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AN EXPERIMENTAL STUDY OF COMPRESSIONAL VELOCITIES IN DEEP SEA SEDIMENTS

by

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SUBMITTED IN PARTIAL FULFILLMENT

OF THE REQUIREMENTS FOR THE

DEGREES OF BACHELOR AND

MASTER OF SCIENCE

at the

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

June, 1960

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Department of Geology and Geophysics, May 20, 1960

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ABSTRACT

by

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Submitted to the Department of Geology and Geophysics on May 20, 1960, in partial fulfillment of the requirements for the degrees of Bachelor and Master of Science.

A 560 kilocycle acoustic probe has been used to measure several hundred compressional velocities in three deep sea cores from the Mediterranean Sea. Velocity-depth profiles as determined with the probes have been compared with moderate positive correlation to subbottom reflections recorded with a high resolution echo sounder. The spread of velocity measurements is from 1.47 to 1.63 kilometers per second. These values are in good agreement with results given by previous investigators. An accuracy of .03 km/sec., or 2% of the measured velocity, was attained with the probes. Many clays were observed with velocities less than that of sea water. This has been predicted by Urick (1947). Compressional velocity has a strong inverse dependence upon water content (or porosity). From the results of eleven specific gravity determinations it is concluded that specific gravities of the clay solids do not vary by more than a few per cent from a value of 2.80. Therefore the high velocity correlation is associated with the bulk density of the sediment. No velocity-depth relationship was found. Lithologic changes that involve differences in water content have a much greater influence on velocity than any depth relationship over the depths of 25 feet here considered . The liquid limits of sediments from the cores consistently showed values a few per cent less than the natural water content. This is taken as evidence that no large variations in water content occurred between the times the cores were taken and when they were examined. Visual evidence and data from a consolidation test is used to indicate intense deformation in one core.

Thesis Supervisor: J. B. Hersey

Title: Associate Professor of Oceanography

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SYMBOLS AND SUBSCRIPTS

Symbols

B = bulk momulus = $-\frac{1}{V}\left(\frac{\partial V}{\partial P}\right)_{\substack{\text{constant}\\ \text{entropy}}}$ = adiabatic compressibility

B_c = structural bulk modulus

d = diameter

e = void ratio = volume voids/volume solids

e_o = initial void ratio

f = frequency

f_c = critical frequency

G = specific gravity of soil

K = $1/B$ = adiabatic incompressibility

L = length

L.L. = liquid limit

n = porosity = $e/(1+e)$

P = pressure

P.I. = plastic index

t = time

T = temperature in °C.

u = shear, or rigidity, modulus

v_c = compressional velocity

w = water content = Weight water/Weight solid

W = weight

ρ = density

ϕ_i = volume fraction of component i

Subscripts

s solid

l liquid (usually water)

m the composite medium under consideration

ACKNOWLEDGMENTS

The writer wishes to thank Doctor J. B. Hersey of Woods Hole Oceanographic Institution, who acted as my thesis advisor during this investigation. Other individuals at the Woods Hole Oceanographic Institution also contributed in collecting the samples and in furnishing advice.

Several individuals at the Massachusetts Institute of Technology Soil Mechanics Laboratory, especially Professor J. E. Roberts, were very helpful in ideas, in locating a place for me to work, and in providing some of the materials used in this study.

My thanks to Ann C. Henry for typing the manuscript.

AN EXPERIMENTAL STUDY OF COMPRESSIONAL VELOCITIES IN DEEP SEA SEDIMENTS

I. Introduction

During the CHAIN 7 cruise (Woods Hole Oceanographic Institution) in June 1959, Doctor J. B. Hersey suggested to me that a study of acoustic velocities in deep sea cores might prove most interesting and informative. A comparison of velocity-depth profiles from reflection data and from measurements on the cores themselves was suggested.

Work on this project was begun on the ship during the summer of 1959 and then continued at the M. I. T. Soil Mechanics Laboratory from September 1959, to April 1960.

This problem is part of a general one of securing measurements of physical properties of materials below the ocean bottom. Only a few tens of meters of the deep sea bottom have been sampled. Below these depths only acoustic data are presently available. Correlations between seismic information and other physical properties are thus desired. Such correlations might be extended to greater depths through laboratory experimentation on consolidated bottom samples.

Acoustic data on bottom sediments are relatively easy to take compared to sampling. In some instances a continuous reflection profile of the upper layers of sediment can be made. However, acoustic velocities in these layers are not obtained with continuous profiling devices such as the high resolution echo sounder or a sparking source. Velocity determinations are usually accomplished with an explosive source. Nevertheless, the velocities of layers just below the bottom are often unobtainable with the conventional explosive methods. A low velocity layer, which has a velocity sometimes less than that of the water above, is often either masked or not resolvable without the use of a two ship, wide-angle reflection profile.

An attempt is described in this paper to measure velocities in materials from this low velocity, high porosity layer. Velocity-depth profiles for three cores approximately 25 feet long have been plotted from the experimental data. This is compared with reflection data taken with a high resolution echo sounder. Other tests of physical properties are as follows:

Water content	Specific gravity of solids
Atterberg limits of sediments	Consolidation
Grain size	Visual description
Mineralogy	Extent of disturbance.

II. Some Physical Properties of Deep Sea Sediments

Sediments exhibit a wide range of physical properties. In size they range from suspensions and colloids to sands and lithified sediments. Many mechanisms are in effect through this range. Consequently, a theory of strength and wave propagation applicable to all of these would be complex and difficult to manage even if it did exist. Some simple theoretical models of moderate applicability to high porosity sediments like clays and silts are discussed in section III.

The comparison of velocities and other physical properties of sediments can be difficult in several respects. Equations relating velocity and elastic constants are based on such assumptions as elastic deformation in a homogeneous medium. While propagation of low amplitude acoustic waves in sediments is probably in the linear region, other strength tests involve extremely non-linear macroscopic processes such as rearrangement of structure and the sliding of grains past one another. Strength moduli as determined by these two methods would not be expected to be comparable since they do not represent the same physical processes.

At the very least the physical properties of sediments are dependent on the constituents — solid, water, and possibly air or gas. Air can be neglected in the case of saturated sediments. However, dissolved gases can affect physical properties, especially in sediments with a high organic content.

In most sediments the concentration of mineral matter in the liquid is high enough so that considerable physical and electrical interaction takes place. For this reason grain size alone is not the great controlling factor of physical properties as early soil workers had thought it might be. This is especially true in clays where the terms "grain size", "pore size", and "grain-to-grain contact" are not very applicable. In sands, however, such terms have much greater physical meaning.

Theories of wave propagation in unconsolidated sediments include considerations of bulk composition, grain contact, and grain size to some extent. No attention has been given to electrical or electrolytic processes as they might affect wave propagation. Relative differences in velocity in sediments of high porosity are largely a function of water content. This is so because the compressibility of water is large compared to that of the minerals involved. However, any theory of attenuation or dispersion in unconsolidated sediments will likely require a consideration of the sediment structure and electrical interactions of small particles. These effects should become more important as the sediment is consolidated.

Differences in acoustic properties between sands and clays must include not only a consideration of grain size but also of the adsorbing nature of the clay particles. Many of the physical properties of clays can be accounted for in terms of changes in the adsorbed water layer around clay particles. Water content of these clays is thus a function of forces acting between the clay particle and its surrounding envelope of water. Salt concentration can affect this water envelope drastically by neutralizing charges on the particle's surfaces and thus decreasing the amount of adsorbed water. Particles in a non-salt floc tend to line up parallel to one another, whereas a salt floc has no directional properties. In the sense that electrolytic concentration can affect the structure, it should have some effect on wave propagation. To my knowledge no experimental information on this is available. Such an effect on acoustic properties would likely be small at high porosities. Further considerations of clay structure and strength can be found in Lambe (1958).

III. A Review of the Literature — Some Theoretical Models for Acoustic Propagation in Sediments

Several simple models for composite media have been discussed in the literature. Good reviews of the subject have been given by Laughton (1957) and Shumway (1960).

Urick (1947) assumed dilute solutions of suspended sediment can be approximated by an ideal solution such that

$$\text{Eqn. (1)} \qquad V_m = V_1 + V_s$$

where V_m, V_1, and V_s stand for volumes of the composite mixture, liquid, and solid respectively. For particles smaller than the wavelength of sound passed through the medium, Urick assumed scattering to be negligible. In this case particles are not in contact. Therefore all components will be subjected to the same pressure P. An expression can be derived for the compressibility of the mixture. All compressibilities are understood to be adiabatic.

$$\text{Eqn. (2)} \quad B_m = -\frac{1}{V}\left(\frac{\partial V}{\partial P}\right) = -\frac{1}{V_1 + V_s}\left[\left(\frac{\partial V_1}{\partial P}\right) + \left(\frac{\partial V_s}{\partial P}\right)\right]$$

$$\text{Eqn. (3)} \quad B_m = \phi_s B_s + \phi_1 B_1 = \phi_s B_s + B_1 (1 - \phi_s)$$

where ϕ_i = volume fraction of component i

This compressibility relationship was assumed by Urick (1947). The above derivation is similar to that given by Laughton (1957). In like manner

$$\text{Eqn. (4)} \quad \rho_m = \rho_s \phi_s + \rho_1 (1 - \phi_s)$$

An expression for compressional velocity in the mixture is as follows

$$\text{Eqn. (5)} \qquad v_{cm}^{2} = 1 / \rho_m B_m$$

If the following substitutions are made in Eqn. (5)

$$g = \frac{v_{cl}}{v_{cm}} \; ; \; a = (\rho_s/\rho_1) - 1 \; ; \; b = (B_s/B_1) - 1$$

It follows that

$$\text{Eqn. (5')} \qquad g^2 = (1 + a\phi_s)\ (1 + b\phi_s)$$

A maximum or minimum in $g(\phi_s)$ will occur for

$$\text{Eqn. (6)} \qquad \phi_s = - (a + b)/2ab$$

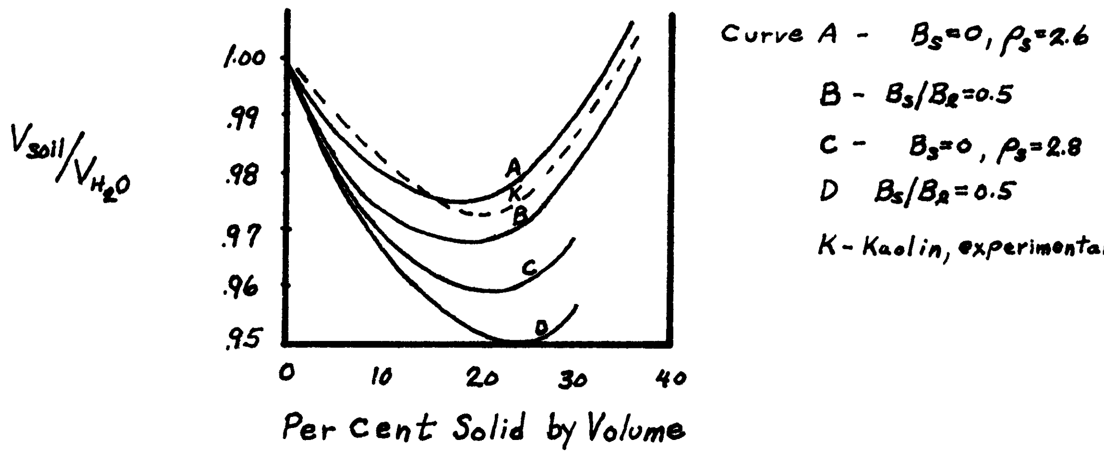
$$\text{Eqn. (7)} \qquad g^2_{max} = - (a - b)^2/4ab$$

For sediment particles in water ρ_s is greater than 2 gm/cm^3; $\rho_1 = 1$ gm/cm^3. "a" is therefore greater than 1. For a real solution of "g", it is required "b" be less than zero; i.e., B_1 is greater than B_s. The requirement $0 \leq \phi_s \leq 1$ imposes the additional restriction $a > |b|$. This is satisfied in the case of sediments where "a" is greater than 1.

Urick plotted two theoretical curves: one for $B_s = 0$, the other for $B_s/B_1 = .05$. The dashed curve in Figure 1 is Urick's experimental determination for a kaolinite-water mixture. If a value of $\rho_s = 2.80$ gm/cm^3, approximately that obtained in this paper, the curves C and D are obtained. This value of density and that used by Urick (2.60 gm/cm^3) should span the normal range of densities in sediments excluding light pyroclastics.

Urick (1948) considered the case of small, rigid, incompressible

FIGURE 1: Low Velocities in System Soil-Water after Urick (1947)



Low Frequency Behavior of Velocities after Biot (1956)

Figure 2

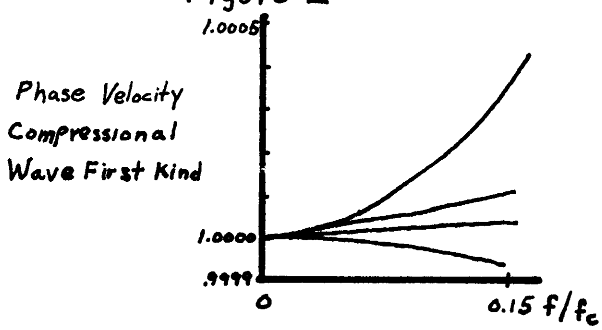
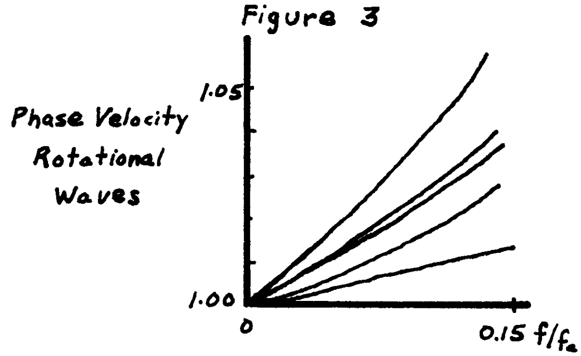


Figure 3



Full Frequency Behavior of Velocities - after Biot (1956)

Figure 4

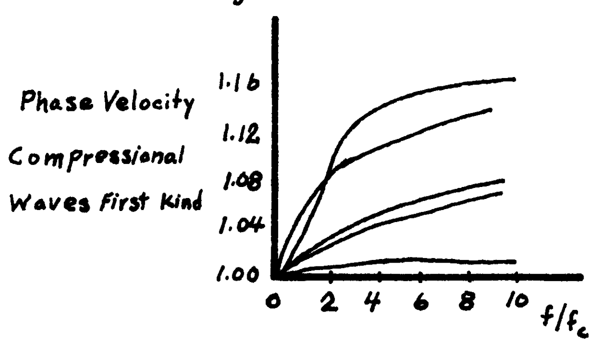
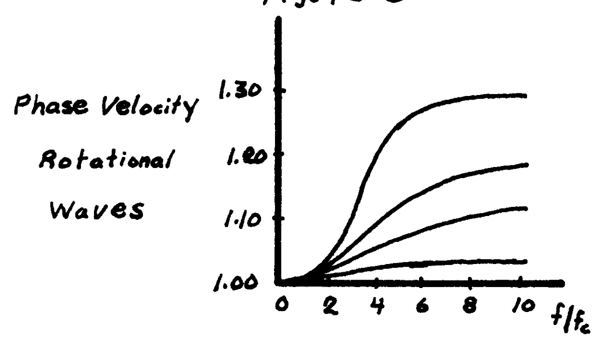


Figure 5



spherical particles free to move in a liquid at low frequencies. Urick derived an absorption coefficient from the mechanics of a spherical pendulum moving in a viscous fluid. Coupling between the components depends, of course, on the driving frequency and the viscosity of the fluid.

