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Columbia University
in the City of New York

LAMONT GEOLOGICAL OBSERVATORY
PALISADES, NEW YORK

# PHYSICAL PROPERTIES OF MARINE SEDIMENTS

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## I. INTRODUCTION

The unconsolidated sediments that blanket the ocean floor are of widely varying thickness but seismic observations indicate that 200 to 400 meters in the Pacific (Revelle, et al, 1955) and one kilometer in the Atlantic (Ewing, et al, 1959) are fairly typical values for deep water. At present direct observation of these sediments is limited to such samples as may be recovered by dredging or coring operations, for drilling has been carried out only in the shallow waters of the coastal shelves. Knowledge of the physical properties of the great bulk of the sediments deeper than the few tens of feet reached by coring equipment is thus necessarily derived from geophysical observations.

The recoverable sediments of the topmost part of the sedimentary column may be studied in detail. Physical properties such as elastic wave velocities, attenuation, density, and thermal conductivity are directly measurable. The bulk properties depend ultimately on the values of other observables. These include water content, particle density, size and shape of particles, age, and chemical composition. Measurements may be static or dynamic. They may be carried out at elevated pressure, they may be made in various frequency ranges, they may sometimes be made in situ. Measured properties on small samples are related only in some average sense to the results of large scale measurements.

It will be noted that, whether on the scale of small samples, or on the scale of seismic refraction experiments, acoustic measurements play a major role in any study of physical properties of marine sediments. They provide direct information on compressional wave velocity, shear wave velocity, impedance, attenuation, and Poisson's ratio. If density is known independently, they supply in addition dynamic values of compressibility, shear modulus, and other elastic constants. Combined with appropriate travel time analysis they give variation with depth of elastic wave velocities and, consequently, of depth variation of any derived quantities. In certain cases, such as dispersion of elastic waves in shallow water, some information on depth variation of density is obtained. Approximate depth variation of many physical properties may be derived from velocity measurements by combining them with empirical relationships connecting velocity with density or velocity with porosity. These relationships are based on experimental observation and inferences from them are capable of being partially checked by comparison of gravity with seismic interpretations. The existence of such empirical connections is a consequence of the limited range of substances ordinarily occurring in sedimentary deposits and is to be regarded as a fact of geology rather than the operation of physical law.

## II. RELATIONSHIPS AMONG OBSERVABLES

The physical properties to be dealt with in this article are listed in Table I, together with the symbols and units employed. Observables that specify particle statistics or sediment type have been omitted from the table but some will be discussed later in connection with results. Also omitted are observables that are not properties of the sediments alone. Group velocity of a dispersed wave in shallow water, for example, depends to some extent upon density and elastic properties but it is especially a characteristic of a particular depth variation of these quantities Heat flow is not a property of the sediments at all except to the extent that sources within the sediments might contribute to the total flow.

Equations relating physical properties that make possible the computation or estimation of one quantity from measured values of others are given in Table II. Some are generally valid both for sediments and sedimentary rocks; others are valid only in limiting cases. Those equations that are strictly valid only for isotropic elastic media may, when applied to porous media, be considered to define dynamic values of such quantities as $\kappa$, $\mu$, $\sigma$, and E.

A marine sediment is a particle aggregate with the interstices completely water filled. A number of theoretical discussions of these porous or granular media have been reported. The reader is referred to papers of Ament (1953), Biot (1956), Zwikker and Kosten (1949), Gassman (1951), and Paterson (1956) for information on their elastic behavior. Parasnis (1960) has computed the thermal conductivity to be expected under certain special assumptions with regard to grain shape. Porous media are in general neither isotropic nor perfectly elastic. Even when isotropic on a scale large compared with grain sizes, two dilatational waves and one rotational wave, each with a different velocity, can occur. Both attenuation and the velocities have been shown to increase with increasing frequency. Sutton, et al (1957) have shown, however, that over the frequency range of from a few tens of cycles per second to one megacycle the velocity change is not likely to exceed one percent for sediments usually encountered. Birch and Bancroft (1938) noted no significant difference between velocities measured on rocks at high frequency and corresponding seismic velocities. In effect, then, particles and fluid may usually be considered to move together. If compressional velocities are observed that are higher than predicted by the Wood equation (Equation 8) the increment will be attributed to an increase of quantity $K + 4/3\,\mu$ caused by compaction and cementation.

Despite the exceedingly complex nature of granular media and the multiplicity of processes that affect the behavior of the aggregate as a whole, it is clear from Figure 1 that some broad general trends may be discerned in the experimental results. The Wood equation (Equation 8) is an approximate lower limit to observed compressional velocities for a given porosity. As Wyllie, Gregory and Gardner (1956) have pointed out, the time average equation (Equation 9) is a fair representation of the main trend of observations for consolidated sediments. The spread of observations is sufficiently limited that rough predictions of porosity and density may be made if velocity is known. Thus if the compressional velocity is 4 km/sec it is unlikely that porosity will be outside the range 0.15 to 0.25 no matter what other attributes the sediment may have. From Equation 6 density is a linear function of porosity, provided the individual grains of the sediment are all of one kind. Nafe and Drake (1957) have shown that nearly the same straight line would fit the observations for sandstones, limestone shales and ocean sediment cores. A porosity range of 0.15 to 0.25 means that density is in all likelihood between 2.3 and 2.5 $gm/cm^3$.

Thus the similarity of particle densities in common sediment-forming substances leads to great simplification in empirical description of physical properties.

Before proceeding to a discussion of methods and the summary of results it is necessary to comment on the small number of actual measurements that have been made. Of reported observations there are disproportionally many from easily accessible areas, particularly from shallow water. The only in-situ measurements other than seismic have been made in depths that permit operations by divers. Precautions against disturbing the sediment samples to be studied, are more effective in the case of shallow water sediments. To include in the discussion of physical properties sufficient information to make possible predictions and estimates of properties not directly observed, some data from bore holes and from artificial compaction experiments are presented. A distinction between observations on actual marine sediments and others of possibly similar character will be maintained throughout the article.

## III. METHODS OF MEASUREMENT

Among the methods described below some are direct measurements and some yield information on physical properties only through a comparison of experiments and theory involving observables other than the ones sought. Thus, velocity may be measured by direct timing or it may be determined indirectly by comparison of measured with computed curves of phase velocity as a function of period. It is not intended to provide a comprehensive summary of available methods but rather to indicate some that are typical and that are currently in use. Methods of static determination of elastic constants are not included on the grounds that they are adequately treated elsewhere. Moreover, results are frequently incompatible with observed velocities whic depend on dynamic values. Particular notice is given to methods having spe al advantages for use at sea.

A. Density, Porosity, and Average Grain Density

These three quantities have been measured (Sutton, et al, 1957; Hamilton, et al, 1956) directly for sediment samples by weighing a sample c known volume and reweighing after drying at 100 to 115°C.

