and in two areas of turbidites was 0.990 (Tucholke and Shirley[55]; Shirley[62]). These ratios were within about 1% of those measured by Tucholke in the cores taken at the time of the *in situ* measurements (and corrected to *in situ* values). The *in situ* values noted above are almost exactly the same as those in Table IV for these sediment types: 0.984 for pelagic clay (122 samples of clayey silt, silty clay, and clay), and 0.993 in turbidites (87 samples of clayey silt, silty clay, and clay). An important conclusion of Tucholke and Shirley[55] and Tucholke[21] was that the method of correcting laboratory to *in situ* velocities developed by Hamilton[64,2] was confirmed in their experiments.

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APPENDIX: REGRESSION EQUATIONS INTERRELATING VARIOUS SEDIMENT PROPERTIES

Regression lines and curves were computed for the illustrated sets of (x,y) data. These constitute the best indices (x) to obtain the 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 standard errors of estimate σ opposite each equation, are applicable only near the mean of the (x,y) values. Accuracy of the (y) values, given (x), falls off away from this region (e.g., Griffiths, 1967, p. 448; Ref. 65). Grain sizes are shown in the logarithmic phi-scale ($\phi = -\log_2$ of grain size in millimeters).

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 the velocity values of minerals at zero porosity or the velocity value of sea water at 100% 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%
  (H) and (P) 70% to 90%
(3) Density, ρ, g/cm³
  (T) 1.25 to 2.10 g/cm³
  (H) 1.15 and 1.50 g/cm³
  (P) 1.15 to 1.70 g/cm³
(4) Clay size grains, C, percent
  (H) and (P) 20% to 85%

Porosity, n (%) versus mean grain size, M_z (ϕ), Fig. 3

(T) $n = 22.01 + 9.24M_z - 0.365M_z^2$ $\sigma = 6.5$
(H) $n = 82.42 - 0.28M_z$ $\sigma = 4.7$
(P) $n = 50.0 + 3.34M_z$ $\sigma = 6.2$

Density, ρ (g/cm³) versus mean grain size, M_z (ϕ), Fig. 5

(T) $p = 2.374 - 0.175M_z + 0.008M_z^2$ $\sigma = 0.11$
(H) $p = 1.327 + 0.005M_z$ $\sigma = 0.09$
(P) $p = 1.869 - 0.057M_z$ $\sigma = 0.11$

Sound velocity, V_p (m/s) versus mean grain size, M_z (ϕ), Figs. 10 and 11

(T) $V_p = 1952.5 - 86.26M_z + 4.14M_z^2$ $\sigma = 29$
(H) $V_p = 1594.3 - 10.2M_z$ $\sigma = 12$
(P) $V_p = 1609.7 - 10.8M_z$ $\sigma = 11$

Sound velocity, V_p (m/s) versus porosity, n (%), Figs. 6 and 7.

(T) $V_p = 2502.0 - 23.45n + 0.14n^2$ $\sigma = 31$
(H) $V_p = 1410.6 + 1.177n$ $\sigma = 13$
(P) $V_p = 1564.6 - 0.597n$ $\sigma = 13$

Sound velocity, V_p (m/s) versus density, ρ (g/cm³), Figs. 8 and 9

(T) $V_p = 2330.4 - 1257.0\rho + 487.7\rho^2$ $\sigma = 33$
(H) $V_p = 1591.5 - 63.4\rho$ $\sigma = 13$
(P) $V_p = 1476.7 + 29.7\rho$ $\sigma = 13$

Sound velocity, V_p (m/s) versus clay size, C (%), Fig. 12

(H) $V_p = 1549.4 - 0.66C$ $\sigma = 10$
(P) $V_p = 1554.7 - 0.65C$ $\sigma = 11$

Density, ρ (g/cm^3) versus porosity, n (%)

(T) $n = 157.6 - 57.8\rho \quad \sigma = 2.1$

(H) $n = 150.1 - 51.2\rho \quad \sigma = 1.1$

(P) $n = 155.9 - 56.0\rho \quad \sigma = 0.8$

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2.6.2 Rocks

For some of the rocks mentioned in paragraph 2.5 the table below gives indicative values for their geo-acoustic parameters.

		Compressional speed [m/s]	Shear speed [m/s]	Density [g/cm^3]
Igneous rocks	Granite	6000	3500	2.65
	Basalt	4800	2500	2.5
Sedimentary rocks	Sandstone	4700	2700	2.5
	Siltstone	3800	2150	2.4
	Mudstone	2200	1000	2.2
	Limestone	6000	3000	2.65

Figure 9 below gives the relation between shear speed and longitudinal (compressional) speed for the various types of rocks discussed.

