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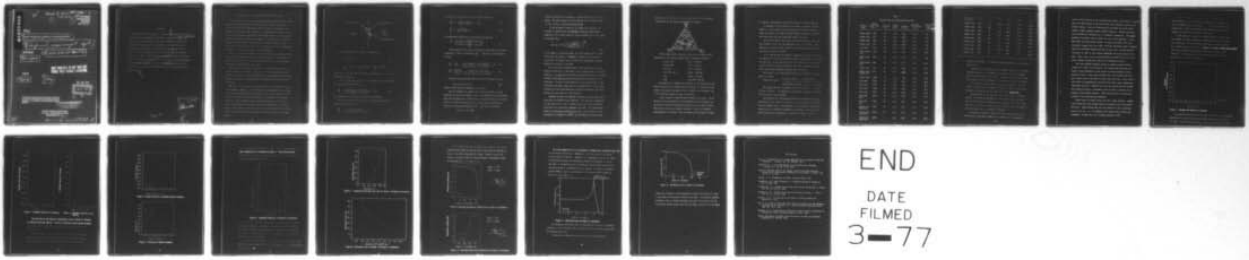
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ACOUSTIC AND PHYSICAL PROPERTIES OF BOTTOM SEDIMENT.(U)
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PREFACE

The purpose of this paper was to summarize the existing studies of the acoustic and physical properties of bottom sediments. ^{were summarized.} The computation of the impedance, reflection coefficient, and the reflection loss for each of the sediments, as well as the graphic presentation, represents the original material, presented in this paper. No new experimental work was carried on, and the sediment core data in Table I were originally prepared by Shumway (1958) and Sutton, Berckhemer, and Nafe (1957). This paper was prepared because it was believed ~~that~~ ^{the} information ~~it contains~~ may be useful to those interested in bottom reflection and its application to bottom bounce sonar.

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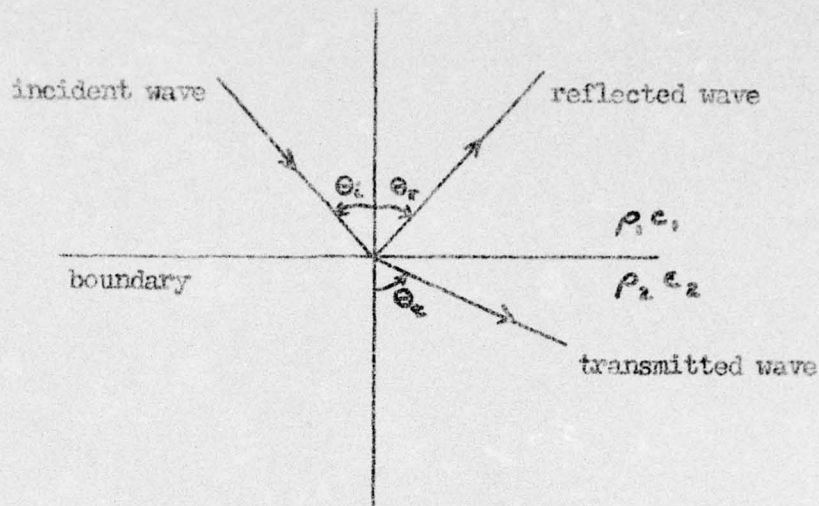
ACOUSTIC AND PHYSICAL PROPERTIES OF BOTTOM SEDIMENTS

The acoustic and physical properties of bottom sediments play a significant role in the bottom reflection method of sound transmission. Reflections from the ocean bottom effect the total transmission loss between a sound source and a receiver. In general these effects are not so readily calculated from known or measurable constants of a region nor may they be readily predicted in advance. An examination of the fundamental relations between the intensity of reflected energy and the constants of the medium is of value in suggesting the order of magnitude of the loss to be expected under various conditions.

When a sound ray is reflected from the ocean bottom, account must be taken of the resultant changes in amplitude and phase. The changes in amplitude and phase for the reflected ray may be given by the Rayleigh reflection coefficient for plane waves.

The ratio of the acoustic pressure in a medium to the associated particle velocity is defined as the specific acoustic impedance of the medium. The specific acoustic impedance is a real quantity of magnitude ρc that is analogous to the mechanical and electrical expressions for resistance.

Consider an ideal plane surface separating two fluid media of acoustic impedances $\rho_1 c_1$ and $\rho_2 c_2$. In the boundary plane between these two media two important conditions must be satisfied at all times. These are (1) continuity of pressure and (2) continuity of normal particle velocity. The first condition requires that the acoustic pressure of the wave in the second medium equal the acoustic pressure of the wave in the first medium. The second condition requires that the component of the particle velocity normal to the boundary plane be the same on both sides of the plane.



In mathematical notation these conditions are:

$$P_i + P_r = P_t \quad (1)$$

$$\frac{P_i}{\rho_1 c_1} \cos \theta_i - \frac{P_r}{\rho_1 c_1} \cos \theta_r = \frac{P_t}{\rho_2 c_2} \cos \theta_t \quad (2)$$

where P_i , P_r , and P_t , are the incident, reflected, and transmitted pressure, and $\theta_i = \theta_r$.

From this the ratio of the acoustic pressures of the reflected and incident waves is found to be

$$\frac{P_r}{P_i} = \frac{\rho_2 c_2 \cos \theta_t - \rho_1 c_1 \cos \theta_i}{\rho_2 c_2 \cos \theta_t + \rho_1 c_1 \cos \theta_i} = R_r \quad (3)$$

This expression is the Rayleigh reflection coefficient.