Then

$$\rho = W/V$$

$$\rho_g = \frac{D\,\rho_w}{W\,\frac{\rho_w}{\rho} - W + D}$$

$$\phi = \left(\frac{W - D}{W}\right)\frac{\rho}{\rho_w}$$

where W, D, represent wet and dry weights, respectively, $\rho_w$ the density of water, $\rho_g$ the average grain density and V the original volume. Altern - tive expressions may be written by replacing W by $V\rho$

Sutton, et al (1957) have described a method for measuring bulk dens: that is particularly useful aboard ship where accurate weighing is not possib: . The density of a known volume of sediment from a freshly obtained core was determined by suspending the sample beneath a modified spirit hydrometer i: distilled water. The sample was obtained by inserting a small cylinder into core, meanwhile holding a piston fixed at the core surface to prevent compaction of the sample. After density measurement samples were preserved in sealed containers so that dry and wet weights could be also determined in the laboratory for determination of $\phi$ and $\rho_g$ and an independent measurement of $\rho$

Hamilton, et al (1956) have described underwater density sample collection by a diver. A metal tube after insertion into the sediment was sealed at the top by a plastic disk. Another disk was placed at the bottom of the tube by burrowing down along the side. As the methods of Hamilton, et al, and Sutton, et al, suggest, it is important to avoid any compaction in the process of sample collection. An excellent example of the distortion in length that can be produced by the act of sampling is shown by Ericson and Wollin (1956). They compare the lengths of corresponding intervals of cores taken with and without a piston. Nearly a 50% reduction in length occurs in the upper part of the core taken with no piston.

B. Compressional Wave Velocities and Attenuation

Compressional velocities and thicknesses of principal layers, and sometimes shear velocities, are obtained by the methods of seismic refraction. Since frequencies are typically of the order of 100 cycles/sec (wave lengths of the order of 20 meters) and path lengths within the sediment may be many kilometers, the resulting velocities are average values over large volumes of material.

Velocity measurements on sediments in situ were reported by Hamilton, et al (1956). This was accomplished by insertion of two probes each with a barium titanate transducer into the sea floor by a diver. The probes were one foot apart and penetrated to a depth of 6 inches. Velocities were then measured by direct timing of 100 kc pulses.

Sutton, et al (1957) measured sound velocities in freshly obtained cores by direct timing of an acoustic pulse across the diameter of the core. A velocity standard was used to obtain the origin time of the transmitted pulse.

Velocity measurement by direct timing, though not on marine sediments, have also been described by Hughes (1957), Paterson (1956), and Wyllie, et al (1956). Laughton (1957) and Kershaw (unpublished) have employed direct timing methods to study the effect of compaction on physical properties of marine sediments.

Shumway (1956, 1960) measured both compressional velocity and attenuation by exciting axial modes of oscillation in sediment samples contained in thin-walled plastic cylinders. The values of velocity reported by Shumway were obtained by comparison of resonance frequencies for a given mode of a sediment filled with a water filled container. Attenuation was measured by determining the Q of the system from the resonance widths at half power points. Toulis (1956) has derived equations relating Q and the resonance frequency to the attenuation and sound velocity respectively for this system. He has also shown how to introduce corrections for finite thicknesses of the cylindrical container.

C. Shear Wave Velocities

Although there is no doubt that shear waves occur in marine sediments there are many practical difficulties to be overcome in measurements and few reliable observations have been made. The coefficient of rigidity is normally so small that measurements by direct timing on recovered samples have been accomplished only under elevated pressure. A. S. Laughton (1957) measured shear wave velocities in a sample of globigerina ooze compressed between porous disks. Shear waves were observed at pressures about 500 $kg/cm^2$ on the initial compression. On decompression shear waves could be identified at pressures as low as 64 $kg/cm^2$. Indirect evidence from seismic refraction measurements for the existence of shear waves in marine sediments has been reported by Nafe and Drake (1957). An excellent measure of an average sedimentary shear wave velocity is to be found in the dispersion of oceanic Love waves and higher mode Rayleigh waves. The observed periods and low values of group velocity near the group velocity minimum cannot be accounted for without assigning a rigidity to the sedimentary column. According to Oliver and Dorman an average shear velocity between 0.25 and 0.60 km/sec is required.

D. Dispersion Analysis

Surface waves traveling in a layered half space are in general dispersed with phase and group velocity determined by the depth variation of elastic wave velocities and density. Normal mode propagation in shallow water, Rayleigh and Love waves in the crust and upper mantle of the earth are familiar examples of such dispersed wave trains. The dispersion may be computed from an assumed structure. By comparison of observed with the computed phase and group velocities the validity of the assumed velocity-density depth variation may be tested. A comprehensive treatment of layered media is to be found in Ewing, Jardetzky and Press (1957).

There are numerous examples of dispersion studies relating to shallow water sediments. Pekeris (1948) has computed dispersion for a number of two and three layer cases, each layer being assumed liquid and of constant density and velocity. Press and Ewing (1950) and Tolstoy (1954) have treated the case of a liquid over a semi-infinite elastic solid. Tolstoy (1960) and Satô (1959) have considered cases of continuously varying material properties.

The most detailed dispersion studies involving properties of sediments apply to shallow water. Some work yields information on deep water sediments. Oliver, et al, used the fluid thickness indicated by Rayleigh wave dispersion across ocean basins to estimate average sediment thickness. Katz and Ewing (1956) identified a constant frequency arrival occurring on deep water refraction records at ranges of the order of 60 km as a wave propagating for part of its path in a thin layer just below the water-sediment interface with a velocity lower than that in the water. Much more information on deep water sediments may be expected in the immediate future as machine computations of dispersion for a wide variety of assumed structures become available for comparison with observations.

E. Thermal Conductivity

Thermal conductivity has been measured both by steady state and transient methods. Ratcliffe (1960) described apparatus for steady state measurements for a few ocean sediments. He used two disk shaped samples of like dimensions with a disk shaped heat source between them. The sediment samples were confined in ebonite rings. The ends of the sediment disks opposite the heat source were cooled by circulating water. The apparatus was subjected to a vertical load to insure good thermal contact and the whole apparatus was surrounded by glass wool insulation to reduce loss of heat at the edges. From the dimensions, power input and temperatures at the hot and cold end plates, the thermal conductivity was determined. Zierfuss and van der Vliet (1956) employed a steady state method to measure thermal conductivity of sedimentary rocks. They also used cylindrical samples but arranged in such a way that all of the heat passing through the sample passed also through a crown glass standard. Thus the only measurements required were of temperature gradient in sample and standard. The ratio of the thermal conductivities is then equal to the inverse ratio of the temperature gradients. Zierfuss and van der Vliet depended for shielding on a heated glass tube having temperatures at all points very near to those at corresponding points of sample and standard. In this way edge losses were minimized.

Von Herzen and Maxwell (1959) measured thermal conductivity on freshly obtained cores by insertion of a small needle probe containing both a heating element and a thermistor. The needle constitutes basically a line source and after a sufficiently long time the measured temperature varies linearly with the logarithm of the time. The slope of the line is $q/4\pi K$ where q is the heat input per unit length of cylinder and K is the thermal conductivity. For validity of the linear relationships the time must be large compared with the square of the probe radius divided by the diffusivity (diffusivity = $K/\rho$ x thermal capacity). The probe used by Von Herzen and Maxwell was constructed from a hypodermic needle 6.4 cm long and 0.086 cm in diameter. Power input was of the order of 2 watts and time required for measurement 10 minutes or less. Measurements compared favorably with steady state measurements carried out on selected samples. For application of measured values to heat flow determination corrections for change of temperature and pressure between place of measurement and the sea bottom are required. These corrections are discussed by Bullard, et al (1956). They amount to a 5 to 10% reduction on account of the temperature decrease and something less than a 3% increase to correct for pressure.