Urick and Ament (1949) found a complex propagation constant for composite media. In addition to Urick's 1947 theory they added the effect of scattering. The resulting complex propagation constant has a velocity term in the real part and an attenuation factor in the imaginary part. Velocities and absorptions were found in agreement with the theory for concentrations of solid less than 25% by volume.

Ament (1953) carried the 1949 theory a step further by determining a dynamic density $\rho_m(f)$, a dynamic compressibility $B_m(f)$, and hence, a complex propagation constant dependent on frequency.

Chambré (1954) developed a plane wave equation relating density and strain in a composite medium. From this he derived the propagation term term

$$\text{Eqn. (8)} \qquad v_{cm}^2 = \left(\frac{\partial P}{\partial \rho_m} \right)$$

From this Chambré also derived equation (5). He showed that it also follows from considerations of conservation of mass and momentum in the mixture.

Biot (1956) has developed a more general theory for elastic waves in porous solids. His theory treats the case of a porous, compressible solid saturated by a viscous, compressible liquid. Still his theory must assume such things as impervious pore walls and pore sizes close to a mean. Biot has defined a characteristic frequency f_c dependent upon the dynamic viscosity of the fluid and upon the pore diameter. Fluid flow at frequencies much less than f_c will be laminar. However, Biot has considered the full frequency behavior.

For water at 15°C. and assuming the pores are circular in cross section, the following table gives a general idea of the values of f_c to be expected for a given pore diameter. I have equated pore diameter and grain size in Table I as a basis for comparing sediments. Particle diameter squared is inversely proportional to f_c.

Table I

Characteristic Frequency	Pore Diameter	M.I.T. Classification
.65 kc/sec.	0.1 mm	fine sand
65 kc/sec.	.01 mm	medium silt
6.5 mc/sec.	.001 mm	coarse clay

Biot defined a structural factor dependent on pore shape. He concluded there was but little difference between elastic properties for the limiting cases of a long slit and a sphere.

Biot's theory predicts that a rotational and two compressional waves should be expected in porous solids. Each of these waves involves coupled motion between the fluid and solid. That compressional wave which Biot has referred to as the first kind is the fastest of the three; the other compressional wave is highly damped. Paterson (1956) has observed the two compressional waves experimentally in sands. Figures 2-5 are sketches of the frequency dependence of velocities for several different degrees of coupling as given by Biot. Biot predicted an absorption coefficient proportional to f^2 for the cases considered in Figures 2-5. One difficulty with this more expanded treatment is that Biot has introduced many factors which are quite difficult to measure or evaluate.

As seen in Figure 2 little dispersion of phase velocity of compressional waves is expected for frequencies less than f_c. A theory of dispersion is also contained in the Ament (1953) treatment. Laughton (1957) examined the latter theory and concluded that a frequency of 500 mc would be needed to change v_c by 10% from its value at zero frequency. Laughton has criticized the calculation of dispersion based on Ament's theory as presented by Officer (1955). Officer's calculation of dispersion is dependent on values of permeability squared. Officer used values of permeability corresponding to the mean grain size of the sediment. Actually permeability is controlled by the finest ten to fifteen per cent of the sediment. Values of permeability and dispersion smaller than those given by Officer are thus expected. However, the question is still not settled. The Ament theory deals with a dynamic and not a static permeability. Permeability values used in dispersion calculations have been based on the static permeability. Bound water and structure of clays should be important in limiting the effective channel or pore sizes of the water.

Sutton, Berckhemer, and Nafe (1957) have examined predictions of dispersion in the Ament and Biot theories. They concluded that dispersion should be negligible for frequencies below 10 mc for most sediments. Since high frequency transducer measurements on sediments have yielded minimum values of velocity similar to those obtained from seismic data at low frequencies, Sutton, et al., concluded that this is another indication that dispersion is small. However, Officer (1955), reported a seismic study wherein velocity increased with frequency. A possible resolution of these difficulties is that the dispersion noted by Officer may arise from the geometry of the experiment and not from the sedimentary material itself.

Laughton (1957) has examined several elastic models for porous media. When particles immersed in a liquid are not in contact, equation (3) is the valid expression for the compressibility of the mixture. If the particles are in contact in an infinitely compressible fluid, all applied stress will be borne by the particle structure. The most general case will be one intermediate to these two wherein the particles are in contact in a fluid of finite compressibility. In this case an externally applied stress will be balanced by both hydrostatic and inter-particle forces. For this case Laughton derived the following expression

$$\text{Eqn. (9)} \qquad K_m^* = 1/B_m^* = 1/B_m + 1/B_c$$

where B_m^* = total compressibility of the mixture with particles in contact

B_m = compressibility derived in Eqn. (3).

B_c = a structural modulus.

Laughton has shown that an increase of velocity with pressure in all but the most advanced stages (where a shear strength is developed) can be accounted for by an increase in the incompressibility of the particle structure. The total expression for compressional velocity with a shear modulus u and a structural modulus included becomes

$$\text{Eqn. (10)} \qquad v_c^2 = \frac{K_m^* + 4/3\ u}{\rho\ m}$$

Several theoretical studies of elasticity applicable to sands have been made. Among these are works by Iida (1939), Gassman (1951), Morse (1952), Brandt (1955), and Paterson (1956).

IV. A Review of the Literature— Experimental Studies of Velocity in Sediments

The large porosity dependence of acoustic velocity is the subject of estimates of porosity at depth in the ocean bottom by Nafe and Drake (1957). From an inspection of seismic records for the North Atlantic, Nafe and Drake recognized a higher velocity gradient with depth in shelf areas than in deep sea regions. The inference is thus that the low velocities represent higher porosities. Some possible explanations (some of my own) of this difference in porosity are listed below.

1. Lithification is inhibited to a greater extent in the deep sea bottom than in shelf areas.
2. Different rates of loading are responsible.
3. Dissimilar structural arrangements of particles result from different methods of deposition.
4. Textural differences may account for at least a part of this effect. Shelf areas should be more abundant in coarser grained, low porosity materials.

Hamilton (1959) has recently discussed consolidation and thickness of deep sea sediments in the light of both geophysical data and consolidation tests on rocks and soils. This work is in part an extension of the theories of Nafe and Drake. Hamilton pointed out that the velocities down to several kilometers depth as interpreted from geophysical evidence are consistent with data on consolidation and lithification of sediments. This view is more in accord with the presumed rates of sediment accumulation than with the hypothesis that less than a kilometer of sediment is present in the average oceanic section.

Laughton (1957) mounted 1 megacycle compression and shear transducers in a consolidometer with one-dimensional drainage. In this manner he was able to measure void ratios and velocities of unlithified sediments in the pressure range 10 to 1000 kg/cm^2.

Laughton could vary the hydrostatic pore pressure independently. He found no velocity variations greater than 3% for hydrostatic pressures up to 400 kg/cm^2. The velocity increase from 0 to 5 kg/cm^2 pore pressure was the same for clay (terrigenous mud) as for pure water. Further increases of this pressure did not increase the velocity at the same rate. The change in compressibility of the water fraction at low pressures was responsible, presumably, for the velocity increase.

Globigerina ooze showed a decrease followed by an increase of velocity with increasing pore pressure. The increase is explained as for the clay. The decrease for the ooze is an indication grain-to-grain contact was being decreased by increasing pore pressure. Apparently the effect of grain contact is so small in the clay that the compressibility of water predominates. Since the clay has only fine grains (1-2 microns) while the ooze has a coarse fraction up to 100 microns, the ooze might develop more grain contacts than the clay where water around particles prevents grain contact as such.

Laughton observed velocity differences as porosity varied. However, several of his tests showed that all velocity changes could not be traced to porosity alone.

In the consolidation test, loads are added incrementally to achieve

Figure 6

Schematic of Velocity Change with Pore Pressure after Laughton (1957)

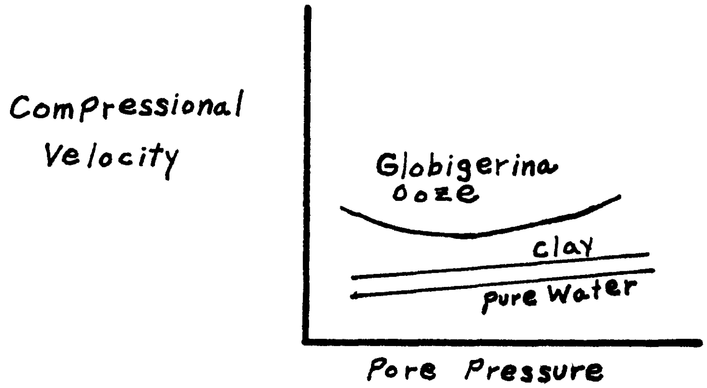


Figure 7

Velocity Variation With External Pressure during Consolidation after Laughton (1957)

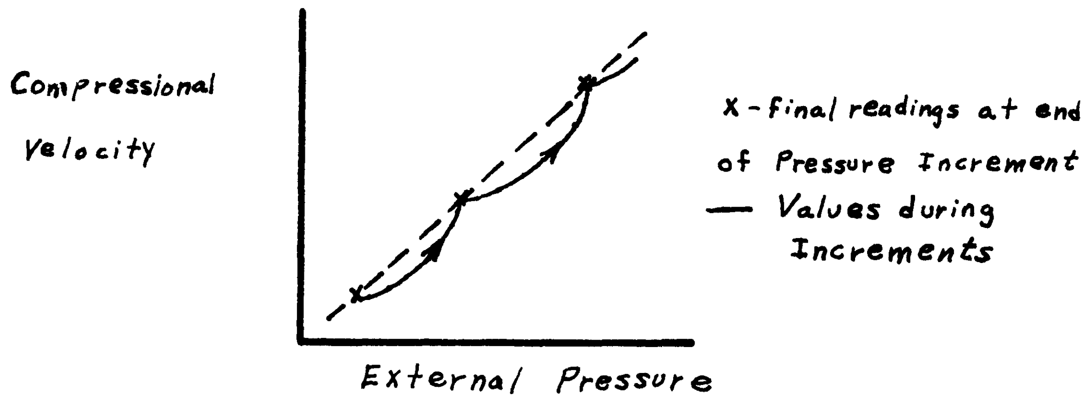
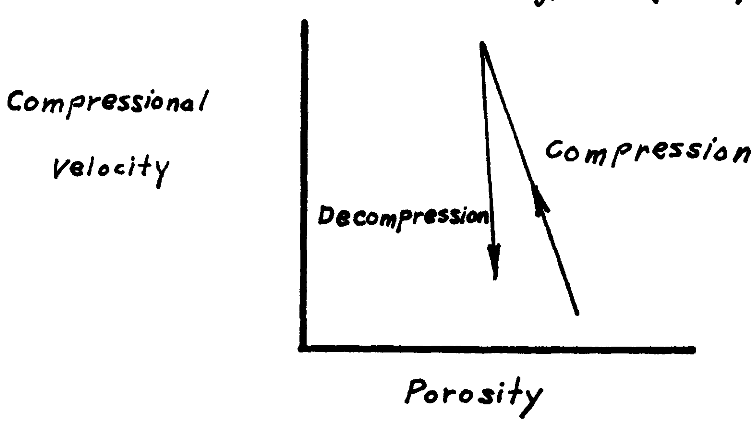


Figure 8

Velocity Variation during a Consolidation Cycle after Laughton (1957)



the highest pressure in the test. The first effect of the load is to put a stress on the pore water. In a drained test this pressure is relieved in a few minutes by drainage of free water from the soil. A secondary compression follows this. Water is slowly squeezed from the sediment structure. This may last for a period of years in a clay, but is practically non-existent in a sand.

Laughton noticed an increase of velocity during secondary compression. However, all values of velocity measured during a given pressure increment lay below the straight line joining the final values for each increment. (See Figure 7).

Laughton observed (Figure 8) that velocity increased much less than porosity decreased for a complete compression-decompression cycle. Hence velocity is not a function of porosity entirely.

Laughton hypothesized that velocity increases are primarily due to the establishment of a particle structure as pressure is increased. His development of a structural bulk modulus is given in the section on theoretical models. Laughton believes time effects may be of importance in that secondary compression, and hence velocity, may be much greater in nature for a given external pressure than for determinations in the laboratory under fast loading. Chemical reactions, especially those in carbonates, might be effective during geological times.

The velocity-applied pressure curve given by Laughton shows no indication of a decrease in slope at higher pressures.

During the compression Laughton observed no shear arrivals at all in the terrigenous mud and none in the Globigerina ooze until pressures greater than 500 kg/cm^2 were reached. Shear waves sent through the soils were partially converted to compressional arrivals which came in ahead of the shear waves. On the compressional cycle of the tests, shear waves could not be identified positively until they reached sufficient amplitude to show up in the p arrival. On decompression Laughton could follow the shear arrival down to 64 kg/cm^2. Difficulties can arise in coupling the shear generator to the sediment.

Hughes and Jones (1950) reported that shear arrivals could not be identified for most rocks for pressures less than approximately 35 kg/cm^2. This absence of shear waves in rocks at low pressures comes from air spaces and minute cracks that are closed with pressure in rocks. In the case of saturated sediments the water may play a similar role to the air in rocks, although cracks should not be present in the former.

Laughton also made velocity measurements in directions parallel and transverse to the main axis of the core. The transverse measurements (parallel to the earth's surface) always gave the higher velocity. Some values obtained are listed in Table II.