The ratio of the acoustic pressures of the transmitted and incident waves is found to be

$$\frac{P_t}{P_i} = \frac{2 \rho_2 c_2 \cos \theta_t}{\rho_2 c_2 \cos \theta_t + \rho_1 c_1 \cos \theta_i} \quad (4)$$

At normal incidence ($\theta_i = 0$) these equations reduce to

$$\frac{P_r}{P_i} = R = \frac{\rho_2 c_2 - \rho_1 c_1}{\rho_2 c_2 + \rho_1 c_1} \quad \text{and} \quad (5)$$

$$\frac{P_t}{P_i} = \frac{2 \rho_2 c_2}{\rho_2 c_2 + \rho_1 c_1} \quad (6)$$

The reflection coefficient, R , may also be written as

$$R = \frac{\rho_2 \cos \theta_t - \rho_1 \left[\left(\frac{c_1}{c_2} \right)^2 - \sin^2 \theta_i \right]^{1/2}}{\rho_2 \cos \theta_t + \rho_1 \left[\left(\frac{c_1}{c_2} \right)^2 - \sin^2 \theta_i \right]^{1/2}} \quad (7)$$

These pressure ratios may be transformed into ratios of acoustic intensity by using the relation $I = \frac{P^2}{\rho c}$. The ratio of intensities is then

$$\frac{I_r}{I_i} = \frac{P_r^2}{P_i^2} = \frac{[\rho_2 c_2 \cos \theta_t - \rho_1 c_1 \cos \theta_i]^2}{[\rho_2 c_2 \cos \theta_t + \rho_1 c_1 \cos \theta_i]^2} \quad \text{and} \quad (8)$$

$$\frac{I_t}{I_i} = \frac{P_t^2 (\rho_1 c_1)}{P_i^2 (\rho_2 c_2)} = \frac{4 \rho_1 c_1 \rho_2 c_2 \cos^2 \theta_i}{[\rho_2 c_2 \cos \theta_t + \rho_1 c_1 \cos \theta_i]^2} \quad (9)$$

The loss in decibels per reflection may be computed by using

$$\text{Loss in db} = 20 \log R \quad (10)$$

where R is the Rayleigh reflection coefficient.

When the velocity of sound in the bottom is greater than the velocity of sound in the ocean ($c_2 > c_1$) the amplitude of the reflected wave will increase from a finite value at normal incidence to a value of unity (perfect reflection) at the critical angle and remain at unity out to grazing incidence ($\theta_i = 90^\circ$). The critical angle is

$$\theta_c = \theta_i = \sin^{-1} c_1/c_2 \quad .$$

The most rapid rate of increase is immediately below the critical angle. The phase change is from 0 degrees at the critical angle to 180 degrees at grazing incidence ($\theta_i = 90^\circ$)

If the velocity of sound in the sediment is less than the velocity of sound in the ocean ($c_2 < c_1$) the reflection coefficient decreases from a finite value at normal incidence to 0 at an angle of incidence θ_c .

$$R = R_r = c_1 \eta^{-1} \left[\frac{1 - \left(\frac{c_1}{c_2}\right)^2}{\left(\frac{c_2}{c_1}\right)^2 - 1} \right]^{1/2}$$

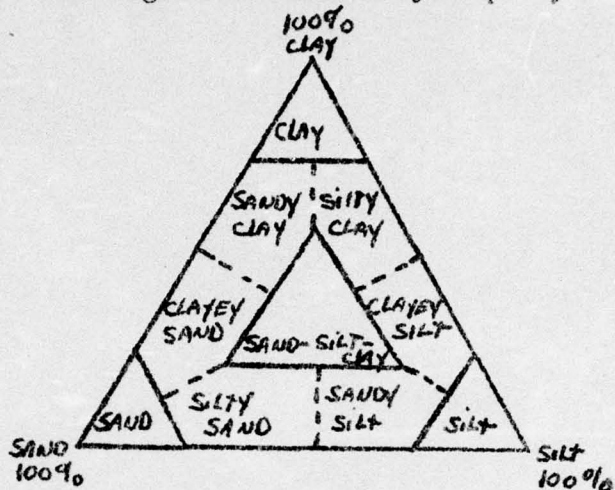
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At the angle of intramission, θ_c , all sound energy will be transmitted into the bottom. For $\theta_i > \theta_c$ the reflected wave is reversed in phase by 180 degrees and the reflection coefficient increases rapidly to a value of 1 at $\theta_i = 90^\circ$.

One of the major uncertainties in the application of the reflection coefficient is determining the in situ density of deep sea sediments. In situ sediment velocities may be determined by seismic methods whereas in situ sediment densities are determined in the laboratory. Sediment sound velocities are also calculated in the laboratory and agree quite well with seismic results. Serial measurements of water temperature and salinity are necessary for determining sound ray paths and water density.

Whenever acoustic and physical properties are discussed, sediments must be defined rather precisely. The term mud is useful as a field term or as a term on charts, but is too vague when exact determinations of physical properties are made. The nomenclature used in analyzing the sediments presented in this report is that recommended by Shepard in which a sediment name is derived by the

relative amounts of sand, silt, and clay present. The following illustration is the diagram recommended by Shepard;



The sediments within any one division may be further named according to the Wentworth grade scale of particle diameters.

coarse sand	1.00 - 0.50 mm
medium sand	0.50 - 0.25 mm
fine sand	0.25 - 1.125 mm
very fine sand	0.125 - 0.062 mm
silt	0.062 - 0.004 mm
clay	less than 0.004 mm

The density of a sediment is dependent on the densities of the solid constituents or mineral grains, the water in the pore spaces between the grains, the gas trapped or formed in the sediment, and on the relative amounts of these constituents.

A small number of common minerals compose most rocks. The important solid constituents of unconsolidated marine sediments are the mineral particles derived from these rocks, the mineral grains which form in place, and organic deposits. After the sediments are deposited on the ocean floor additional minerals may be formed

by chemical precipitation or by the alteration of other minerals.