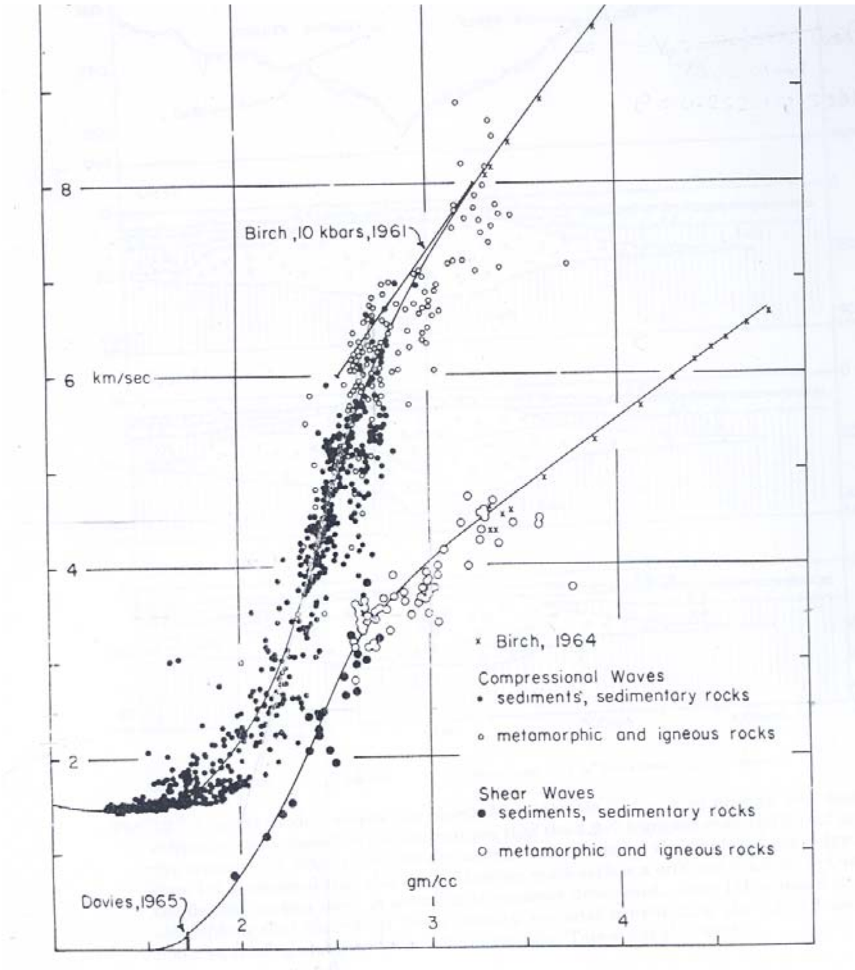

Figure 9

Glossary

Archipelagic apron – a gentle slope with a smooth surface on the seafloor
Bank – an elevation of the seafloor located on a shelf
Basin – a depression of variable extent, generally in a circular or oval form
Borderland – a region adjacent to a continent that is highly irregular with depths in excess of those typical of a shelf
Canyon – a narrow, deep depression with steep slopes
Continental margin – a zone separating the continent from the deeper sea bottom, generally consisting of the rise, slope and shelf
Continental rise – a gentle slope rising toward the foot of the continental slope
Continental shelf – zone adjacent to a continent or island from the waterline to the depth at which there is usually a marked increase f slope to greater depth (shelf break)
Continental slope – zone between the continental shelf and the continental rise
Cordillera – an entire underwater mountain system including all the subordinate ranges, interior plateaus and basins
Escarpment – an elongated and comparatively steep slope of the seafloor separating flat or gently sloping areas
Fan – a gently sloping, fan-shaped feature normally located near the lower termination of a canyon
Fracture zone – an extensive linear zone of unusually irregular topography of the seafloor characterised by large seamounts, steep-sided ridges, troughs or escarpments
Gap – a depression cutting transversely across a ridge or rise
Hill – a small elevation rising less than about 200 m above the seafloor
Hole – a small depression on the seafloor
Knoll – an elevation rising less than about 1 km above the seafloor and of limited extent across the summit
Moat – a depression located at the base of many seamounts or islands
Mountains – a well-delineated subdivision of a large and complex feature, generally part of a cordillera
Plain – a flat, gently sloping or nearly level region of the seafloor
Plateau – a comparatively flat-topped elevation of the seafloor of considerable extent across the summit and usually rising more than 200 m above the seafloor
Province – a region composed of a group of similar bathymetric features whose characteristics are markedly in contrast with the surroundings areas
Range – a series of generally parallel ridges or seamounts
Reef – an offshore consolidation with a depth below the sea surface of less than about 20 m
Ridge – a long, narrow elevation of the seafloor with steep sides
Saddle – a low part on a ridge or between seamounts
Seachannel – a long, narrow, shallow depression of the seafloor, usually occurring on a gently sloping plain or fan, with either a U-shaped or V-shaped cross-section
Seamount – an elevation rising 1 km or more above the seafloor, and of limited extent across the summit
Shoal – an offshore hazard to navigation with a depth below the sea surface of 20 m or less, usually composed of unconsolidated material
Sill – the low part of a ridge or rise separating ocean basins from one another or from the adjacent seafloor
Spur – a subordinate elevation, ridge or rise projecting from a larger feature
Tablemount – a seamount having a comparatively smooth, flat top (also called a bench)
Trench – a long, narrow and deep depression of the seafloor, with relatively steep sides
Trough – a long depression of the seafloor, normally wider and shallower than a trench
Valley – a relatively shallow, wide depression with gentle slopes, the bottom of which grades continually downward (as opposed to a canyon)