Table II

Transverse and Longitudinal Velocities in Soils and Sediments

Transverse/Longitudinal Velocity	Material Tested	Investigator
1.05	Globigerina ooze	Laughton (1957
1.13, 1.22, 1.36, 1.43	Terrigenous mud	Laughton
1.34	Red clay	Laughton
1.05	London clay	Ward, Samuels, and Butler (1959)

Sutton, Berckhemer, and Nafe (1957) made high frequency measurements on 26 cores both immediately after coring and three months later. With only one exception they observed no systematic change of velocity during this interval. The various cores represented most of the typical deep sea sediments and various geological ages. Most of these measurements were made every foot along the core. From a statistical analysis, Sutton, et al. derived the following formula for compressional velocity in bottom sediments.

$$\text{Eqn. (11)} \qquad v_c = 2.093 \text{ Km/sec.} + (.0414 \pm .0060)\log_2 M + (.00135 \pm .00038)j - (.44 \pm .15)n$$

where n = porosity; M = median grain size in mm.; j = per cent of material soluble in HCl. A large number of samples had a velocity less than that of water as predicted by Urick (1947). Higher attenuation of the acoustic wave was noticed in the coarser grained sediments. This observation has been reported by several investigators.

Neither Sutton, *et al.*, nor Laughton observed any sizable increase of velocity with depth. In only a few cases were increases observed. Hamilton (1959) states that porosity variations with depth in cores are not observed.

Sutton, *et al.* attempted to account for higher velocities in coarser grained materials by the statement that coarse grains are approximately round, while clay grains are platy. More grain-to-grain contact per unit volume is made in the coarser materials; hence, lower velocities are observed in fine sediments. However, the question should be raised about the meaning of grain-to-grain contact in fine sediments. A more reasonable answer to the question of decreasing velocity with increasing fineness at constant porosity would be that the clay particles surround themselves by a water envelope, whereas sands, for example, do not. The difference in velocities is not a result of grain shape *per se*.

Sutton, *et al.* calculated that differences resulting from temperature and pressure variations between their velocity measurements and those encountered in the sea bottom are smaller than one per cent and of the same order as their instrumental error.

Hamilton, *et al.* (1956) used two velocity techniques on a series of samples collected in San Diego harbor. In one case divers inserted transducer probes in the bottom. Sediment samples were taken at these stations. A resonant chamber technique (25-35 kc) as described by Toulis (1956) was used for velocity and absorption studies on the samples. The standard deviations for the probe *in situ* method and the resonance chamber were .035 and .015 km/sec. respectively. A correlation coefficient of

0.9 and a mean difference of .006 km/sec. were calculated. Sediment velocities are indicated in Table III.

Shumway (1960) has performed measurements with the resonance chamber on a wider range of unconsolidated marine sediments ranging from shallow water sands to deep sea clays. Over one-third of his values are less than the velocity of sea water.

Hamilton (1956) has analyzed the low velocity data obtained by Hamilton, et al. (1956). He found that fine grained sediments departed from the Urick (1947) theory for porosities less than 77%. He reasoned that grain contact was in effect below that limit. Hamilton found the Urick theory did not hold for highly porous sands. Sands are not suspensions as required by the Urick theory.

Busby and Richardson (1957) examined sound absorption through sands and through glass beads immersed in a liquid. They found an increase of absorption with frequency in the range 0.5 to 3.0 megacycles/sec.

Table III

A Tabulation of Compressional Velocities for Sea Bottom Sediments

Investigator	Velocity-km/sec.	Nature of Sediment
Laughton (1957)	1.47-1.54	Red clay
	1.54-1.63	Calcareous ooze
Shumway (1956, unpublished) In Hamilton (1959)	1.48-1.57	Red clay
	1.49-1.52	Calcareous ooze
Sutton, *et al.* (1957)	1.43-1.57	Red clay
	1.48-2.77	Calcareous ooze
Hamilton, *et al.* (1956)	1.46-1.80	All samples tested
	1.68	Fine sand
	1.65	Very fine sand
	1.55	Silty-sand
	1.47	Medium silt
	1.46	Silty clay
Shumway (1960)	1.474, lowest value	Red medium clay
	1.785, highest value	Medium sand
Officer (1955)	1.46-1.76	Minimum seismic velocities quoted by Officer
Sykes (1960, in this report)		Terrigenous (?) clays, silts
	1.47-1.59	Piston Core 2
	1.49-1.60	Piston Core 3
	1.52-1.63	Piston Core 6

Wyllie, Gregory, and Gardiner (1956, 1958) have measured velocities in sedimentary rocks of interest to the oil industry. In their 1956 article they mentioned velocity experiments on a pile of glass slides. Only when the faces of the slides in contact were wet, did the average velocity of the pile equal that of an individual slide. These authors used this as an example to show how poor coupling or cracks in materials can cause errors in velocity measurements.

Murphy, Berg, and Cook (1957) made velocity measurements by a pulse method on porous cores of artificial material. Their results are summarized in the formula

$$\text{Eqn. (12)} \qquad v_c = kv_1 + C$$

where k depends on the porosity and type of cement; C, on the nature of the rock frame.

Hughes and Cross (1951) examined velocities in sedimentary materials as a function of pressure and temperature. Dry and water saturated runs on a sandstone and on Solenhofen limestone are illustrated in Figure 9. At low pressures the dry sandstone has a lower velocity than the water saturated. In effect we can reason that that part of the travel through the void space is faster for water than for air. At higher pressures the air is forced out in the dry case and a grain contact established. Water prevents this in the second. Neither effect is large in the limestone, and the order of the dry and saturated are reversed in sense compared to the sandstone.

Hughes and his associates (see bibliography), Birch (1960), and Ide (1937)

Figure 9
Velocities in Two Rocks as a Function of Pressure
after Hughes and Cross (1951)

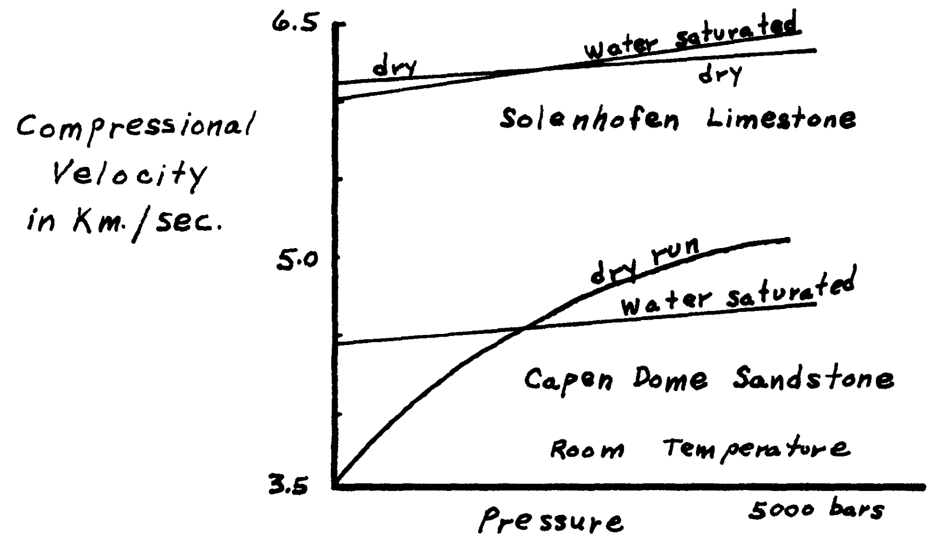
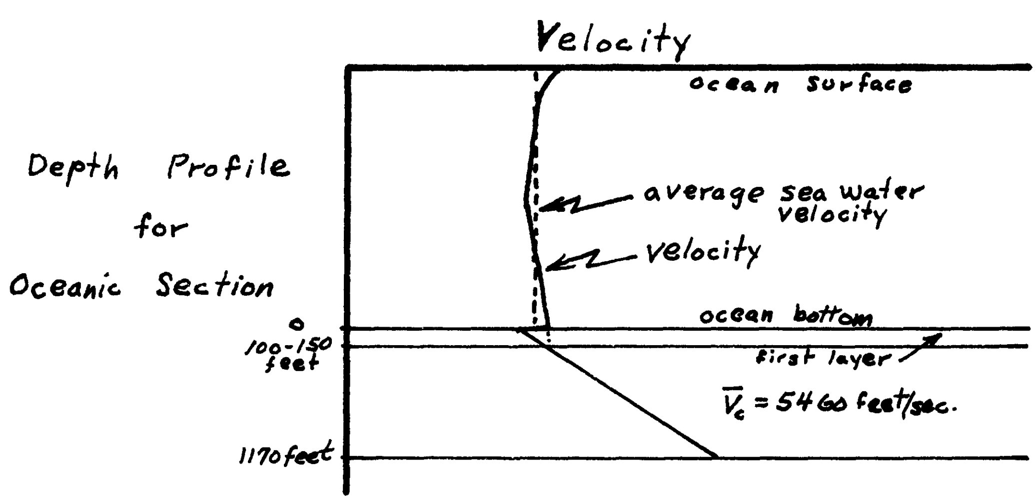


Figure 10
A Hypothetical Velocity-Depth Model as Given by
Officer (1955)



have made velocity measurements on a range of rock materials. Their techniques for velocity measurement are similar to those used on sediments.

Officer (1955) obtained information about a low velocity later just beneath the sea bottom with a wide angle seismic reflection profile. Officer reported an horizon at 100 to 150 feet and another at 1170 feet below the sea bottom. Velocity contrasts between the first layer and the sea water of 0.9 to 1.1 fit the reflection data for the first layer equally well. The average velocity for the second layer was 1.67 km/sec. From the shape of the distance-travel time plot, Officer concluded a velocity gradient must exist in that layer. Since this second arrival did not cross the bottom reflection at grazing incidence, a lower velocity just below the sea bottom is indicated. From this information and a consideration of the theoretical studies by Ament (1953) and Urick (1947), Officer constructed a plausible velocity-depth model. (See Figure 10.) Officer's bibliography includes other references to seismic studies of the low velocity layer.

V. A Description of some Methods for Velocity Determination

The development of transducer ceramics, particularly barium titanates, coupled with improvements in oscilloscopes has increased the ease with which precise sound velocity measurements can be made on small samples.

The interferometer gives very precise velocity and absorption data but only on gases and liquids. Direct measurements of solids are made difficult by distortion of the original signal. Hence, the advent of the direct arrival is often hard to distinguish from noise.

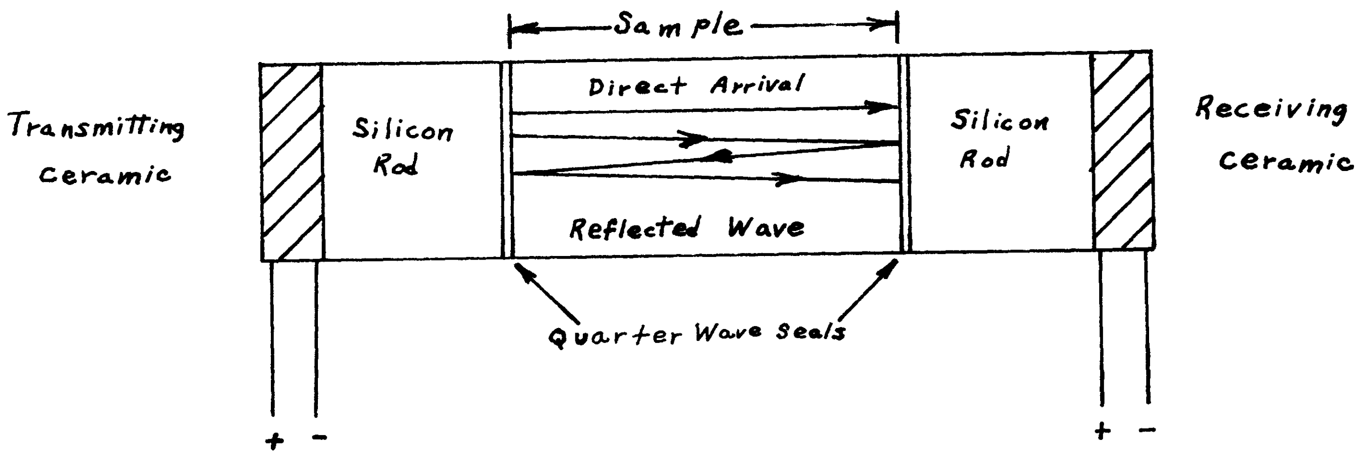
If a sample is placed between two ceramic transducers, the "distance" corresponding to the "time" measurement is not necessarily the distance between the ceramic faces, but rather a distance somewhat greater. Samples with a high attenuation such as sediments must be small. Thus the distance measurement is critical.

Consequently, several methods have been devised whereby a material of unknown velocity can be compared to that of a known. Laughton (1957) and Birch (1960) have used parallel delay lines. One branch contains the unknown; the other a variable mercury delay. The received waveforms are displayed on a double beam oscilloscope, and the length of the mercury column is adjusted until its waveform coincides with that of the waveform of the unknown. A travel time for the unknown is calculated from the velocity and length of mercury.

McSkimin (1950, 1957) has compared the travel times of direct and reflected waves as illustrated in Figure 11. Instead of using a rectangular voltage pulse across the transmitting crystal, McSkimin has employed a gated sinisoidal pulse. The frequency spectrum of this packet has a higher

Figure 11

Device Used by McSkimen (1950) for Velocity Measurements



concentration near the frequency of the sinisoidal input than the spectrum of a rectangular pulse.

A "Sing-around circuit" has been used by Cedron and Curran (1954) and by Myers, McKinnon, and Hoare (1959). The received pulse triggers another transmitted signal. Travel times are determined by measuring the frequency of the applied pulses.

The resonance chamber technique used by Shumway has been described by Toulis (1956).

VI. Experimental Procedures

Three cores taken in the Mediterranean Sea on CHAIN cruise number 7 (Woods Hole Oceanographic Institution) have been investigated by the author.

The cores (see Table IV) were taken by a piston corer with a plastic tube as a liner. After recovery of the coring device from the water, the liner containing saturated sediments was removed from the rest of the corer; sections of the liner were cut into four foot lengths, and the ends were sealed with scotch electrical tape and plastic. These cores were kept moist inside a deep freeze with the power turned off while aboard ship.

Velocity measurements were made with a set of probes on Piston Core 6 during the three weeks after that core was taken. However, bending of the probes in the sediment resulted in very poor data. All velocity data shown in this paper was taken in the M.I.T. Soil Mechanics Laboratory after improvements had been made in the probe holder and thicker probe rods had been installed.

To make velocity measurements, the plastic liner was cut in half lengthwise and one of these halves was removed. The exposed sediment was covered with Saran Wrap during the probe measurements to prevent water loss. Samples taken upon opening one core section and after the velocity measurements showed no water loss (less 0.2%).

A probe separation of about 1.4 inches was used in all velocity measurements. During a series of sound measurements, readings with the probes in distilled water would be taken at intervals so that any drift of the oscilloscope could be observed. Velocities in the sediments could be computed from the formula

$$\text{Eqn. (13)} \qquad v_l/v_m = (1 - v_l t_{ml}/L)$$

where v_m = compressional velocity of the sediment, v_l = compressional velocity of distilled water, t_{ml} = travel time difference between sediment and water, and L = probe separation.