## IV. SUMMARY OF RESULTS

### A. Density-Porosity

In Figure 2 are plotted measurements of density as a function of porosity for water saturated sediments and sedimentary rocks. For a two component mixture of fluid and particles of a single kind the relationship is necessarily linear. It is remarkable that when observations are plotted without regard to sediment type the observed points come so close to fitting a single straight line. In the figure are plotted some 300 points representing measurements on limestones, dolomites, sandstones, shales, clays, sands, gravels, and ocean sediments. Measurements have been taken from Birch (1942), Hamilton, et al (1956), Shumway (1960), Sutton, et al (1957), and Kershaw (unpublished) and others. The range of porosity values is 0 to 0.25 for limestones, 0 to 0.4 for sandstones and 0 to 0.5 for shales. Clays, sands, and gravels cover the middle range from $\phi = 0.2$ to 0.6. Ocean sediments range from $\phi = 0.4$ nearly to 0.9. Deviations from the straight line drawn on the figure are greatest in the case of limestones and dolomites. Of about forty points plotted, two thirds lie almost on the line and the remaining one third lie above the line. A few large deviations occur in the middle range of porosities for clays and gravels but even most of these lie close to the mean line.

B. Compressional Velocity-Porosity

Observed compressional velocities are plotted against porosity in Figures 1 and 3. It is apparent that they fall into three groups: the sedimentary rocks at high velocity, the ocean sediments at low velocity, and artificially compacted sediments intermediate between the others. The apparent separation into groups is probably a consequence of easy accessibility of samples in the high and low velocity ranges. Wells are cased at the top; coring devices have limited penetration. In Figure 1 observations are compared with the Wood equation, the time average equation, and with a representative of a one-parameter family of equations proposed by Nafe and Drake (1957). Plotted points for ocean sediments have been obtained from Sutton (1957), Hamilton (1956), Shumway (1960), Laughton (1957) and Kershaw (unpublished), and those for sedimentary rocks from a variety of sources. In Figure 3 observations are distinguished according to sediment type. It will be noted that most of the observations fall within the n = 4 to n = 6 range defined by the equations of Nafe and Drake. Since these equations take roughly into account the effect of grain to grain contacts, they do not fit the distribution of observations on uncompacted samples. On the other hand, they do outline the main trend of points and in particular fit well to trajectories on the V-$\phi$ graph of artifically compacted sediments. It is clear from Figures 1 and 3 that many samples can be found for which the observed points lie close to the Wood equation even at low porosity. Observations suggest that initial distribution on the V - $\phi$ plot is governed by the Wood equation but that subsequent compaction and cementation have the effect of displacing the points upward along the direction of the curves shown in Figure 3.

The occurrence of measured velocities less than water velocity is now well documented. Katz and Ewing (1956) found evidence of a low velocity layer at the top of the sediments on twelve seismic refraction stations in the Atlantic Ocean Basin west of Bermuda. Similar conclusions were reported by Officer (1955) on the basis of wide angle reflection studies. Hamilton (1956) and Sutton, et al (1957) have observed velocities less than water velocity on recovered sediment samples. Such low velocities are not to be confused with much lower values that occur in gassy sediments. The effect is caused by initial loading of the fluid by particles of higher density in the high porosity range. Within the experimental error the reduction of velocity is just that predicted by the Wood equation. Lowest measured values are about 1.45 km/sec.

Porosity is clearly a major factor in determining compressional velocity as Figures 1 and 3 indicate. The scatter about the main trend may be partially understood by examination of the effects of other variables. Sutton, et al (1957) carried out a detailed study of a group of cores from a variety of environments in the Atlantic Ocean. They measured median grain size, porosity, carbonate content, sorting coefficient, density, average grain density, and salt content with the aim of determining separately the influence of each variable on sound velocity. Using a multiple regression analysis involving as variables, porosity, carbonate content, and median grain size, they found velocity to increase with decreasing porosity and increasing grain diameter and carbonate content. Shumway (1960) in a study of a group of samples from the Pacific and Arctic Oceans found a positive connection between velocity and the porosity, grain size and phi deviation measure. Unlike Shumway's samples those described by Sutton, et al, showed no strong correlation between porosity and grain size. Thus the effects of grain size and porosity on velocity could be clearly separated.

In the high velocity region of Figure 3 it should be noted that limestones lie for the most part above 4.5 km/sec and sandstones between 3.5 and 5 km/sec. The choice of 6 km/sec for the upper limit of the plotted curves is arbitrary. As the density data will show, some observations on limestones and dolomites occur at velocities as high as 6.5 to 7 km/sec. The dependence of measured velocities on temperature and pressure has been discussed by Shumway (1956) and Sutton, et al (1957). The change with temperature results almost entirely from the change of velocity of the water fraction. Change with temperature, for a deep water sample raised to the surface is approximately the same, though opposite in sign, as the change with pressure. Since bottom and surface water velocities differ by about 1% in situ velocity should differ from that measured on deck by about the same amount.

C. Compressional Velocity-Density

Velocity-density data for sediments and sedimentary rocks are shown in Figure 4. Included in the figure are results for most of the samples plotted on the velocity-porosity figures. Independent measurements of density were made for most of the ocean sediments occurring in the low velocity range and for most of the values at very high velocity. Points representing artificially compacted sediments have been derived from measurements of volume of expressed water together with a measurement of initial density. The curve drawn through the points is intended only as a fair indication of the main trend and one that may serve to provide estimates of density from velocity in the absence of any other information. Cracking of samples or the presence of small amounts of gas in samples of ocean sediments will have the effect of reducing velocity for a given density and anhydrites have exceptionally high densities for a given velocity. In drawing the curve of Figure 4, therefore, relatively less weight was given to observations for which velocity reduction or density increase from such causes is a possibility.

All points plotted in Figure 4 represent observations on water-saturated materials. Those published numbers for which the degree of saturation was not indicated have been omitted.

It is interesting to note that when data for igneous and metamorphic rocks are plotted together with those for sedimentary rocks the velocity density curve may be continued upward to higher velocities with no noticeable offset, though the curve has a slight decrease in slope.