In detrital mineral studies it is common to separate the minerals in a heavy liquid such as bromoform, so that the minerals which float are called the "light" minerals (less than about 2.89 g/cc), and those which sink are called the "heavy" minerals.

The light minerals, which make up about 98% of many sands, are mainly quartz (2.66 g/cc) and feldspars (2.57 to 2.77 g/cc). The dominance of light minerals results in an average grain density in most sands of about 2.65 g/cc.

If the sediment has an unusually large amount of heavy minerals the density of the solid constituents will be greater than the average of 2.65 g/cc. It has been found that the heavy minerals are concentrated in the finer sizes of sand, and that 70% of the heavy mineral grains have median diameters between 0.25 and 0.06 mm.

The density of the solid constituents in a sediment may be computed as follows:

$$\text{Density of solids} = \frac{\text{wet wt of sediment} - \text{wt of sea water}}{\text{total volume of sample} - \text{volume of sea water}}$$

The in situ density of shallow bottom water varies between 1.02 and 1.03 g/cc. The density of bottom water increases with depth to about 1.05 at 3,000 fathoms.

The porosity of a rock may be defined as the percentage of pore space in the total volume of the rock, i. e., the space not occupied by solid mineral matter. The porosity is the total pore space as contrasted with the effective or available pore space. The total pore space includes all interstices or voids and is larger than the

Table 1

Sediment data and computed bottom loss

Sediment type	Median grain diam (mm)	Porosity (%)	Wet Density (g/cc)	Velocity (ft/sec)	Impedance (10^3 g/cm ² sec) (ρc)	Reflection loss (db)
Medium sand	.284	39	1.99	5608	3.40	8.5
Medium sand	.270	59	1.50	6690	3.06	9.8
Medium sand	.259	42	1.95	5504	3.27	9.0
Medium sand	.250	43	1.94	5855	3.46	8.4
Fine sand	.176	43	1.96	5576	3.33	8.8
Sandy coarse silt	.0594	61	1.62	5576	2.76	10.9
Sandy coarse silt	.0563	57	1.77	5904	3.19	9.5
Sandy coarse silt	.054	65	1.60	4907	2.40	13.4
Coarse silt	.053	63	1.64	4953	2.47	12.8
Clayey sandy coarse silt	.046	66	1.56	4987	2.33	13.9
Medium silt	.0101	61	1.63	5543	2.76	11.3
Fine silt	.0094	65	1.60	5215	2.54	13.0
Fine silt	.0078	69	1.54	4953	2.33	14.8
Fine silt	.0071	61	1.62	5150	2.54	13.0
Fine silt	.0070	69	1.44	5248	2.28	15.3
Very fine silt	.0066	79	1.35	4884	2.01	17.6
Very fine silt	.0063	52	1.74	5510	2.92	10.5
Clayey very fine silt	.0058	76	1.41	4897	2.10	16.5
Clayey very fine silt	.0052	78	1.37	4887	2.04	18.7
Clayey very fine silt	.0050	—	1.43	4953	2.16	16.4

Clayey very fine silt	.0046	62	1.57	5248	2.51	12.8
Coarse clay	.0038	69	1.49	4986	2.27	15.4
Coarse clay	.0033	55	1.77	5510	2.97	10.3
Coarse clay	.0030	74	1.50	4986	2.28	15.5
Coarse clay	.0028	69	1.52	4952	2.27	15.4
Coarse clay	.0025	83	1.31	4874	1.95	18.5
Coarse clay	.0020	81	1.32	4887	1.97	20.0
Coarse clay	.0020	80	1.30	4986	1.98	20.0
Coarse clay	.0020	69	1.45	5248	2.32	15.3
Coarse clay	.0019	82	1.31	5051	2.2	15.5

effective pore space. In a water saturated sediment the void space is water filled.

Porosity may be divided into two classes: original and secondary. Original porosity is an inherent characteristic and was determined at the time the sediment was formed. Secondary porosity results from later changes, which may increase or decrease the original porosity.

The original porosity of the sediment is affected by (1) uniformity of grain size, (2) shape of grains, (3) method of deposition and manner of packing, and (4) compaction during **deposition**.

Theoretically, actual grain size has no influence on porosity. However, it has been found that the finer-grained sediments have a higher porosity than the coarse-grained sediments. It must be noted that size is related to other properties, such as shape which may be an influencing factor in porosity differences.

Sand grains, because of their large size, sink quickly to the bottom and as they are too large to be affected by the adsorbed

water on their surfaces or by intermolecular forces, they assume a position among the other grains on the bottom under the influence of gravity and bottom currents. Sand and coarse-grained silt-sized particles either assume a single grained or mixed grained structure. Whether the grain size is uniform or nonuniform is of fundamental importance. The highest porosity is obtained when the grains are all of the same size. It has been shown that equal-sized spheres, if regularly packed, will have porosities ranging from 26% to 48%. Naturally deposited sands of uniform size have porosities ranging from 23% to about 50%. If finer-grained particles are added to the uniform assemblage the porosity will be lowered. This is due to the interstices being occupied by the finer particles, thereby leaving less room for the interstitial water.

The finer-grained sediments generally possess greater porosity chiefly as a result of two factors. The finer-sized particles have films of water absorbed on their surface and are affected by intermolecular forces. When these particles fall to the bottom they will most likely stick to the grain on which they first fall. They will be held there by intermolecular forces and will form a three dimensional honey-comb structure. Smaller particles also tend to be less rounded than larger particles, a phenomenon which is known to produce an increase in the original porosity. These factors tend to increase the porosity of the finer-grained sediments.

Grain shape is another factor that will affect porosity. Laboratory studies show that minimum porosity is obtained with disc-shaped particles. These studies show that the addition of platy minerals, such as clay and mica, to sediments will increase the porosity considerably. Crushed mica has a maximum porosity of 92%.