Probe measurements were made in the center of each core sample and parallel to the long axis of the core cylinders. As a rule measurements were made at least every four inches along this axis. Readings near the ends of the core sections were avoided because of possible disturbances in these regions.

After velocity measurements had been completed, the sediment was sliced parallel to the long axis with a piece of thin sheet metal covered with Saran Wrap. Water content samples of 5 to 10 grams were taken immediately at the same places velocity determinations had been made. Half of the core was preserved for further testing; the other, allowed to dry partially in the air so that photographs could be made. Maximum visual contrast among various strata is attained when the sediment has partially dried.

Two of the Atterberg limits, the liquid and plastic limits, were determined at points along the three cores in the same manner as indicated in Lambe, Soil Testing for Engineers. The liquid limit is the water content at which a fine grained soil or sediment has an extremely small macroscopic shear strength as determined by an empirical test. This limit is a close approximation to that water content marking a transition between liquid and plastic states. Similarly the plastic limit is the water content of the transition between plastic and solid states. The difference between

Table IV

Core Data

Piston Core	Date Taken	Dates Examined	Length (feet)	Water D Depth	Position Taken
2	13 June 1959	10 Mar. 1960 19 Mar. 1960	27 1/2	1881 fm.	40-02N, 12-16.5 E Tyrrhenian Sea
3	17 June 1959	16 Feb. 1960 8 Mar. 1960	24	1870 fm.	40-08 N, 12-19E Tyrrhenian Sea
6	11 July 1959	17 Dec. 1959 28 Jan. 1960	24	1340 fm.	41-35 N, 04-52 E Algiers Provencal Basin

Table V

Specific Gravity Data of Sediment Solids

Piston Core	Soil Type	Depth in Core in Inches	Specific Gravity
2	clay	137-139	2.82
2	clay	139-141	2.95
2	silty-clay	215-217	2.82
2	fine, micaceous sand	223-225	2.715 2.710
2	clay	251-253	2.78
3	clay	68-70	2.80
3	clay	200-204	2.80
6	clay	114-116	2.83
6	clay	171-173	2.785 2.775

these two limits, or plastic index, is a relative measure of the extent of plasticity in the soil.

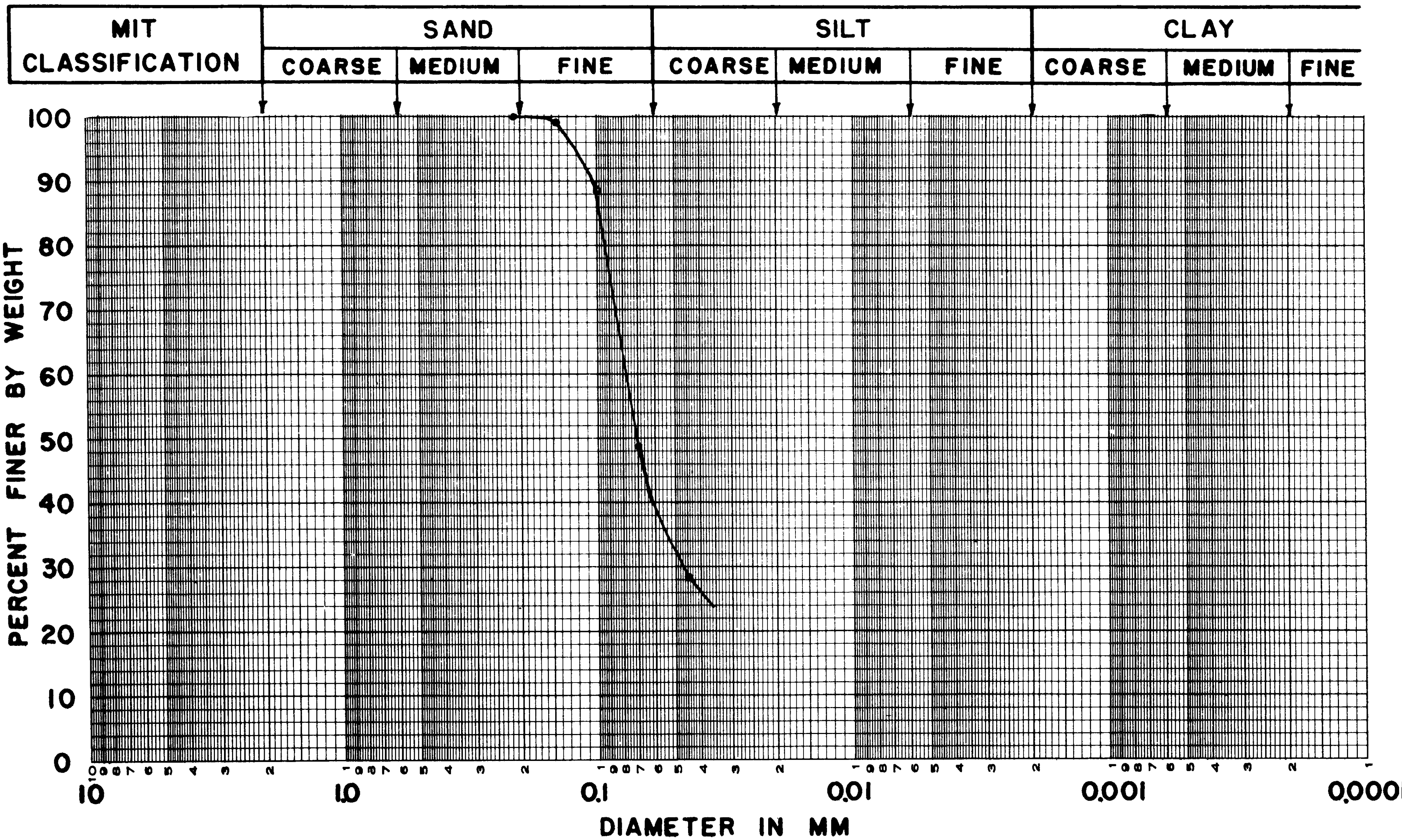
Eleven specific gravity determinations of the sediment particles were made as outlined by Lambe (as cited). These results are indicated in Table V. In this test as well as in water content determinations, a "dry weight of sediment" enters the calculations. Since all water cannot be removed easily from a fine grained sediment, the "dry weight" is defined for a sample heated at a given temperature for a given time. In this case samples were held at 105°C. for one day.

The large specific gravity in Table V of 2.95 is presumed in error. Such a high value is rare and unusual in soil engineering studies. The sediment listed above it with a value of 2.82 was taken from the core next to this sample and had visual properties and a water content identical to the sample in question. As can be seen from Table V, an assumed specific gravity of 2.80 in calculations for clays in these cores would no be in error by more than a few per cent at most. Although water content samples are easy to do in the laboratory, they are made extremely difficult at sea because a sensitive balance is required. Since specific gravities are so similar, density measurements would give accurate porosity data. An immersion technique not requiring weighing could be used to find the density.

Two grain size determinations were made on core samples. Grain size tests take longer to perform and give less information about physical properties of fine sediments than the Atterberg limits. The test on the fine, micaceous sand (Figure 12) was performed with a set of wet sieves; that on the clay (Figure 13), by the hydrometer method as described in

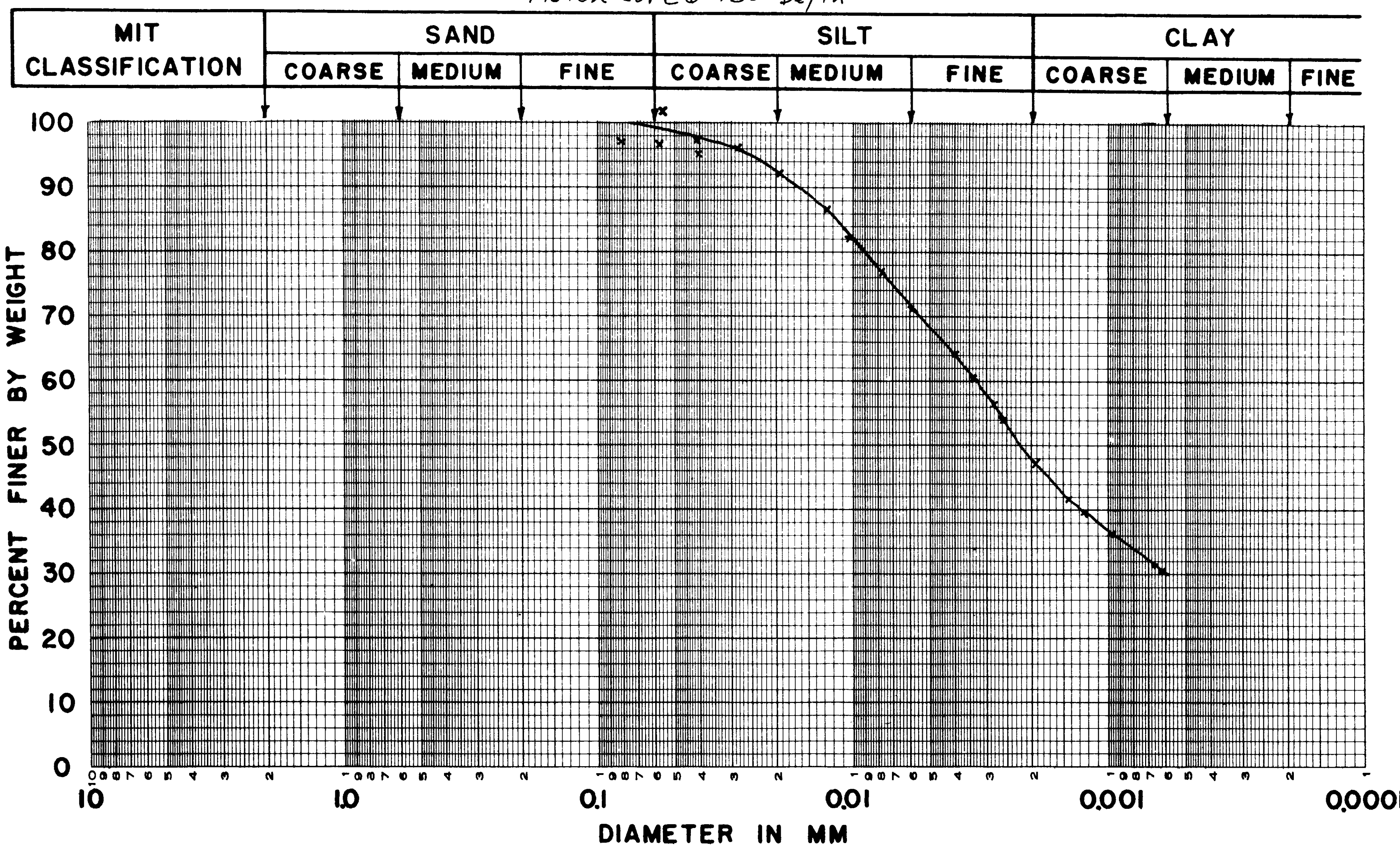
GRAIN SIZE DISTRIBUTION

Figure 12
Fine Micaceous Sand
Piston Core 2 - 224-226"



GRAIN SIZE DISTRIBUTION

Figure 13
Gray Marine Clay
Piston Core 6-130" Depth



Lambe (1951). The extremely good sorting in the sand as compared to the silty-clay are distinguishing features.

A one dimensionally drained consolidation test was run on a sample from Piston Core 6, depth 169-171 inches. Load increments from 1/8 kg/cm^2 were doubled every 24 hours up to a maximum of 16 kg/cm^2.

VII. Description of the Cores and the Coring Process

A visual sketch of the sedimentary column in each of the three cores is given on the graphs of measured physical properties along each core (Figures 28-30). Little emphasis should be placed on color differences in the clays. Some color alteration did go on during storage even though little change in physical properties is presumed to have occurred.

Two sizes of cores were taken on CHAIN 7, the 2.5 inch piston cores, and 1.5 inch diameter Stetson gravity cores. So far, velocity measurements have been made only on the former.

These three cores were taken in approximately two miles of water. Because of the motion of the ship, the corer must be dropped into the bottom and removed in a matter of seconds. Deformation of the sediment due to impact undoubtedly occurs. Deformation increasing with depth is especially noticeable in Piston Core 6. Pulling the corer out of the bottom also induces large stresses. The presence of the piston is believed to greatly alleviate this latter problem. In some cases, however, separations were observed in the sedimentary columns. No gap in sedimentation is believed possible. Visual correlation on opposite sides was observed in many cases. These separations are attributed to an imperfect seal by the piston during retrieval of the core from the bottom. No core catcher was used on the piston cores.

The sampler disturbs material both beneath it and in the tube. Changes in pore water pressure, structure of the sediment, and possibly some intergrannular stress would be expected. Since little can be done to avoid stresses in the sediment, an effort is made to prevent strain through use of a piston.

The piston action is initiated by weights which hang below the coring device and perform the "trip" when they hit the bottom. A soft bottom might not actuate the piston action. Hence, the cored section might not start until a deeper depth. In Piston Core 3, this was believed to have happened and was so recorded in the log of the CHAIN 7 Chief Scientist.

Piston corers used by soil engineers usually have very thin walls. This tends to minimize the amount of soil material that must be displaced either into or outside of the coring tube. Much thicker sampler walls are needed for deep sea coring. During the initial coring stages, sediment recovery would be greater than the natural section because of the displaced material. Wall friction becomes the dominant factor as more sediment enters the core tube. Sediment recovery then becomes less than unity. Moreover, this recovery ratio is a function of radius in the core barrel, with the smallest recovery at the walls. The piston acts to minimize these effects, but it cannot be fully effective in a deep sea corer where the piston is "floating" and not fixed to a rigid platform. The contrasting plastic natures of sand and clay are seen in some of the core photographs where sand has tended to exert a greater wall friction than the clay and hence is dragged along the core walls and mixed with clay. However, one possible difficulty does not seem to be present. The core section does not appear to deviate from normal incidence to the sea bottom.

VIII. Apparatus for Velocity Measurements

A series of experiments were tried aboard CHAIN 7 in June 1959 to measure the velocity of sound in deep water sediments.

Two matched transducers and probes had been loaned to Dr. J. B. Hersey by Dr. A. B. Wood at the Admiralty Research Laboratory, England. In these early experiments the two probes were mounted by clamps to a ring stand. In this way the probe separation could be varied, and the probes could be lowered into the medium being investigated. These probes as well as later versions were coated on all sides except the radiating faces with micro-balloons to prevent any disturbance to the internal wave by changes in the medium surrounding the probes.

As already mentioned, the probe rods were so thin they bent when inserted in the sediment. Clamping devices were none too steady. In September the probe device was altered. The probe rods were changed in size from .081 inch to .109 inch. The latter probes are illustrated in Figure 14. A probe holder which keeps the probes a fixed distance apart but allows them to be raised and lowered as a unit is shown in Figure 15.