D. Shear Wave Velocities and Poisson's Ratio

The information that exists on shear wave velocities in marine sediments consists mainly of direct measurements by Laughton (1957) on artificially compacted sediments, the occurrence reported by Nafe and Drake (1957) of refracted arrivals consistent with the supposition that they propagated over part of their path as shear waves, and evidences of Oliver and Dorman that higher mode Rayleigh and Love wave dispersion is affected by sediments having a shear wave velocity of .25 to .60 km/sec. Hamilton has computed values of rigidity by attributing the discrepancy between observed compressional velocities and those predicted by the Wood equation to arise entirely from rigidity. Actually an increment to incompressibility may also be a contributing factor so that Hamilton's values are in fact upper limits to the rigidity. The available information is summarized in Figure 5 on a graph of Poisson's ratio plotted against porosity. Poisson's ratio is related to shear velocity through Equation 5. The solid curves were given by Nafe and Drake (1957). They represent an estimated range based on an approximate theory of the compaction process. The three solid points were derived by Laughton (1957) from his measurements of compressional and shear velocities in compacted globigerina ooze. Points obtained by Hamilton (1956) are indicated on the figure and are upper limits.

The remaining points are measurements on sedimentary rocks by Evison (1956), Hughes and Jones (1951, Birch, et al (1942) and others. The range of average sedimentary shear velocities required by Oliver and Dorman to fit their dispersion data is an indication that estimates of $\sigma$ by Nafe and Drake were too low for a given porosity. It is also likely that Laughton's points are higher than would be found for sediments in situ where cementation can occur. Thus the dashed curve may represent an approximate upper limit except at low porosities. Seismic refraction data normally yield values of $\sigma$ higher than 0.25. Limestones at low porosity tend to group around 0.30. Evison (1956) has measured a value of $\sigma$ for concrete of 0.20 at low porosity. Thus considerable scatter may be expected.

The degree of water saturation is especially important in determining $\sigma$ for unsaturated sedimentary rocks actually show a decrease of $\sigma$ with increasing porosity rather than increase as do saturated rocks.

E. Thermal Conductivity-Porosity

Observations of thermal conductivity for ocean sediments and sedimentary rocks are plotted against porosity in Figure 6. They are compared with computations of Parasnis (1960) for thermal conductivity of mixtures of particles and fluid for two convenient choices of particle shape. The results are applied in one case to the particles of quartz and in the other to calcite. While shape is an important factor there appears also a clear separation of materials according to composition as is shown by the data for sandstones and limestones.

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Conductivities of ocean sediments are principally affected by the water content, as has been pointed out by Ratcliffe (1960) and Bullard, et al (1956).

F. Selected Values of Other Measured Physical Properties

Table III provides a selection of values of physical properties for ocean sediments from a number of different marine environments measured by several different methods. These numbers are not necessarily typical. The table has been constructed to give at least one value of each quantity listed, to illustrate measurements at high and low porosity, to compare shallow water and deep water environments to show changes with frequency and to provide some indication of results that may be expected for different sediment types. Numbers in brackets are not directly measured but are derived through one or another of the formulae in Table II. Although most of the observations were made on marine sediments a few entries for non-marine sediments have been added at the bottom of the table to illustrate unusual situations such as inclusion of gas or to provide estimates of probable values where actual measurements on marine sediments are few or lacking altogether. It is particularly important to assemble information of shear wave velocities.

The only values of impedance listed are those of Hamilton, et al, (1956). These are not measured but are derived by taking the product $\rho$ V. It is probably better to find impedance by taking this product than to derive it from a measured reflection coefficient for measured reflection coefficients are almost invariably complicated by the occurrence of multiple reflections with a consequent increase in the apparent value of the impedance. The error so introduced is much more serious below about 1000 cycles/sec than it is above that frequency.

Velocities derived from critical angle reflections and sub-bottom reflections (Katz and Ewing, 1956) doubtless apply not to sediments close to the water sediment interface but to those at some distance below. On the same refraction profiles leading to the tabulated numbers there was clear cut evidence for the existence of a layer close to the interface with a velocity less than water velocity.

The frequency dependence of the attenuation coefficient $\gamma$ is of particular interest. Shumway (1960b) has reported $\gamma$ to be proportional to $f^{1.79}$ for sixty-five samples with a standard deviation of 0.98 for the exponent. Measurements were made in the frequency range 20 to 40 kc/sec and attenuations apply to compressional waves. At much lower frequencies, though not for ocean sediments, McDonal, et al (1958) have observed approximately linear dependence of attenuation on frequency. The attenuation in db per thousand feet being 1.05 f for horizontally traveling shear waves from 20 to 152 cycles per second; 0.12 f for vertically traveling compressional waves from 50 to 450 cycles per second. Their measurements were made in boreholes in the Pierre Shale. Since the exponent may depend on the degree of water saturation (Born, 1941) this example is introduced mainly to illustrate the fact that attenuation for shear waves may differ widely from that for compressional waves.

It will be noted that in not one single instance have measurements been made of all the quantities of interest. In fact, it is frequently the case that in publication of results no clear indication is given of the degree of water saturation though all evidence points to degree of saturation and water fraction as the most significant variables in determining physical properties. All quantities listed for ocean sediments apply to cases of 100% saturation.

G. Depth Dependence of Compressional Velocity

Evidence for depth variation of compressional velocity in marine sediments is derived from reflection and refraction measurements. Arrivals reflected at a sub-bottom interface or refracted in the upper part of the sedimentary column commonly show the effects of gradients ranging from 0.5 to 2 $sec^{-1}$. Hill (1952) observed gradients of the order of 2 $sec^{-1}$ in the Eastern Atlantic. Officer (1955) reported reflections at a location northeast of Bermuda that indicated an average gradient of about 1.0 $sec^{-1}$. Katz and Ewing (1956) reported gradients of about 0.6 $sec^{-1}$ at a number of stations in the Western Atlantic. Evidence from both sub-bottom reflections and from refracted arrivals was summarized by Nafe and Drake (1957). In Figure 7 are plotted observations of velocity against thickness of sediment overlying the refracting horizon for a number of stations in the Western Atlantic. Points of the figure are not distinguished as to water depth. The only discernible pattern is shown by comparison of the measurements with velocities found by Laughton (1954) for artificially compacted globigerina ooze. Laughton's velocity-depth curve, based on an assumed parabolic gradient is an effective lower limit to the points measured by the refraction method. This is to be expected, for processes of cementation occurring over an extended time interval should have the effect of increasing the velocity.

In Figure 8 points representing observed velocities are distinguished according to locality. The observations from deep water stations appear to have velocities distinctly lower than those from shallow water stations with the same depth of burial. It should be pointed out that most shallow water observations in Figure 8 are from the continental shelf of northeastern America. Shallow water environments with velocity-depth curves similar to that for deep water sediments are known to exist. The equations given in the figure are not significant in themselves. They are simply convenient representations of the main trends of the data. Velocities observed on the Blake Plateau, east of Florida are higher than those at corresponding depths of burial elsewhere in the shallow waters of the North Atlantic. This is probably a consequence of the high carbonate content of the sediments of the Blake Plateau and is consistent with the observation that for a given porosity sediments of high carbonate content have higher velocities than those of low carbonate content. Figure 8 suggests as does the velocity porosity data, that the zero porosity limit for the velocity of commonly occurring sediments is near 6 km/sec. Additional evidence for the existence of this approximate velocity limit may be seen in the trend of deep water points at depths of burial greater than about 1.5 km. The trend to much higher velocities below 1.5 km may be a consequence of loading of the sediments by the water column when water filled interstices are isolated from each other and do not have free access to the sea.