The relationship between wet density ρ , grain density ρ_s , water density ρ_w , and the porosity expressed as a fraction, n , may be expressed as $\rho = (1-n)\rho_s + n\rho_w$. This is a straight line relationship in which wet density varies inversely with porosity between the extremes of no water and 100% water.

If the mean grain density of a sample is known it would then be possible to predict porosities for sediments of any grain size in a given area. The known density is plotted, at zero porosity, on a density versus porosity diagram (Fig 1). A line is then drawn between this known density and the density of sea water (1.025 at 100% porosity). The wet density of any similar sediment in the area should lie on, or close to, this line.

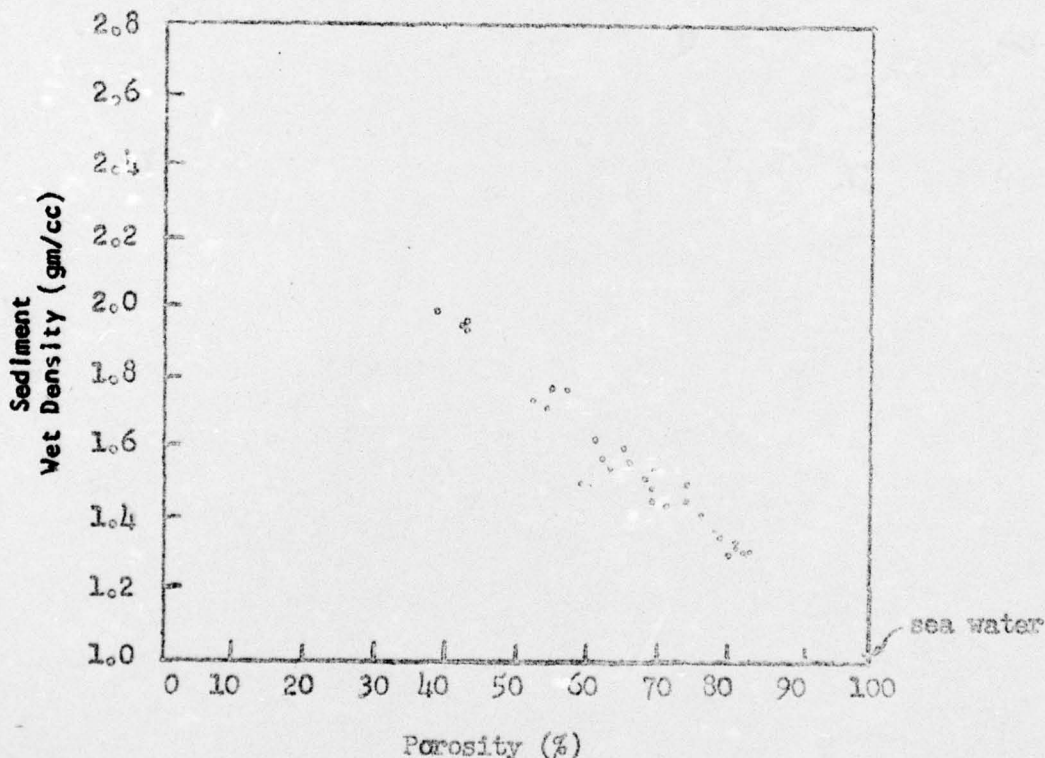


Figure 1 Sediment Wet Density vs Porosity

For sediments ranging from medium sand to coarse silt the porosity varied from 39 to 65% while the wet density range was from 1.50 to 1.99 gm/cc. The porosity range for the fine silts and clays is from

52 to 83 % while the wet density reaches a minimum at 1.30 gm/cc.

The velocity of sound in an ideally elastic homogeneous medium is

$$V_L = \left(\frac{[3(1-\sigma)k + \mu]}{\beta\rho} \right)^{1/2} = \left(\frac{k + \frac{4}{3}\mu}{\rho} \right)^{1/2} \quad (12)$$

where V_L = velocity of longitudinal elastic wave, k = bulk modulus, β = compressibility, ρ = density, μ = coefficient of rigidity, and σ = Poisson's ratio. When rigidity is lacking ($\mu = 0$), Eq. (12) reduces to

$$V_L = \left(\frac{k}{\rho} \right)^{1/2} = \left(\frac{1}{\beta\rho} \right)^{1/2} \quad (13)$$

Any increase in rigidity above zero, which is to be expected in sediments, would increase the velocity of sound as the value in the numerator is increased to greater than one.

Laboratory studies indicate that the compressibility of a suspension can be considered as an additive property of the individual compressibilities of the liquid and the suspended particles. A "bulk compressibility" of the aggregate may be obtained from known mineral compressibilities and relative abundances of the various constituent minerals. The bulk compressibility, β_s , of the wet sediment may be determined by using

$$\beta = (1-n)\beta_s + n\beta_w \quad (14)$$

where n = fractional porosity, β_s = bulk mineral aggregate compressibility, and β_w = water compressibility.

Equation (13), which applies to fluids and suspension, may be used to determine approximate compressibilities for sediments with porosities greater than about 65%.

Equation (13) does not hold for sediments with median diameters in the fine and very fine sand range because of the grain-to-grain contact which causes a more rigid structure. Equation (14) gives some idea of the compressibility values for these sediments.

An examination of the data shows that the velocity of sound in the bottom sediments, at various stations, is less than or near **the velocity of sound in the water just above the bottom.** Laboratory studies have provided a theoretical explanation for the low velocities in the fine-grained highly porous sediments.