During the CHAIN 7 cruise Dr. Earl Hays adapted an Edgerton "Pinger" circuit for exciting one of the transducers. During the summer a Tektronix oscilloscope with a variable delay unit was used to magnify that part of the oscillogram in the region of the direct arrival. Such an oscilloscope was not available for use at M.I.T. Consequently, the scheme as presented in Figure 16 with a delay line and Hewlett-Packard scope was used in all velocity determinations reported in this paper. The frequency measured was near the upper limit of this latter scope. Accurate data could not be

Figure 14 - Sketch of Acoustic Probe

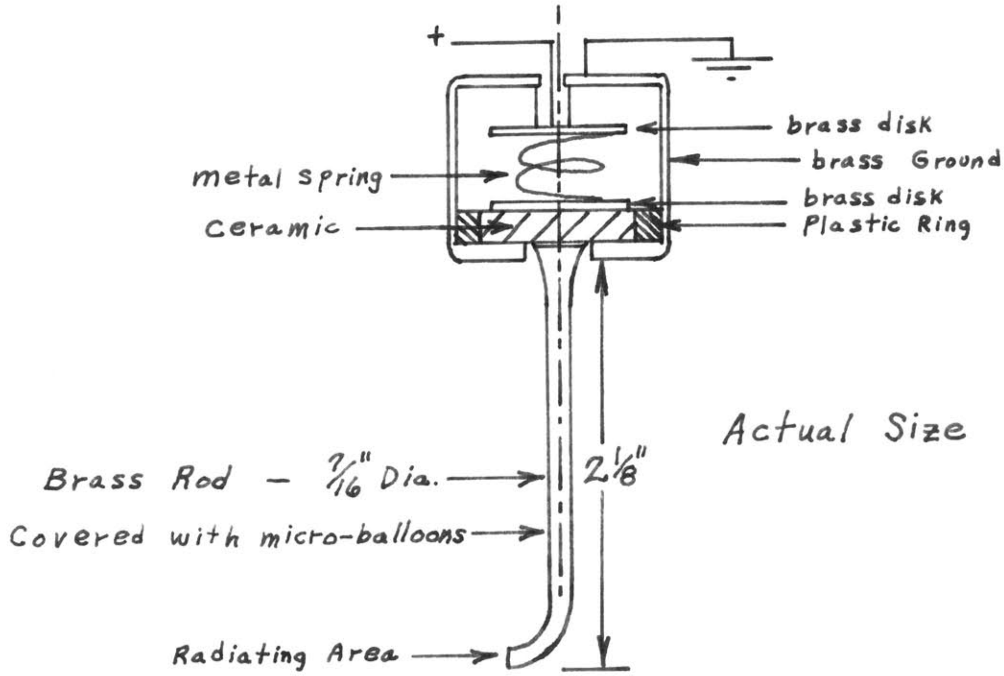


Figure 15 - Photograph of Probe Holder

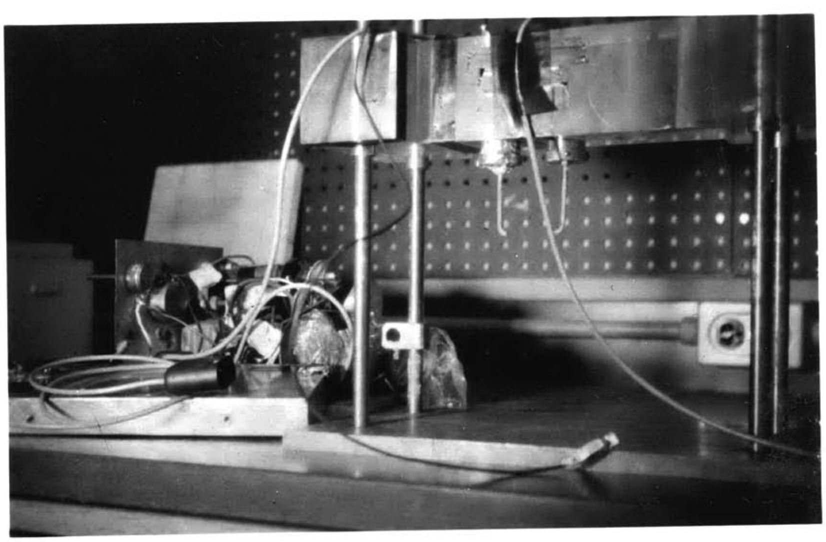
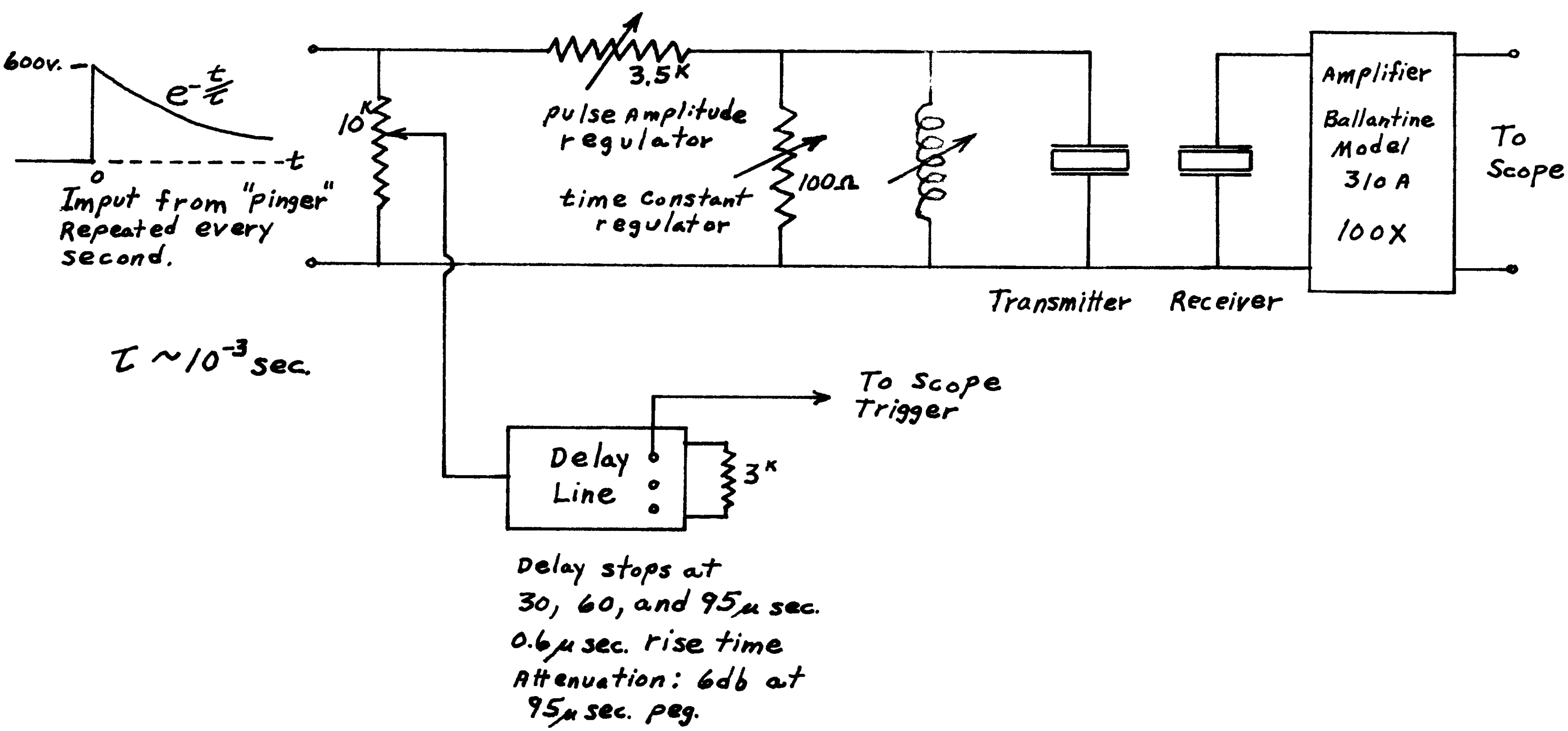


Figure 16
Circuit Diagram of Transducer Driver



read directly from the scope face with ease. With the variable delay line, an arrival could be centered on a hair-line. Polaroid photographs were necessary without this variable delay. Arrivals were read off the photos with a lined plastic rule. For each determination in either distilled water or sediment, photographs were taken on both 5 and 2 microsecond/cm. sweeps. The former was used to make certain the correct part of the waveform was being measured; the latter, for greater accuracy in readings.

The disk transducers used have two resonant frequencies, a radial mode at 157 kc and a thickness mode at 560 kc. The ceramics were driven at the higher frequency by a voltage applied to the transducer faces. With the equipment available, a gating circuit could not be constructed that would apply the necessary voltage (approximately 80 volts) to the transmitting ceramic. Therefore the same L-R-C circuit driven by the "pinger" continued to be used. An exponentially decaying sinusoid of frequency 560 kc was the nearest thing to a gated sinusoid that could be produced. The voltage applied across the transmitting ceramic is shown in Figure 17; a typical received pulse is illustrated in Figure 18.

However, unless the ceramic is driven with a waveform consisting of a single frequency (not possible for a pulse) the radial mode is excited. This results in a shear wave propagated into the probe rod. Mode conversion in a cylindrical rod (see Hughes, Pondrom, and Mims, 1949 and Cooper, 1947) leads to compressional waves of frequencies of 157 and 560 kc. The 157 kc component was especially bothersome. The use of a strong bonding agent (epoxy cement) between rod and crystal insured a good mechanical connection but also good coupling for the 157 kc component. One simple solution which I have not tried would be to filter the lower

Figure 17
Voltage Applied to Transmitting ceramic

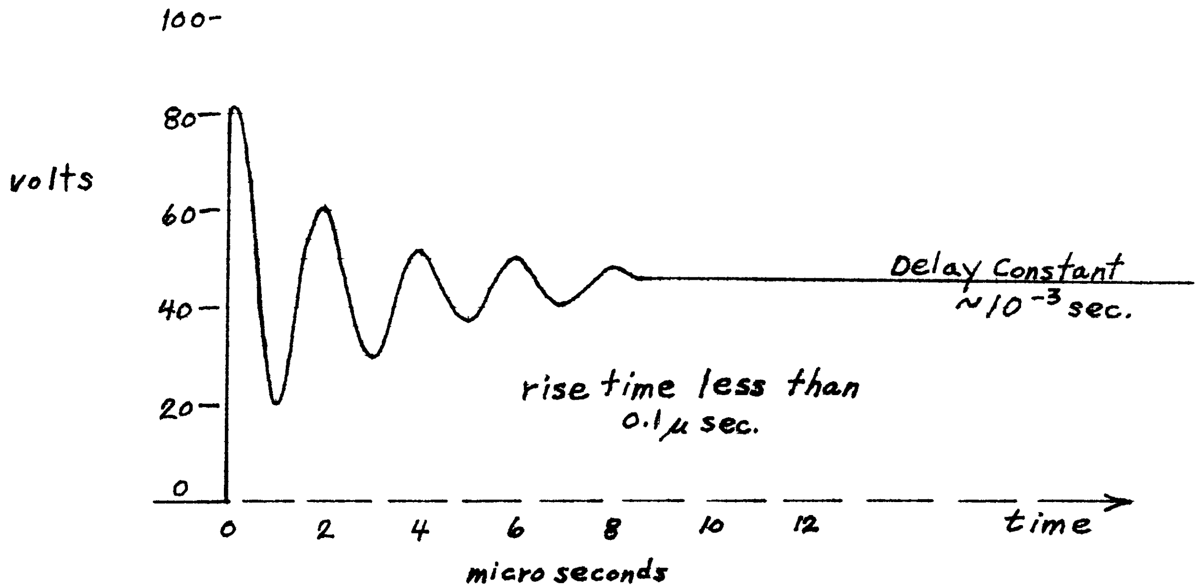
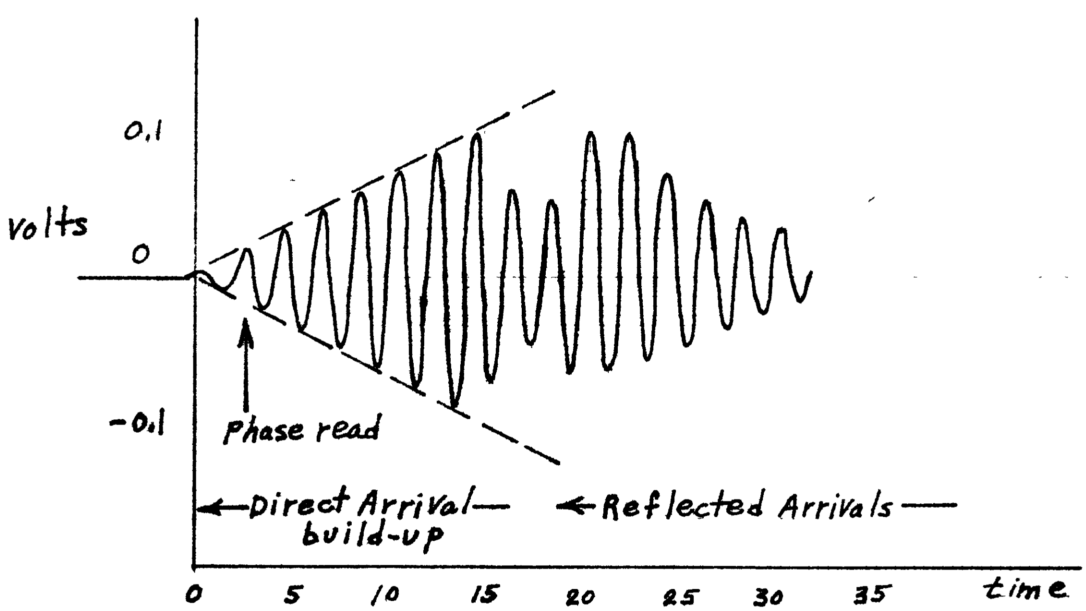


Figure 18
Received Waveform



frequencies out before displaying the received waveform on the scope.

By measuring water arrivals at varying probe separations, the approximate travel time in the probe rods was calculated. This agreed well with the travel time as computed for rods of diameter smaller than the wavelength as given by the formula

$$\text{Eqn. (14)} \qquad v_{bar}^2 = (E/\rho)$$

where v_{bar} = the so called bar velocity, E = Young's modulus. Rod travel time from measurements equaled 55 microseconds; from calculations, 56.4.

A metal shield around the receiving transducer was successful in eliminating electromagnetic pick-up from the transmitting ceramic.

A large "Q" of the circuit representation of the mechanical part of the transducer leads to a "ringing" of the transducer after it is pulsed. However, the disturbing factor is that arrivals build up slowly. Hence, the incidence of a wave is hard to detect.