Nafe and Drake (1957) have suggested that in deep water, the principal contribution to the increase of velocity with depth is that of compaction alone. Many other processes can occur in a shallow water environment such as circulation of water, deposition of solid material in the interstices between particle grains, and erosional unloading. All would tend to increase the compressional velocity. Thus in a shallow water environment velocity might well increase at a greater rate with depth than in deep water. It is unlikely that the rate of increase could be less.

## V. CONCLUSIONS

Consideration of available observations together with data on sedimentary rocks suggests that a known value of one quantity may limit the range within which values of other quantities are likely to be found. Estimation of such limits may be useful in determining the consistency of results of measurements of different kinds. For example, densities to be used in gravity interpretation have been assigned in accordance with seismic measurements of velocity. However useful, estimation is not measurement, and much more detailed information is needed on marine sediments. The close connection between density and porosity seems well established. Very little is known about shear velocity or about attenuation at low frequency. Compressional velocities are easily measured for near surface samples but direct measurements of velocity as a function either of density or porosity at points well within the compacting sedimentary column are needed. Artificial compaction experiments suggest that the trend of observations will be roughly that of the $n = 4$ to $n = 6$ curves of Figure 3.

The value of a physical quantity is in a sense a function of the use for which that quantity is required. A static determination of Young's modulus, as Evison (1956) has pointed out, may differ widely from the Young's modulus obtained dynamically through substituting into equation 4 constants found by measuring compressional and shear velocities. Indeed the dynamic value of Young's modulus found by timing compressional and shear velocities over 100 feet of sediment in situ may differ widely from that obtained by measurements on a hand sample in the laboratory. If interpretation of seismic data is to provide information on sedimentary structure, then the physical constants most needed are dynamic values and values determined by relatively large scale measurement.

## Acknowledgments

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TABLE I

| Property | Symbol | Unit |
|---|---|---|
| Compressional velocity | $V_\rho$ or V | km/sec |
| Shear velocity | $V_s$ | km/sec |
| Compressibility | C | $(dynes/cm^2)^{-1}$ |
| Incompressibility | k | $(dynes/cm^2)$ |
| Rigidity | $\mu$ | $(dynes/cm^2)$ |
| Young's modulus | E | $(dynes/cm^2)$ |
| Poisson's ratio | $\sigma$ | dimensionless |
| Attenuation | $\gamma$ | $(length)^{-1}$ or Db/length |
| Characteristic impedance | $Z = \rho V$ | $gm/cm^2\ sec$ |
| Density | $\rho$ | $gm/cm^3$ |
| Porosity | $\phi$ | dimensionless |
| Thermal conductivity | K | cal/cm C° sec |

## TABLE III

| Sample or Station | Location | Water Depth Fathoms | Method | $V_p$ Km/Sec | $V_s$ Km/Sec | $\mu$ $10^{-11}$ dynes/cm$^2$ | $\kappa$ $10^{-11}$ dynes/cm$^2$ | $\sigma$ | $\gamma$ Db/ft | $f$ cycles/sec | gm | $V_s$ m/Sec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **(A) Marine Sediment** | | | | | | | | | | | | |
| Grey clay | Abyssal Pl. | 2570 | Core | 1.57 | | | | | | | | |
| Tan & Grey clay | " | 2570 | " | 1.51 | | | | | | | | |
| Cream Cal-carinite | Sea Mount | 940 | " | 1.70 | | | | | | | | |
| Red arena-ceous clay | Continental Slope | 820 | " | 1.61 | | | | | | | | |
| Fine Sand | Shallow Water | <15 | in situ | 1.68 | | (0.06) | (0.472) | (0.44) | | | (3. | |
| Very Fine Sand | " " | " | " | 1.66 | | (0.05) | (0.458) | (0.45) | | | (3. | |
| Silty Very Fine Sand | " " | " | " | 1.55 | | | | | | | (2. | |
| Medium Silt | " " | " | " | 1.47 | | (0.00) | (0.356) | (0.50) | | | (2. | |
| Clayey Fine Silt | " " | " | " | 1.46 | | (0.00) | (0.332) | (0.50) | | | (2. | |
| Coarse Silt | " " | < 30 | resonance | 1.514 | | | | | 2.202<br>3.811 | 23,214<br>30,965 | | |
| Medium Sand | " " | " | " | 1.785 | | | | | 5.72 | 31,406 | | |
| " " | " " | " | " | 1.75 | | | | | 3.65 | 23,335 | | |
| Fine Sand | " " | " | " | 1.550 | | | | | 6.27 | 23,943 | | |
| " " | | | | 1.773 | | | | | 4.43 | 31,239 | | |
| Silty Coarse Clay | Continental Slope | 560 | " | 1.491 | | | | | 0.435<br>0.401<br>1.315 | 20,500<br>22,768<br>30,225 | | |
| Clayey Med. Silt | Cont. Border Land | 100 | " | 1.506 | | | | | 0.569<br>1.364 | 23,068<br>30.655 | | |
| Glob. Ooze | | 2577 | Compaction | 1.62 | | | | | | | | .20 |
| " | | " | " | 2.68 | 1.20 | (0.31) | (1.13) | (0.37) | | | | .42 |
| " | | " | " | 2.69 | 1.42 | (0.45) | (1.15) | (0.34) | | | | .57 |
| " | | " | " | 3.06 | 1.57 | (0.56) | (1.38) | (0.32) | | | | |
| Terrig. Mud | | 1140 | " | 1.51 | | | | | | | | |
| " " | | " | " | 1.83 | | | | | | | | |
| " " | | " | " | 2.46 | | | | | | | | |
| A179-8 | Abyssal Plain | 2830 | Sub bottom refl | 1.7-2.0 | | | | | | | | |
| A179-8 | " " | 2830 | Crit Angle | 1.88 | | | | | | | | |
| A179-9 | " " | 2730 | Sub bottom refl | 1.8-2.3 | | | | | | | | |
| A179-9 | " " | 2730 | Crit Angle | 1.81 | | | | | | | | |
| G 8 | N.of Tobago Trough | 960 | Refraction | 1.68 | | | | | | | | |
| br.Zeolitic Clay | | 2440 | Needle probe | | | | | | | | | |
| Clayey Calc. Ooze | | 1625 | " " | | | | | | | | | |
| Buff Calc. Ooze | | 1680 | " " | | | | | | | | | |
| Clayey calc. Ooze | | 1440 | " " | | | | | | | | | |
| Calc Silicious Ooze | | 2080 | " " | | | | | | | | | |
| **(B) Non Marine Sediments** | | | | | | | | | | | | |
| Organic silt | Hudson River | 2 | Refraction | 0.538 | | | | | | | | 80 |
| Ignimbrite | " " | - | Direct Timing | 2.08 | 0.80 | | | (0.41) | | | | 83 |
| Pierre Shale | | - | Refraction | 2.28 | 0.85 | | | (0.42) * | | | | |
| " " | | - | Refraction | 2.36 | 0.86 | | | (0.42) * | | | | 86 |

† $\varphi$ = - $\log_2$ (grain diameter in millimeters)