These sediments are lacking a rigidelastic structure and the sediment particles are largely free to move in the sound field. The velocity behavior will follow that of a suspension. It appears that when the mineral concentration reaches 23% (77% porosity) in a fine grained sediment, a structure is assumed in which the grain - to - grain contacts cause the sediment to assume elastic properties which were not present in the suspension. At this point the sediment assumes a measurable rigidity, its Poisson's ratio decreases from 0.50 for fluids, and its compressibility becomes less than that of a fluid mixture. The elastic properties, previously small or absent, now become important in the computation of the velocity of sound and this velocity increases. Laboratory measurements indicate that there are many fine grained sediments with a porosity greater than about 60% which possess a loose structure and can be considered to conform approximately to the behavior of a suspension.

A plot of velocity vs porosity shows an increase in velocity with a decrease in porosity(Fig 2).

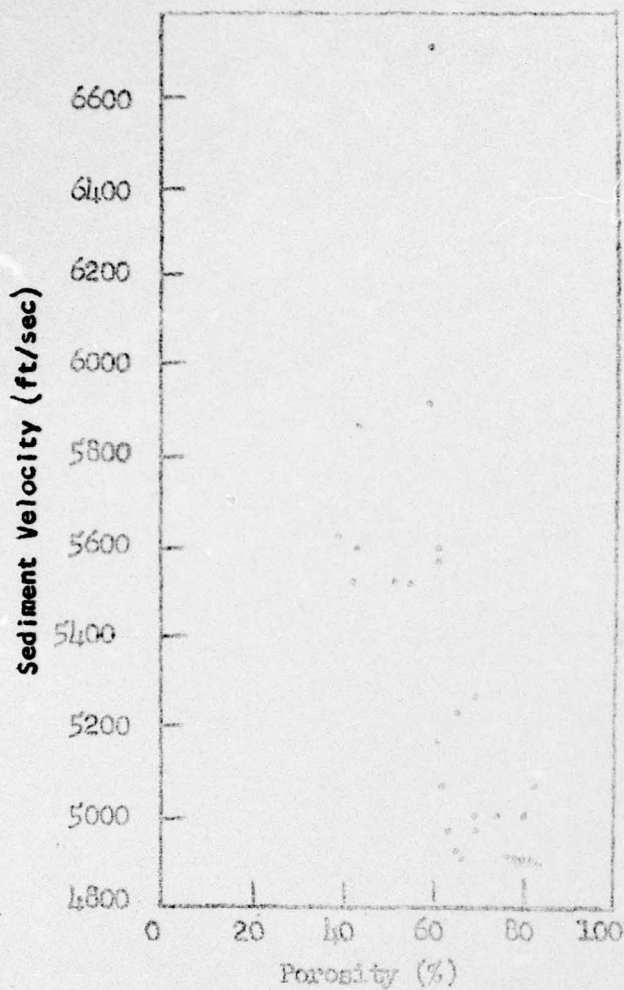


Figure 2 Sediment Velocity vs Porosity.

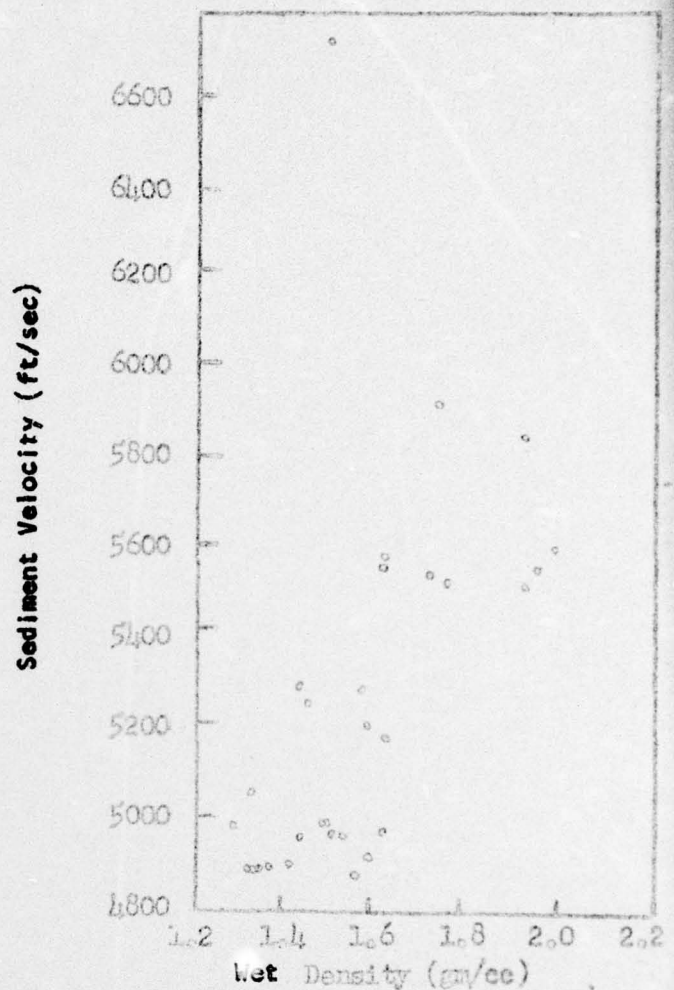


Figure 3 Sediment Velocity vs Wet Density.

The velocity vs wet density relationship (Fig 3) shows an increase of velocity with wet density. A plot of velocity versus median diameter shows a general increase of velocity with increasing median grain size (Fig 4).

The relationship between porosity and grain size was discussed previously. Figure 5 is a plot of porosity vs median diameter and indicates an increase in porosity with a decrease in grain size.