Later arrivals in the oscillogram arise from a variety of factors. Some contributors are ringing of the transducers, internal reflections in the ceramics and probes, reflections from the container of the sediment or water, and surface reflections in the water. During an individual velocity run the first two would not be expected to vary. The effect of the core tube should not be observed before approximately 24 microseconds after the direct arrival for a probe separation of 1.40 inches, a core diameter of 2.5 inches, and the probes in the center of the sediment. Later arrivals in water would not be expected before this time provided neither the water surface nor the container sides are closer to the probes than 0.80 inches.

Since the start of the direct arrival is hard to detect, time differences between water and sediment were found by comparing the time difference between similar phases arriving shortly after the start of the direct wave, but before any reflections could arrive.

The wavelength of sound in water at 560 kc is .106 inches (0.27 cm.). Scattering and diffraction effects in the sediment tested should be small for grain sizes less than one-tenth the wavelength, or about 0.27 mm. This is a medium-sand in the M.I.T. Classification for soils. Thus this requirement is satisfied for all of the sediments in the three cores tested. However, sands tested experimentally gave poor waveforms when any could be obtained. Other investigators (Sutton, Laughton) reported large attenuation in coarse sediments.

IX. Results and Error Analysis

Velocity-depth profiles for three piston cores as determined by probe measurements are given in Figures 28, 29, and 30. Natural water contents, Atterberg limits, and plastic indices are plotted on the same depth axis. Zero depth is taken at the top of the section as observed in the cores. The failure of the core to trip at the bottom would result in this "zero depth" actually being below the sea bottom.

These three cores were taken in areas where the CHAIN echo sounder (a Precision Graphic Recorder) indicated subbottom reflections. Such reflections have been observed as deep as 100 feet below the sea bottom. Reflections picked from the records are plotted in Figures 28-30. The conversion to a depth scale from a travel time basis is based on an average velocity for sea water in the ocean of 4800 feet/sec. (1.473 km/sec.). Since this velocity is very similar to the average velocity in the cores, no correction to the velocity has been made in the reflection profiles. Although reflections have been plotted as bar lines, the spread is probably of the order of one foot as judged from several determinations on what are quite probably the same strata.

Piston Core 2 has a large number of strata with differing physical properties. A definite correlation between either velocity or water content and the reflection data is made quite difficult by the large number of events. However, most of the small groups of reflections occur at lows in the water content curve, highs in velocity. In some cases layers were so thin that a velocity for that layer could not be determined. Gaps in the core tube where no section is believed missing have been indicated.

Fewer variations in velocity or water content are present in Piston Core 3. The separate groups of reflections correspond in all cases but one to high velocity, low water content contrasts. Sand pockets and thin sand layers were found between 80 and 110 inches deep. This is the same region where a series of strong echoes are indicated. The percentage of sand in this volume is small enough so that no indication of these might be expected in the water content and velocity profiles. However, since the sand is badly disturbed and mixed with the clay, the relative sequence of layers is uncertain. Thus this evidence alone does not substantiate the belief the core penetrated a considerable distance in the bottom before tripping. Lack of deeper reflections than those shown makes a comparison of the velocity profile with any deeper reflections impossible.

Piston Core 6 shows marked changes in lithology only in the top three feet. Here a good inverse correlation exists between water content and velocity. This particular correlation is the most consistent one in the other two cores as well. This is the relation predicted by Urick (1947). The reflections at 15, 24-29, 35, and 120 inches correspond to expected departures in the velocity and water content profiles. Below about 125 inches, deformation in the cores is severe. Even though velocity measurements and water content samples were taken in the center of the sediment, the least disturbed section radially, the disturbance is still great enough so that little reliable data might be expected from this region. The possibility certainly exists that this deformed section was sucked into the core barrel by the piston action from the same layer. In this case, the core section would certainly not match the reflection record.

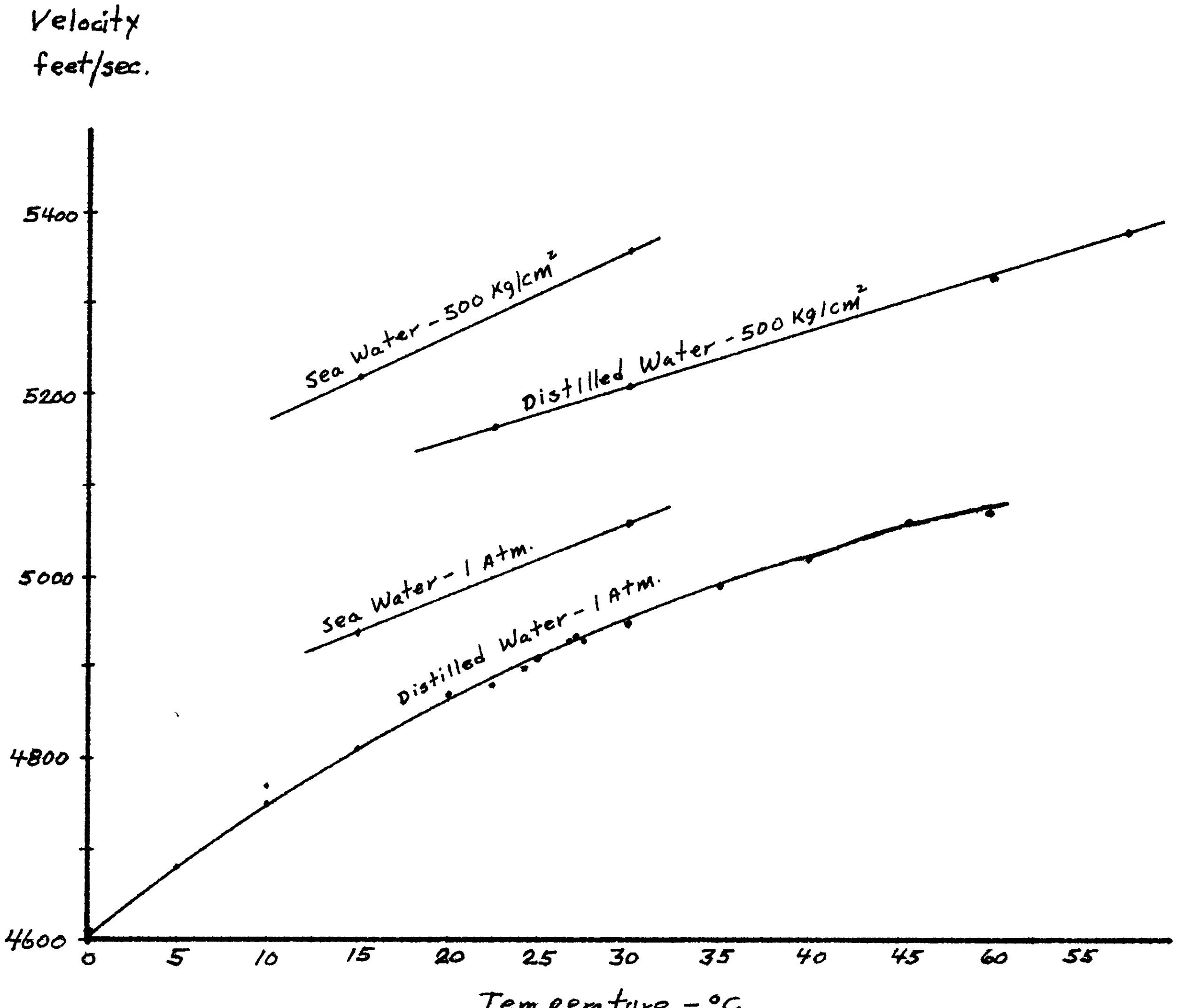
Velocity data for the three cores is plotted as $v_{sediment}/v_{distilled\ water}$. For velocities near the velocity of water this ratio is much less sensitive to variations in v_{water} than is $v_{sediment}$ alone. In Figure 19 velocities of distilled water and sea water have been plotted as a function of temperature. Temperatures in the distilled water and sediment tubes did not vary from a mean of 21°C. by more than ± 3°C. At atmospheric pressure this corresponds to a velocity of 1.487 $\pm$ .012 km/sec. Since this deviation is smaller than experimental error, no temperature corrections have been applied to the data. Variation in velocity of the sediment with temperature was assumed to be the same as for distilled water. For a discussion of this matter see Shumway (1958).

The choice of water as a velocity standard is desirable in the sense that no measurable dispersion (less than 0.1%) has been detected (McCubbin, et al., 1953) from 10^3 to 10^8 cycles/sec. Greenspan and Tschiegg (1956) reported that the effect of dissolved air on sound velocity in distilled water is less than 1 part in 10^5.

The ratio of velocity of sea water to that of distilled water is 1.026 at 15°C. and 1.022 at 30°C. (Figure 25). Applying this ratio to the three cores tested, cores 2 and 3 would have velocities up to a few per cent lower than sea water for substantial parts of their profiles. This is especially true for the clay and silty-clay parts. Core 6 would have no velocities lower than that of sea water. However, systematic errors of a few per cent could alter these generalizations. Minimum and maximum velocities for each core are compared in Table III with those of other investigators. Values used are averages at points along the cores.

Figure 19

Temperature and Pressure Dependence of Velocities of Distilled and Sea Water



Data from: 1. Carnevale, E.H. and T.A Litovitz
Journ. Acous. Soc. Amer. 27, 1955, p 795
2. Holton, G. -1951- Journ. Appld Phys.- 22 pp 1407-1413
3. Hubbard and Loomis -1928- Phil. Mag.- 5- p 1177
4. Smith, A.H. and A.W. Lawson -1954
Journ. Chem. Phys. - 22 - p. 351

No velocity-depth relation is observed. If any exists it is masked by porosity effects and experimental error.

It is estimated the photographs were read to 0.2 microseconds. For a probe separation of 1.40 inches this corresponds to a velocity deviation of .012 km/sec. Distance between the probes was read to an accuracy greater than .01 inch. However, deviation of the probe faces from parallel is probably of the order of .01 inch. This corresponds to a velocity deviation of .011 km/sec. The maximum random error is the sum of these plus the temperature error, or .035 km/sec. This figure is somewhat greater than 2% of the velocities measured. The scatter of most of the experimental velocities at a given place on a core is of the order of .03 km/sec. or less.

Water content is much lower and less variable in Core 6 than in either of the other two cores. The smaller variations probably can be attributed to a more homogeneous sediment. However, the clay water contents of cores 2 and 3 are much higher on the average than those in core 6. This difference is not reflected in the specific gravity of the solids. This deviation probably results from differences in mineralogy and clay structure.

In all cases but in one section of core 6, the liquid limit is a few per cent lower than the natural water content. The liquid limit presumably measures a very small but finite shear strength on a macroscopic scale. Sediments deposited on the sea bottom should have little or no shear strength. That the measured natural water content is greater than the liquid limit is an indication the sediment has not lost more than a few per cent of its water between coring and the time of measurements. Corrections to the velocity profiles for any water loss would be toward decreasing velocity for those values greater than the velocity of water. Corrections for velocities less than that of water cannot be predicted. Velocities would either increase or decrease depending on position of the minimum of velocity. In all probability the water contents would be less than the content of the minimum. Therefore these velocities would also be lower when corrected assuming the Urick (1947) theory gives an approximate prediction of velocity.

One region in core 6 is the only occurrence of natural water contents below the liquid limit. Since this particular section was the first section of any of the cores I examined, and it was consequently opened and sealed several times; a loss of water would not be strange. However, the trends on either side of this section are toward lower water contents. Liquid limits and plastic indices are lower in this region than in any other section of the core except the silt layer at 9 inches depth. These are indications the water content in this section represents the natural state.

The choice of the sample for the consolidation test was unfortunate in that it was from core 6, the core with greatest deformation. This deformation was not detected until sections could be dried. The curve A-B-C (Figure 20) is a void ratio, log pressure curve as calculated from the consolidation data. D-E is the slope of the e-log P curve as a function of pressure, P. Curves, such as this one, that do not approach a straight line on the e-log P plot at higher pressures have been found by Schmertman (1955) to represent disturbed samples. Using techniques outlined by Schmertman, I have constructed the envelope F-H-J-B for the consolidation of the undisturbed sample. The assumed overburden has been calculated from

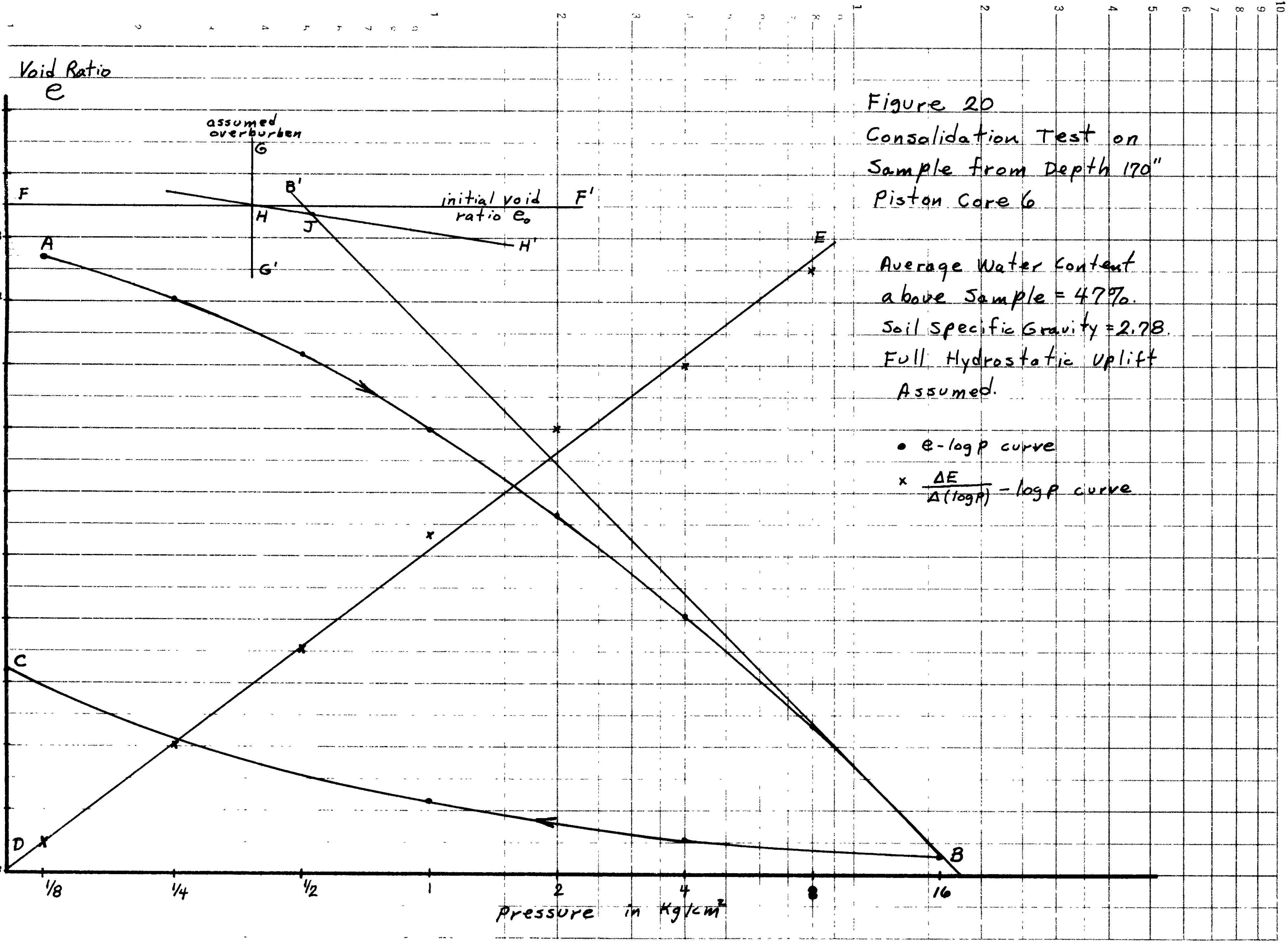


Figure 20
Consolidation Test on Sample from Depth 170" Piston Core 6

Average Water Content above Sample = 47%.
Soil Specific Gravity = 2.78.
Full Hydrostatic Uplift Assumed.

the presumed depth of the sample below sea bottom and the average void ratio and specific gravity of the sediment above this point in the core. The pressure at the point J is an approximation of the minimum past overburden. The fact that J is quite near H is an indication the overburder assumed was correct. Calculations were based on the assumption of full hydrostatic uplift.