* Over 50 to 450 cycle/sec range = 0.12 f for vertically traveling compressional waves in Db per thousand feet. velir
and over 20 to 125 cycles/sec = 1.0 f for horizontally traveling shear waves in Db per thousand feet. aveli

TABLE III

| Vs m/Sec | μ $10^{-11}$ dynes/cm$^2$ | κ $10^{-11}$ dynes/cm$^2$ | σ | γ Db/ft | f cycles/sec | ρV $10^{-5}$ gm/cm$^2$Sec | ρ gm/cm$^3$ | φ | K $10^{-3}$ Cal/C°cmSec | Med. Grain size φ units † | $CO_3$ % | References |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | | | 1.62 | 0.61 | | 7.2 | 23.8 | Sutton et.al.1957 |
| | | | | | | | 1.52 | 0.68 | | 8.6 | 25.0 | " |
| | | | | | | | 1.62 | 0.61 | | 4.1 | 94.8 | " |
| | | | | | | | 1.77 | 0.55 | | 7.3 | 68.4 | " |
| | (0.06) | (0.472) | (0.44) | | | (3.25) | 1.93 | 0.462 | | 2.5 | | Hamilton et.al 1956 |
| | (0.05) | (0.458) | (0.45) | | | (3.18) | 1.92 | 0.477 | | 3.0 | | " |
| | | | | | | (2.60) | 1.68 | 0.613 | | 4.5 | | " |
| | (0.00) | (0.366) | (0.50) | | | (2.49) | 1.69 | 0.609 | | 5.0 | | " |
| | (0.00) | (0.332) | (0.50) | | | (2.34) | 1.60 | 0.656 | | 5.5 | | " |
| | | | | 2.202 | 23,214 | | 1.62 | 0.604 | | 4.2 | | Shumway 1960 |
| | | | | 3.811 | 30,065 | | | | | | | " " |
| | | | | 5.72 | 31,406 | | 1.94 | 0.432 | | 2.00 | | " " |
| | | | | 3.65 | 23,335 | | 1.99 | 0.362 | | 1.75 | | " " |
| | | | | 6.27 | 23.943 | | 1.80 | 0.526 | | 2.68 | | " " |
| | | | | 4.43 | 31,239 | | 1.89 | 0.479 | | 2.62 | | " " |
| | | | | 0.435 | 20,500 | | 1.32 | 0.812 | | 8.05 | | " " |
| | | | | 0.401 | 22,768 | | | | | | | " " |
| | | | | 1.315 | 30,225 | | | | | | | " " |
| | | | | 0.569 | 23,068 | | 1.594 | 0.651 | | 5.60 | | " " |
| | | | | 1.364 | 30.655 | | | | | | | " " |
| | | | | | | | 1.59 | 0.65 | | | 54 | Laughton 1957 |
| .20 | (0.31) | (1.13) | (0.37) | | | | 2.14 | 0.32 | | | | " " |
| .42 | (0.45) | (1.15) | (0.34) | | | | 2.22 | 0.28 | | | | " " |
| .57 | (0.56) | (1.38) | (0.32) | | | | 2.26 | 0.26 | | | | " " |
| | | | | | | | 1.54 | 0.69 | | | 21 | " " |
| | | | | | | | 2.13 | 0.34 | | | | " " |
| | | | | | | | 2.41 | 0.17 | | | | " " |
| | | | | | | | | | | | | Katz & Ewing 1956 |
| | | | | | | | | | | | | " " |
| | | | | | | | | | | | | " " |
| | | | | | | | | | | | | " " |
| | | | | | | | | | | | | Ewing et.al.1956 |
| | | | | | | | 1.19 | (0.88) | 1.63 | | | VonHerzen & Maxwell 19[illegible] |
| | | | | | | | 1.50 | (0.71) | 2.33 | | | " |
| | | | | | | | 1.45 | (0.73) | 2.35 | | | " |
| | | | | | | | 1.39 | (0.77) | 2.14 | | | " |
| | | | | | | | 1.50 | (0.70) | 2.07 | | | " |
| | | | | | | | 1.35 | 0.87 | | | | Drake |
| 80 | | | (0.41) | | | | 1.95 | | | | | Evison 1956 |
| 85 | | | (0.42) | | | | | | | | | McDonal et al 1958 |
| 86 | | | (0.42) | | | | | | | | | " |

...veling compressional waves in Db per thousand feet.
...aveling shear waves in Db per thousand feet.

1

2 2

TABLE II

| Equation | Applicability | Reference |
|---|---|---|
| 1. $V = \sqrt{\frac{k + 4/3\mu}{\rho}}$ | isotropic elastic solid | Bullen (1947) |
| 2. $V = \sqrt{\frac{\mu}{\rho}}$ | isotropic elastic solid | Bullen (1947) |
| 3. $\left(\frac{V}{V}\right)^2 = \frac{2(1-\sigma)}{(1-2\sigma)}$ | isotropic elastic solid | Bullen (1947) |
| 4. $E = 3k(1-2\sigma)$ | isotropic elastic solid | Bullen (1947) |
| 5. $Q = \pi / \gamma cT$ | plane waves c = phase velocity T = period | Birch, et al (1942) |
| 6. $\rho = \sum_i f_i \rho_i$ * | general | Shumway (1960) |
| 7. $C = \sum_i f_i C_i$ * | emulsion or suspension | Shumway (1960) |
| 8. $V^2 = 1/\rho C$ ** | emulsion or suspension (Wood's equation) | Wood (1941) |
| 9. $1/V = \sum_i f_i / V_i$ | transmission perpendicular to a layer sequence (time average equation) | Wyllie, et al (1956) |

* $f_i$ = volume fraction of ith constituent († water = Ø)