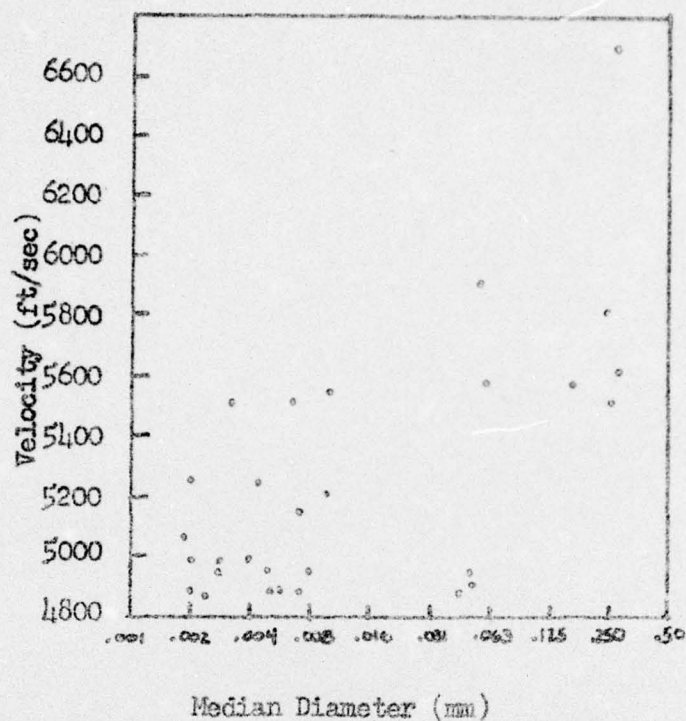


Figure 4 Sound Velocity vs Sediment Median Diameter.

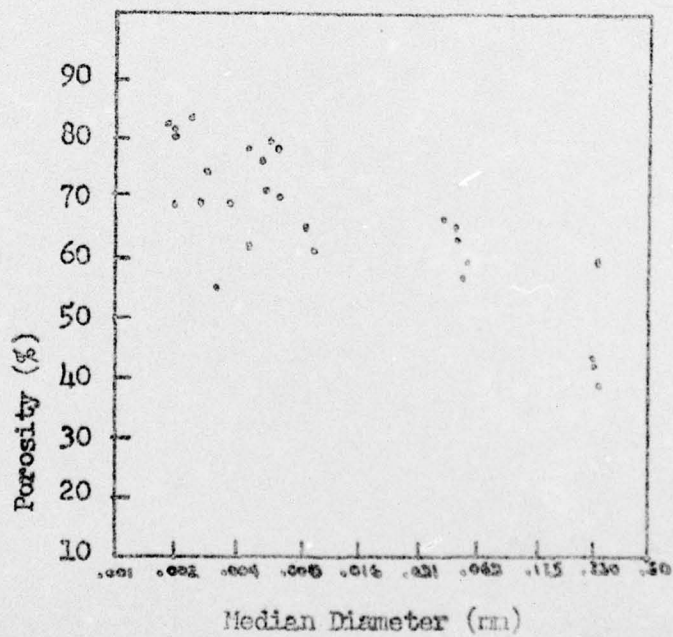


Figure 5 Porosity vs Median Diameter.

The impedance (density times velocity) for the sediments has been computed and is presented on page 7. Since both density and velocity decrease as porosity increases (Figs 1 & 2), the impedance decreases also (Fig 6).

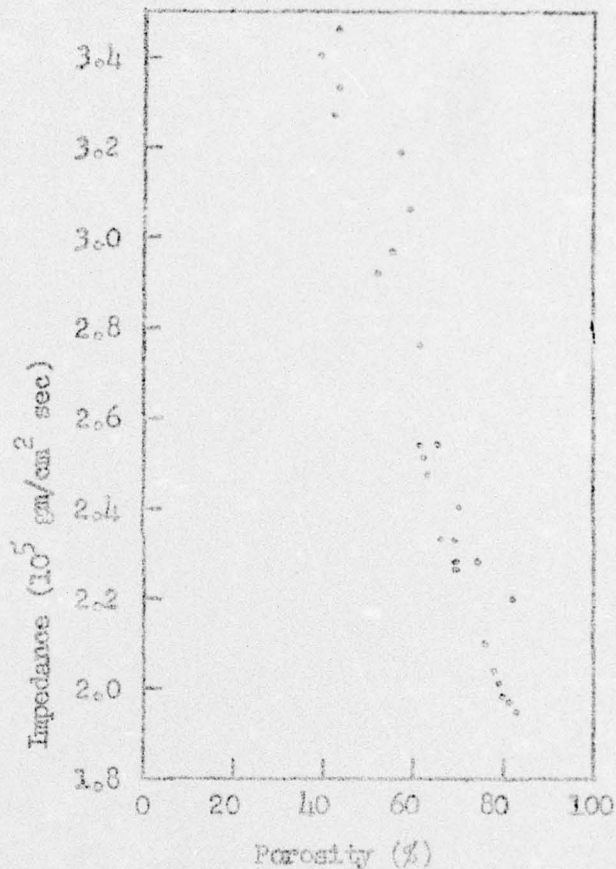


Figure 6 Impedance (Density x Velocity) vs Porosity.

Figure 7 is a plot of computed reflection loss (db) at normal incidence vs porosity. From this illustration it can be seen that as porosity increases the reflection loss increases. This relationship would be expected as a result of the dependence of reflection loss on velocity and density.

A plot of reflection loss at normal incidence vs impedance shows reflection loss decreases as impedance increases. This is the expected relationship (Fig 8).