The e-log P plot determined from the consolidation tests has been added to similar results (Figure 21) as collected by Hamilton (1959).

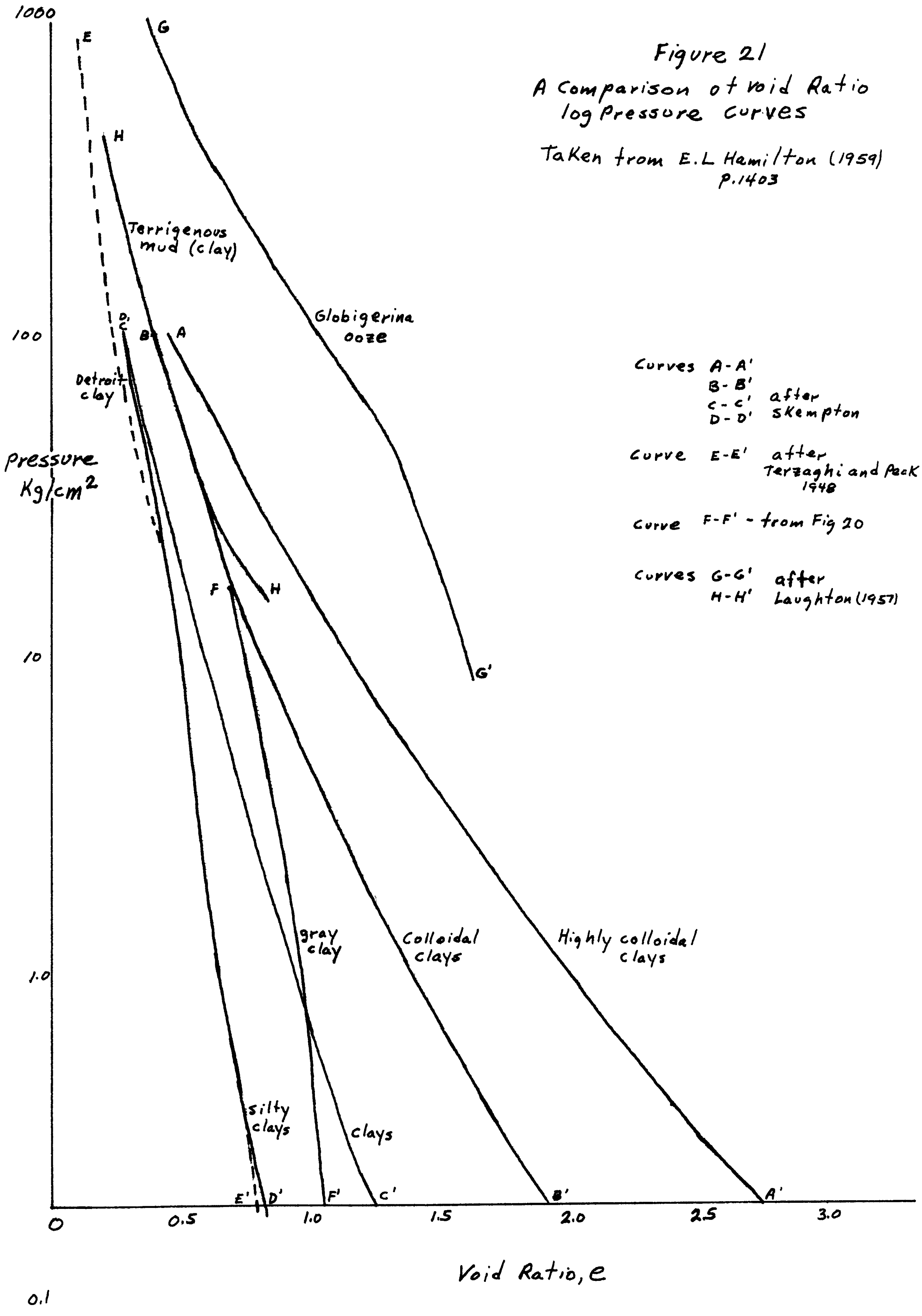


Figure 21
A Comparison of Void Ratio log Pressure curves
Taken from E.L Hamilton (1959) p.1403
1000
100
10
1.0
0.1
Pressure Kg/cm²
Void Ratio, e
0
0.5
1.0
1.5
2.0
2.5
3.0
E
G
H
D
C
B
A
F
H
G'
E'
D'
F'
C'
B'
A'
Terrigenous mud (clay)
Globigerina ooze
Detroit clay
gray clay
Colloidal clays
Highly colloidal clays
silty clays
clays
Curves A-A' B-B' C-C' D-D' after Skempton
Curve E-E' after Terzaghi and Peck 1948
Curve F-F' - from Fig 20
Curves G-G' H-H' after Laughton (1957)

X. Conclusions

The acoustic probe can be used successfully to measure velocities in sediments providing the grain size of the sediment is not too large. This limitation is reached at approximately the size of a coarse silt.

Velocities measured in three deep sea cores are not greatly different from the velocity of water. Many velocities less than that of sea water were determined. These measurements are in agreement with the results of previous investigators.

An accuracy of one to two per cent in velocity determinations is possible by the probe method. A definite advantage is that many readings can be made in a short time just after the core is taken.

Velocity in water saturated bottom sediments is largely a function of water content, or porosity. No velocity-depth correlation is observed in the cores. A consolidation test showed that water content would vary by only a few per cent at most as the effective stress on the sediment sample is increased from zero to an effective stress comparable to that existing at a depth of thirty feet of sediment. Therefore it is not surprising that a velocity-depth correlation was not noticed. Lithological changes that involve differences in water content have a much greater influence on velocity than any depth relationship.

The velocity-depth profiles as determined with the probes have been compared with moderate positive correlation to subbottom reflections recorded by a high resolution echo sounder.

Figure 22

Photograph-- Piston Core 3

9-17 inches on photograph correspond to sample depth of the same. A small clay layer in the center is between sand layers.

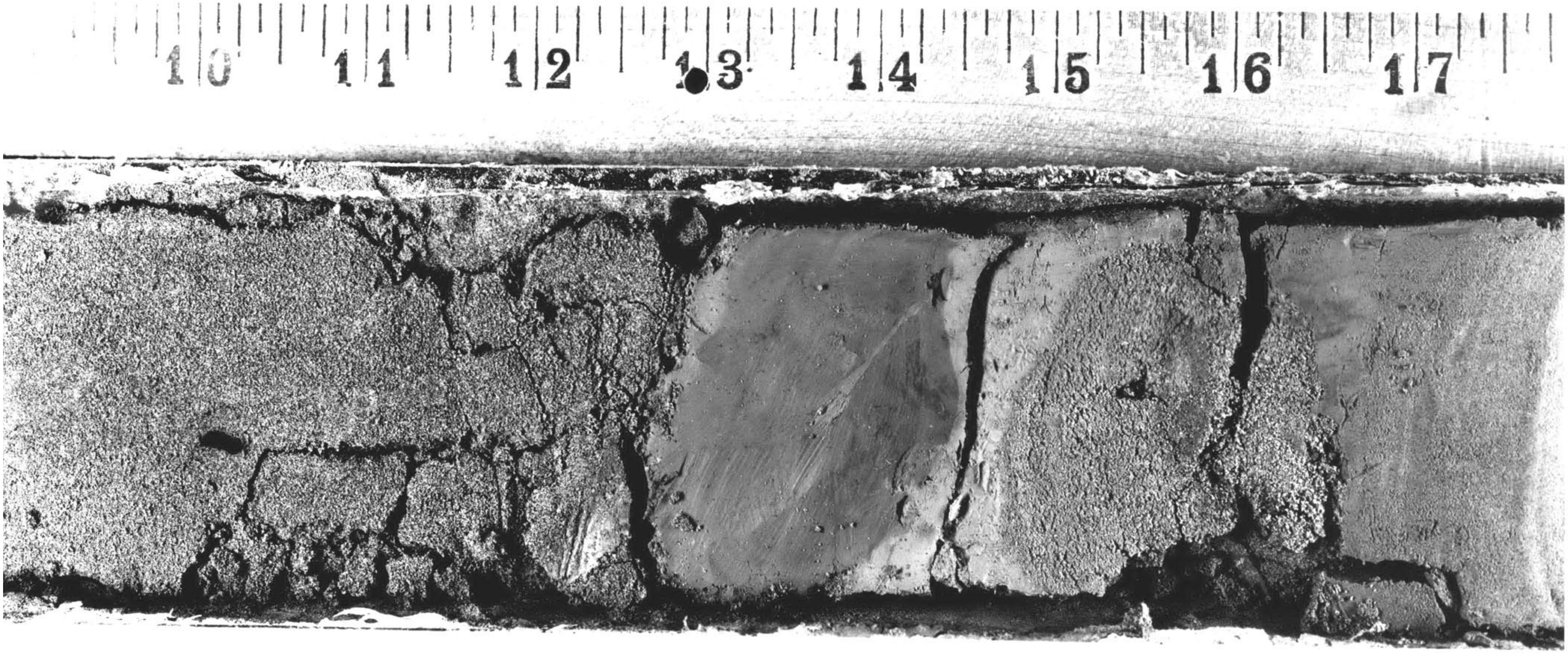

10 11 12 13 14 15 16 17

Figure 23

Photograph-- Piston Core 3

16 to 25 inches on the photograph corresponds to sample depth of the same. Many layers relatively undeformed are seen.

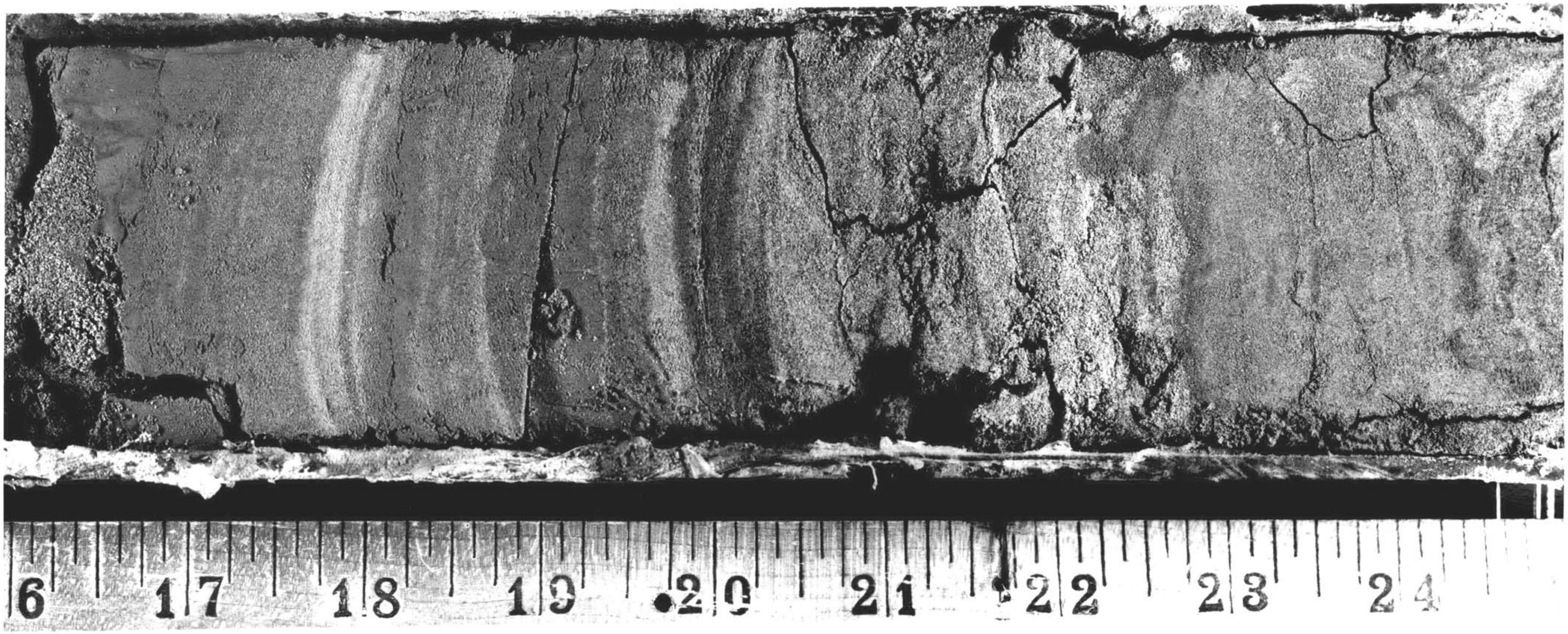

6 17 18 19 20 21 22 23 24

Figure 24

Photograph--Piston Core 3

24 to 32 inches on the photograph corresponds to a sample depth of 108 to 116 inches. A small sand pocket on the left is part of the evidence for correlating reflections in this region with a physical feature in the core.