** $\rho$ and C from equations 6 and 7 respectively

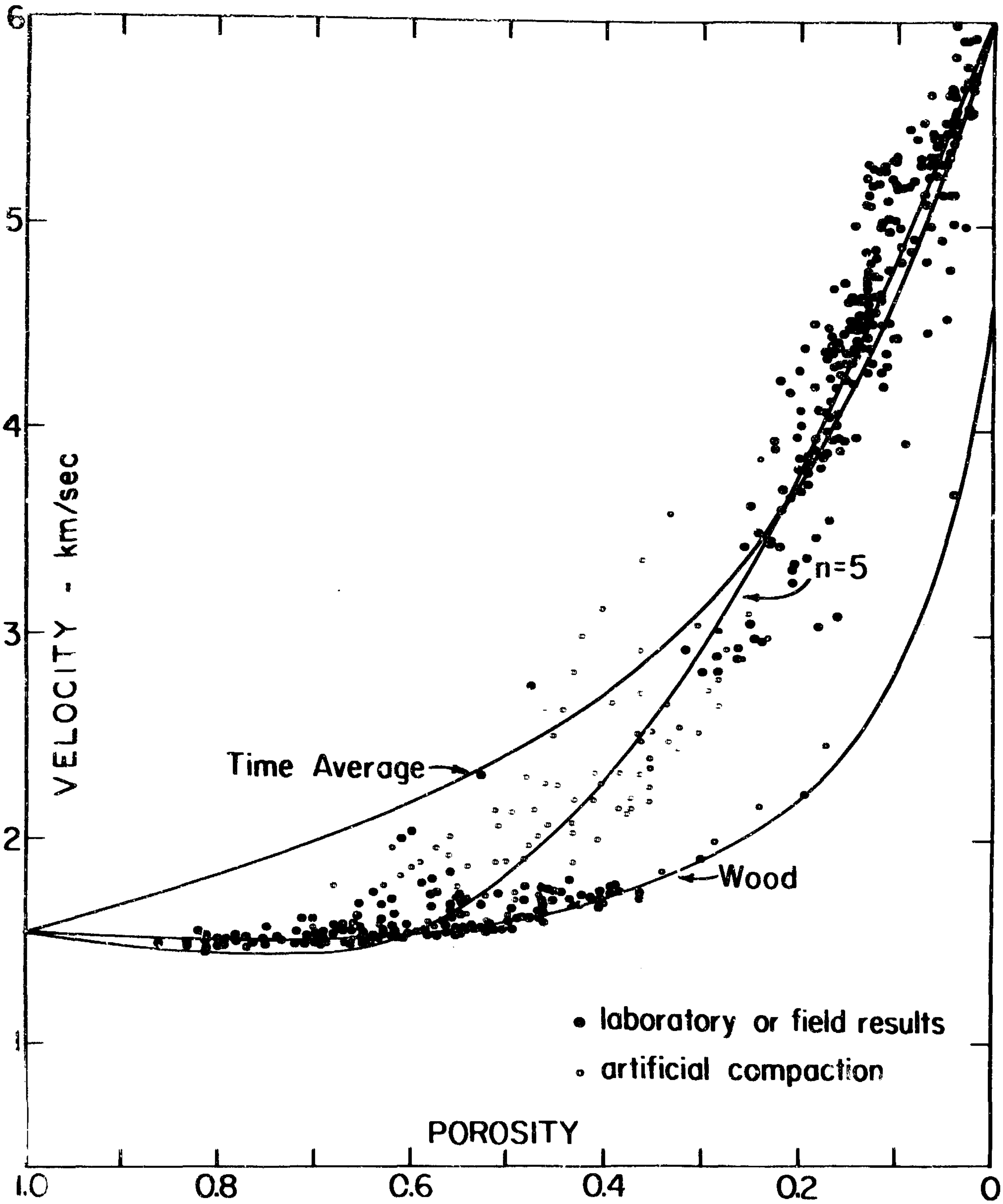


**Figure** **Observations of compressional velocity as a function of porosity. The data are compared with the Wood equation, the time average equation and an empirical equation of Nafe and Drake. Solid points at low velocity represent measurements on ocean sediments.**

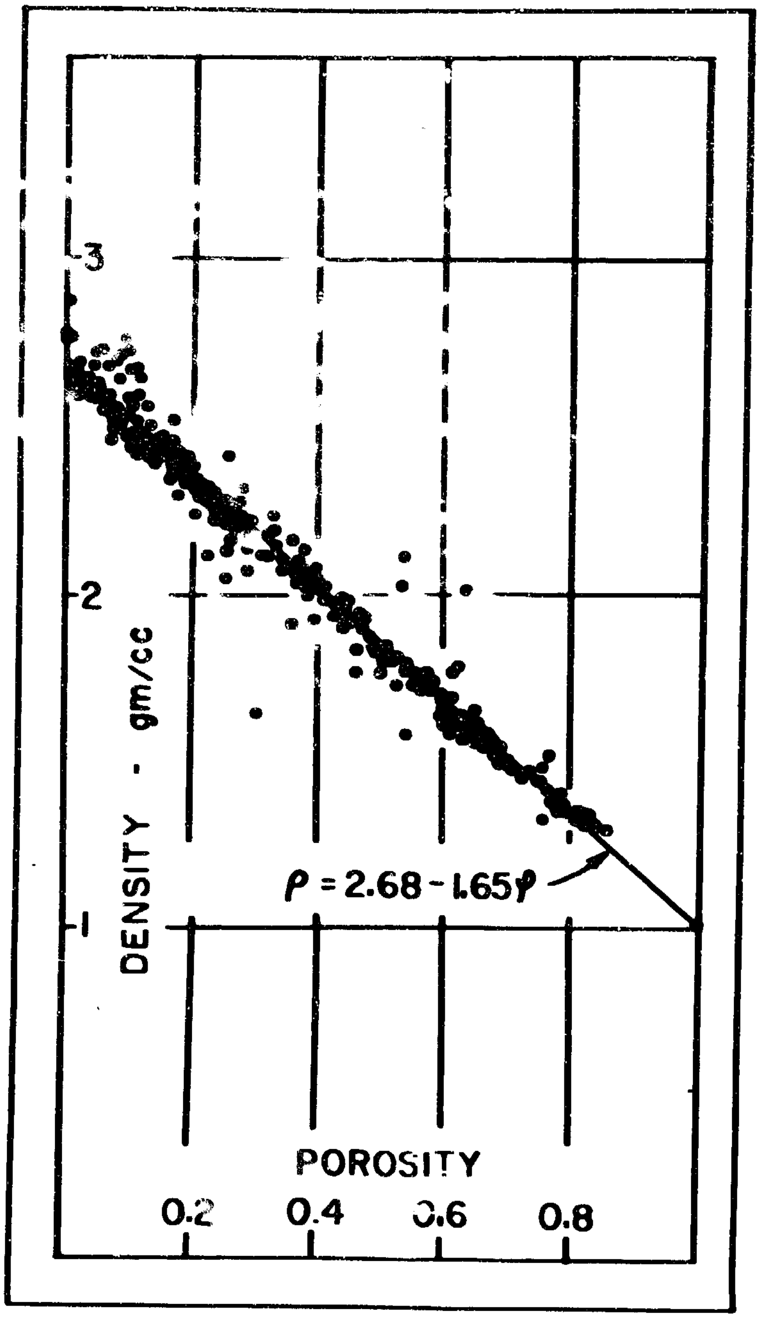


Figure 2 **Porosity-density data for sediments and sedimentary rocks. Measurements on ocean sediments range from porosities of about 0.40 to 0.85.**