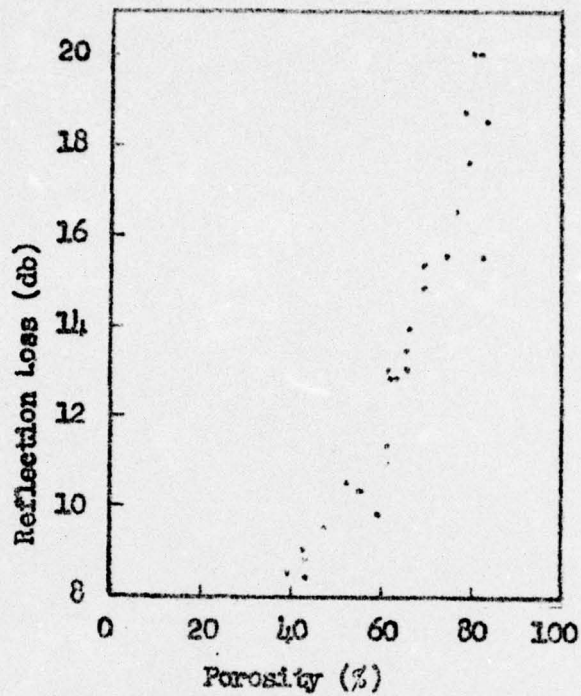


Figure 7. Computed Reflection Loss (db) at Normal Incidence vs Porosity.

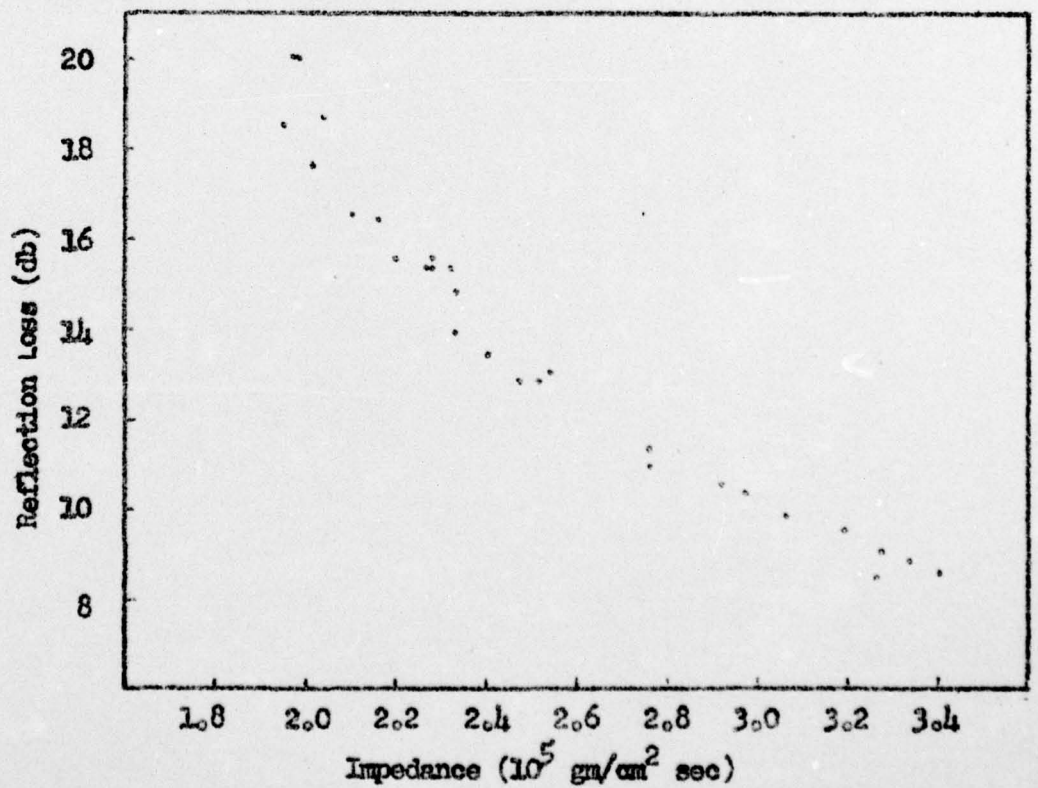
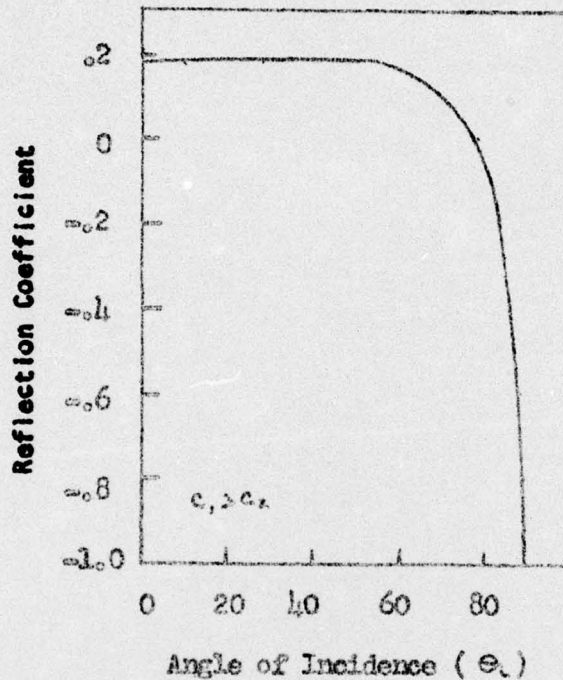


Figure 8 Reflection Loss at Normal Incidence vs Impedance.