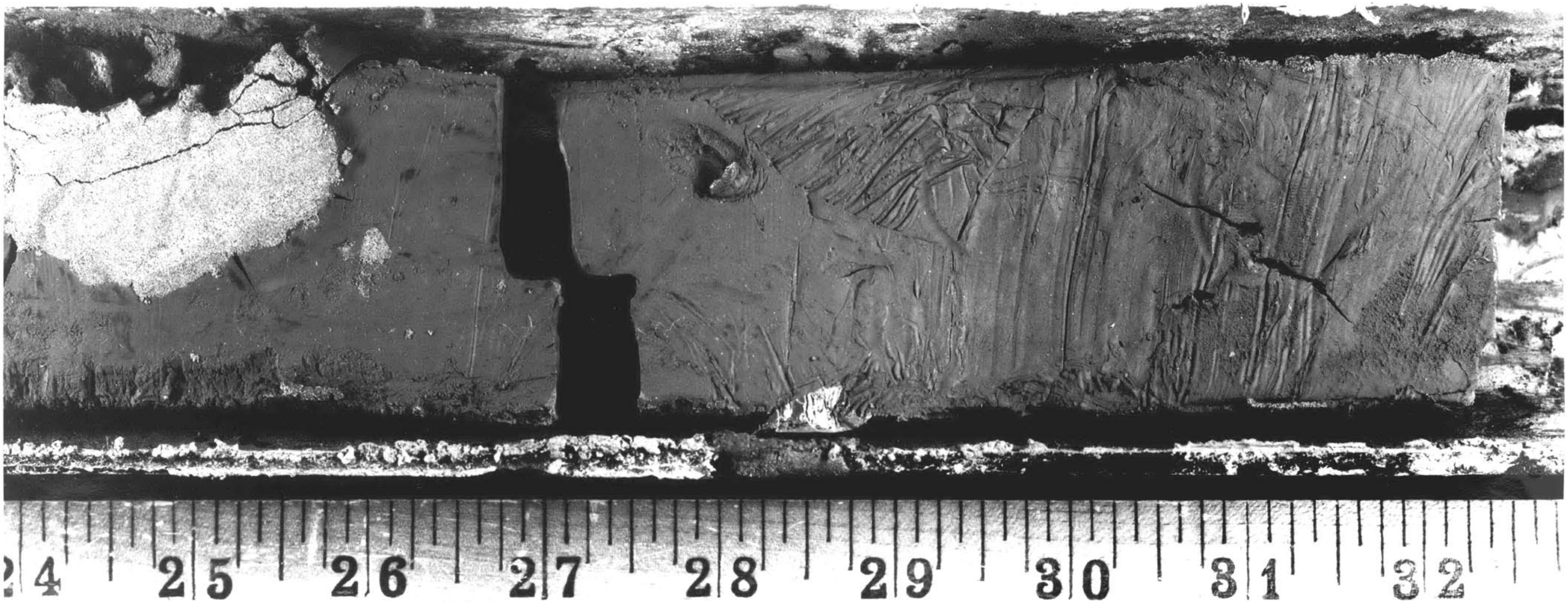

24 25 26 27 28 29 30 31 32

Figure 25

Photograph--Piston Core 3

17 to 25 inches on the photograph corresponds to a sample depth of 141 to 149 inches. Sandy layers have been deformed and drawn further into the core barrel in the center of the tube.

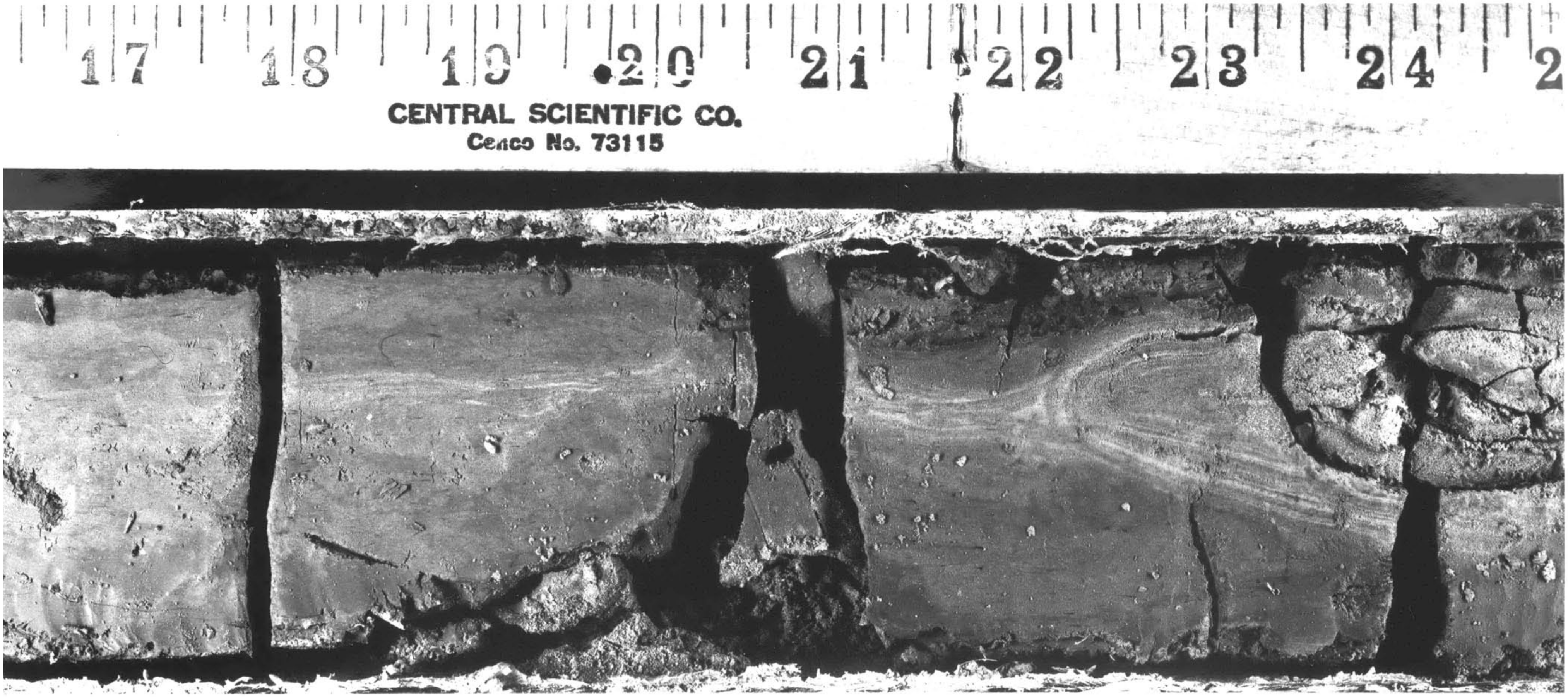

17 18 19 20 21 22 23 24 2
CENTRAL SCIENTIFIC CO.
Cenco No. 73115

Figure 26

Photograph—Piston Core 6

16 to 24 inches on the photograph corresponds to a sample depth of 14 to 22 inches. The effect of the walls in retarding the sediment flow is plainly visible. Small, dark pieces of clay have been drawn out parallel to the layering. These features are believed by the author to have occurred prior to the coring. They might possibly represent a type of turbidity flow.

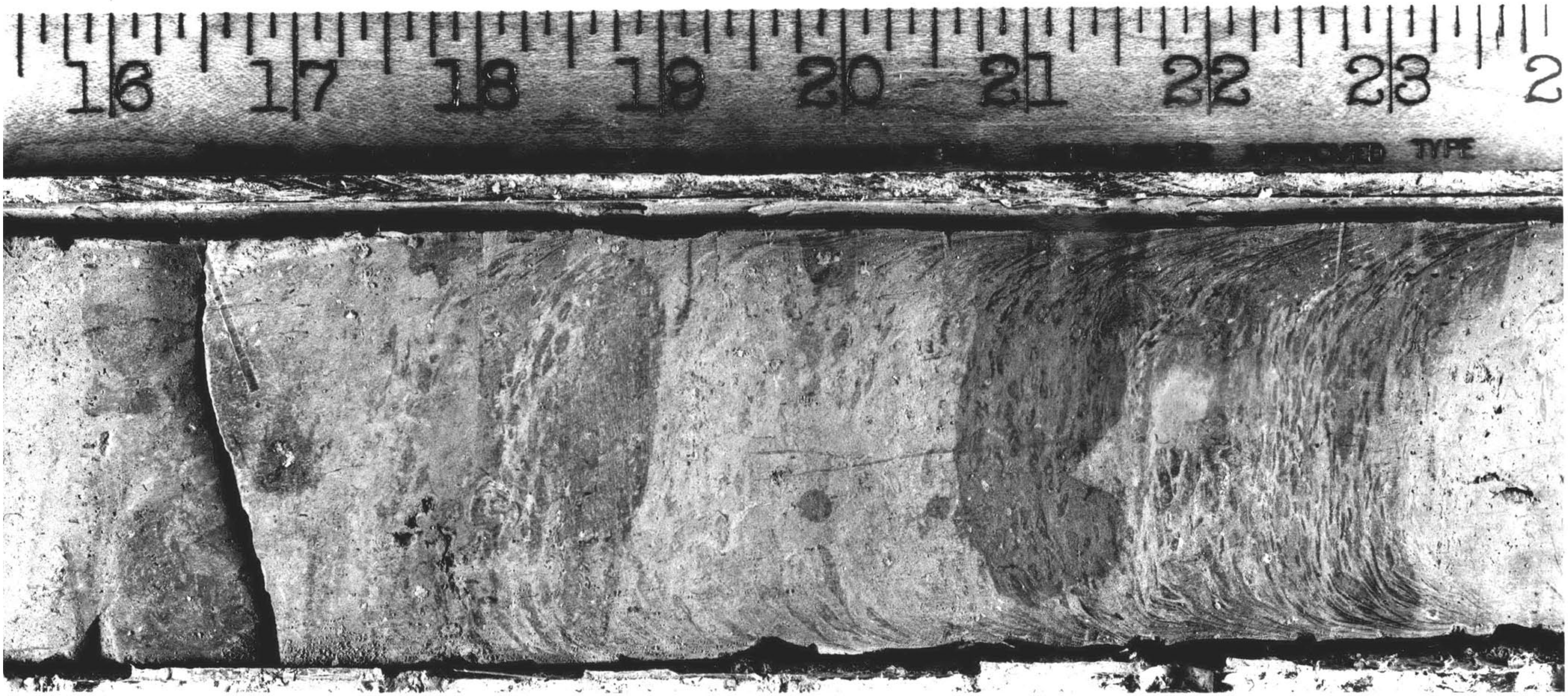

16
17
18
19
20
21
22
23
TYPE

Figure 27

Photograph--Piston Core 6

10 to 17 inches on the photograph corresponds to a sample depth of 258 to 265 inches. This is an illustration of extreme deformation in the lower parts of Core 6.

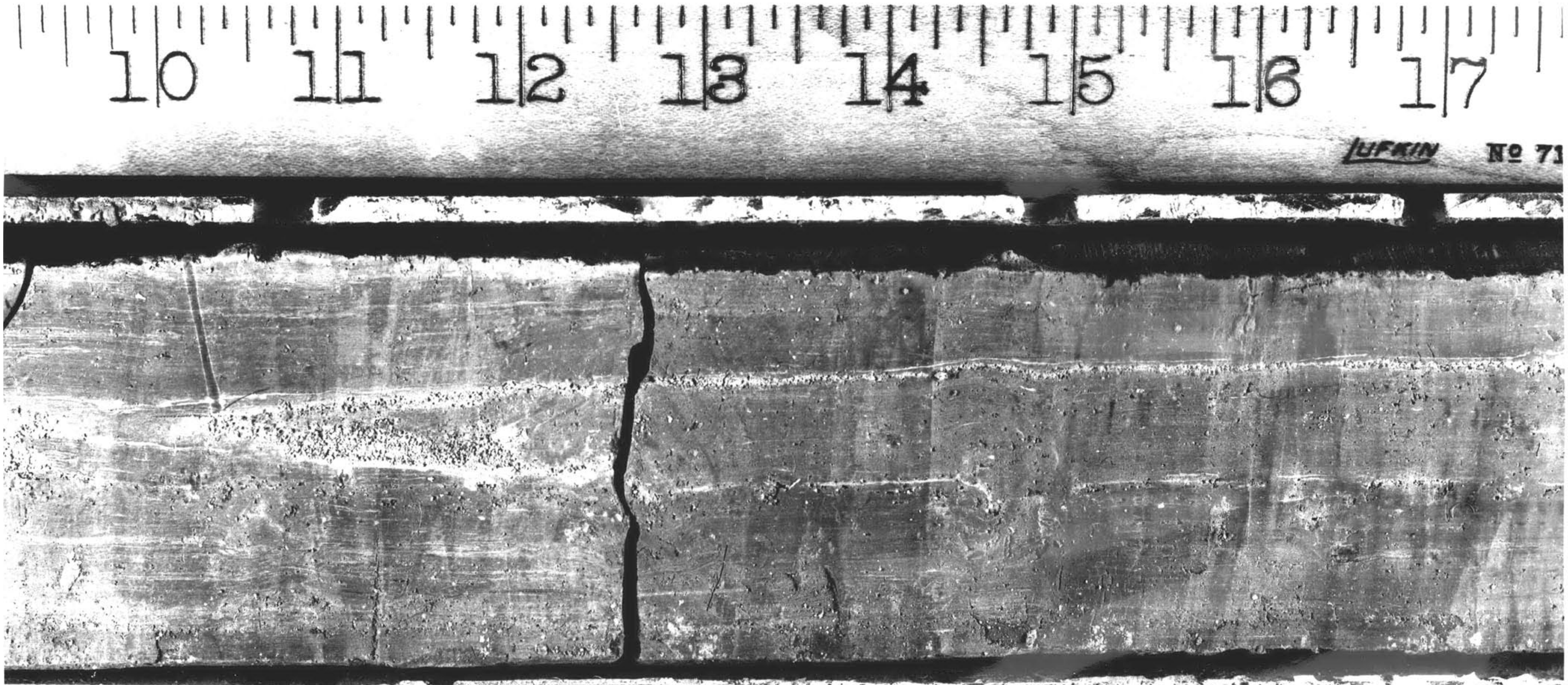

10 11 12 13 14 15 16 17
LUFKIN
№ 71

FIGURE 28 PISTON CORE 2

FIGURE 29 PISTON CORE 3
Core Description
empty
Sand
dark grey clay
Sand
sand pockets
dark gray-brown clay
1/4" sand
dark gray brown clay
mixed sand, clay
sand, clay layers
brown and dark grey-brown clays
Sand
tan clay
sample taken
tan clay
fossiliferous tan clay
Section 1
Section 2
Section 3
Section 4
Section 5
Section 6
Echo Record
Natural Water Content
Water Content (%)
P.I. = 46.5
P.I. = 44.8
P.I. = 33.9
Ratio Velocity Soil / Velocity Water
velocity sea water (approx.)
Key
x - Velocity Data
o - Natural Water Content
L - Liquid limit
P - Plastic limit
Echo Record
Record "picked" at times indicated
1417 - Core taken
S - Reflection Strong
F - Reflection Faint
Depth in Core in Inches
Depth in core in inches
KEUFFEL & ESSER CO. MADE IN U.S.A.

FIGURE 30 PISTON CORE 6

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