As mentioned previously, the velocity of sound in the bottom sediments may either be greater than or less than the velocity of sound in the water overlying the bottom. Figures 9 and 10 are plots of calculated values of the reflection coefficient vs angle of incidence for $c_1 > c_2$ and $c_1 < c_2$.



$$c_1/c_2 = 1.019$$

$$\rho_2/\rho_1 = 1.0448$$

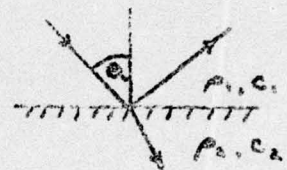
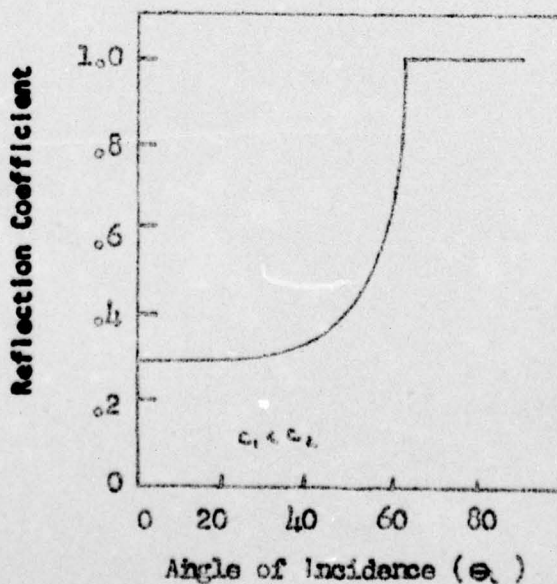


Figure 9 Rayleigh Reflection Coefficient vs Angle of Incidence.



$$c_1/c_2 = 0.885$$

$$\rho_2/\rho_1 = 1.558$$



Figure 10 Rayleigh Reflection Coefficient vs Angle of Incidence.

By using Equation (10) it is possible to compute the reflection loss from the reflection coefficient. Figures 11 and 12 are plots of reflection loss vs angle of incidence. Figures 9 and 11 ($c_1 > c_2$) show that as the angle of incidence increases the reflection coefficient decreases to zero at the angle of intromission and the reflection loss increases until all of the sound energy is transmitted into the bottom. At angles of incidence greater than the angle of intromission the reflection loss rapidly decreases to zero at $\theta_i = 90$.

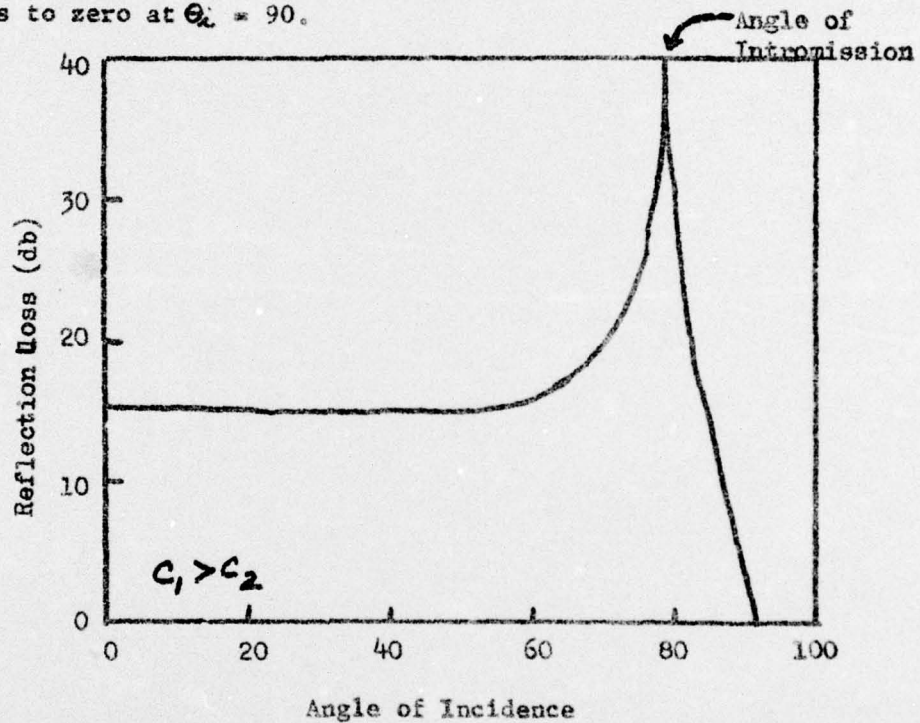


Figure 11 Reflection Loss vs Angle of Incidence.

When $c_1 < c_2$ the reflection loss will decrease as the angle of incidence increases. At the critical angle and beyond, all of the sound energy will be reflected (Fig 12).

A comparison of Figures 11 and 12 shows that high reflection

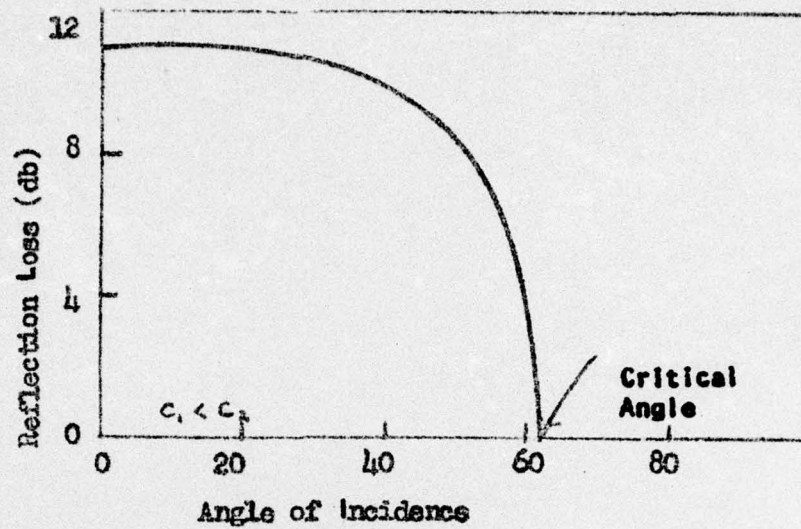


Figure 12 Reflection Loss vs Angle of Incidence.

losses are obtained in those sediments in which the velocity of sound is less than the velocity of sound in the water. The average computed reflection loss, at normal incidence, was found to be about 9 db for the fine and medium sands, 13 db for the silts, and 16 db for the clays.

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