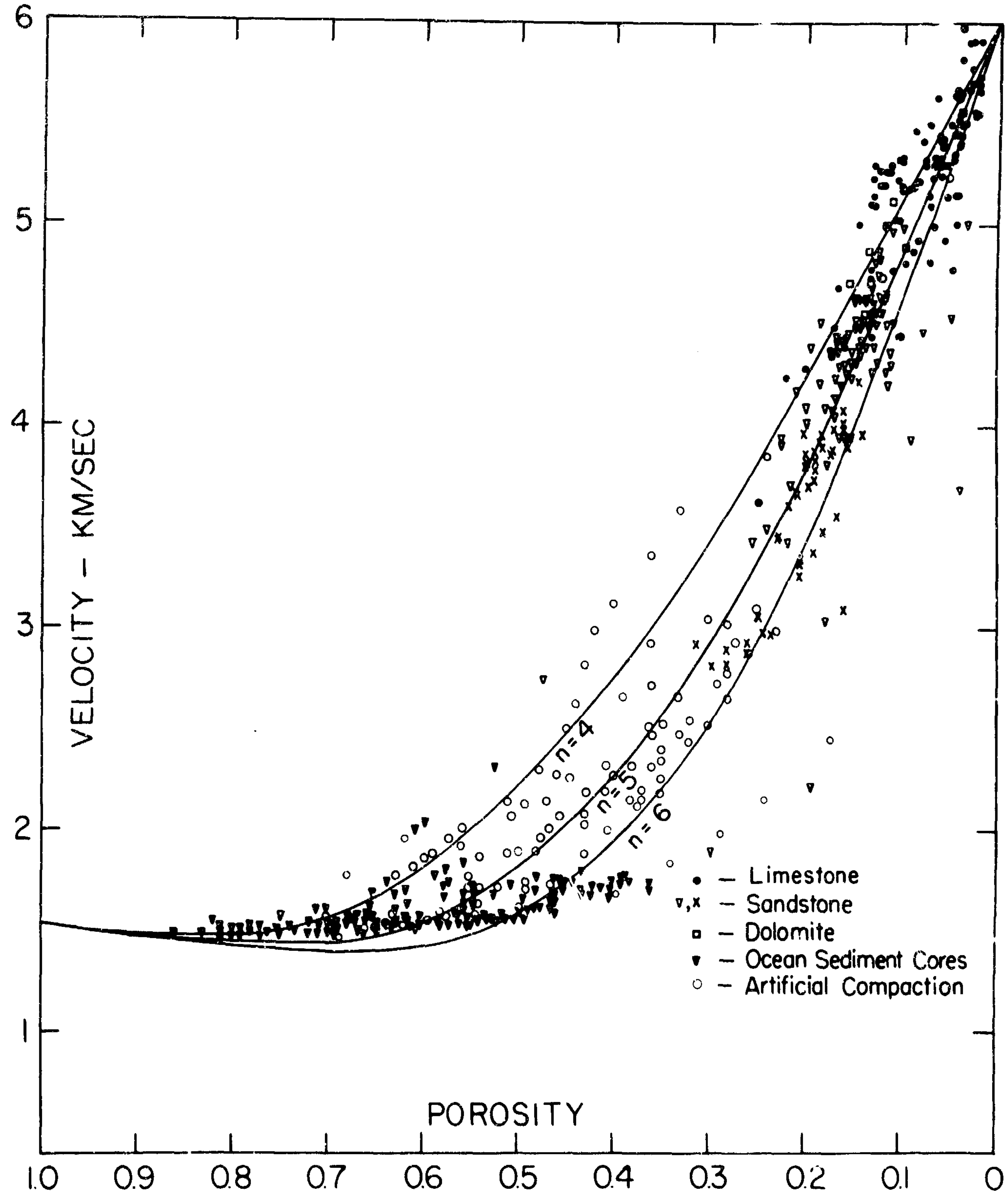


**Figure 3** **Compressional velocity-porosity data for sediments and sedimentary rocks compared with empirical equations of Nafe and Drake.**

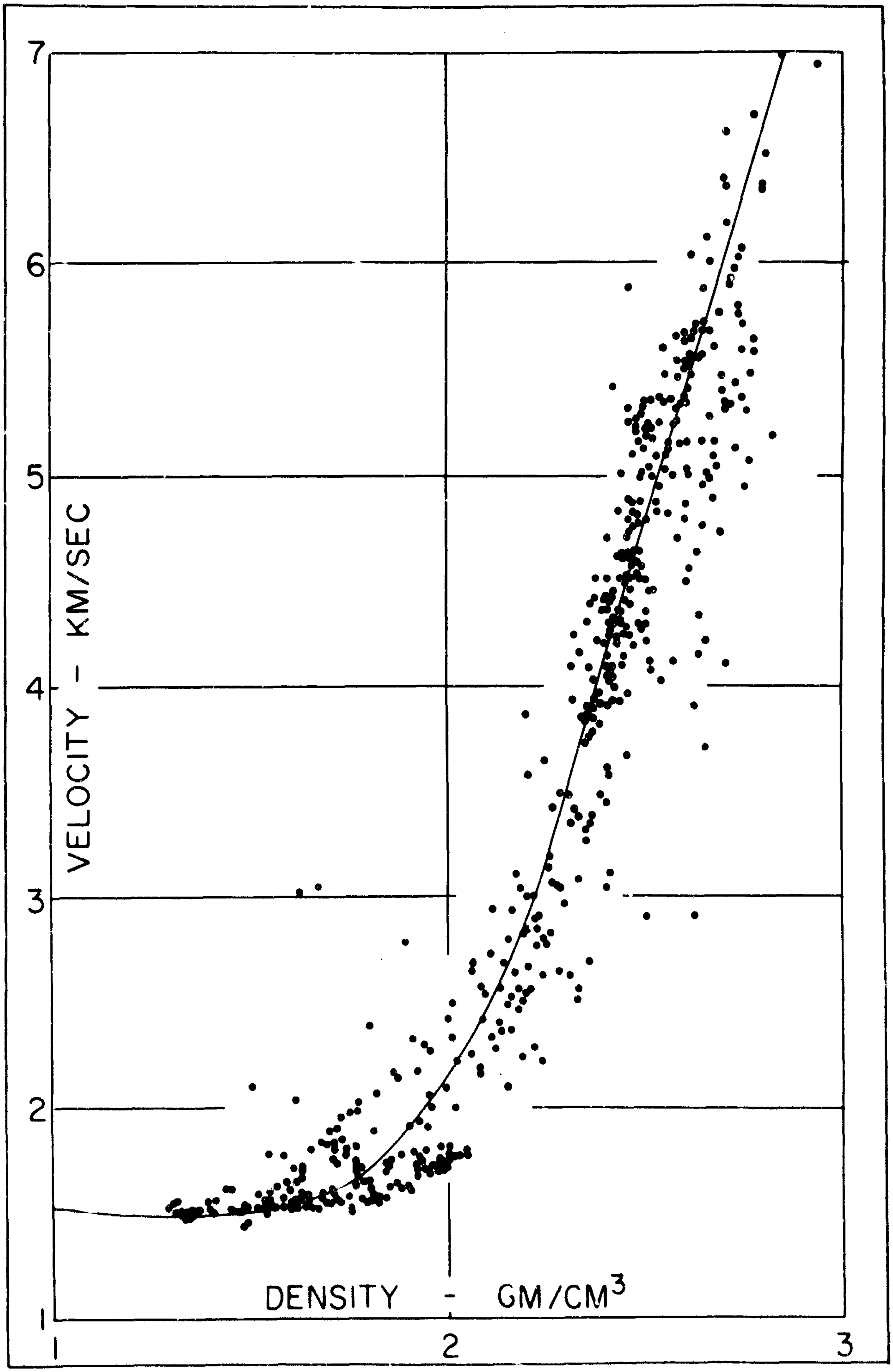


Figure 4 Compressional velocity as a function of density. Most observations on ocean sediments lie below 2 km/sec in velocity.

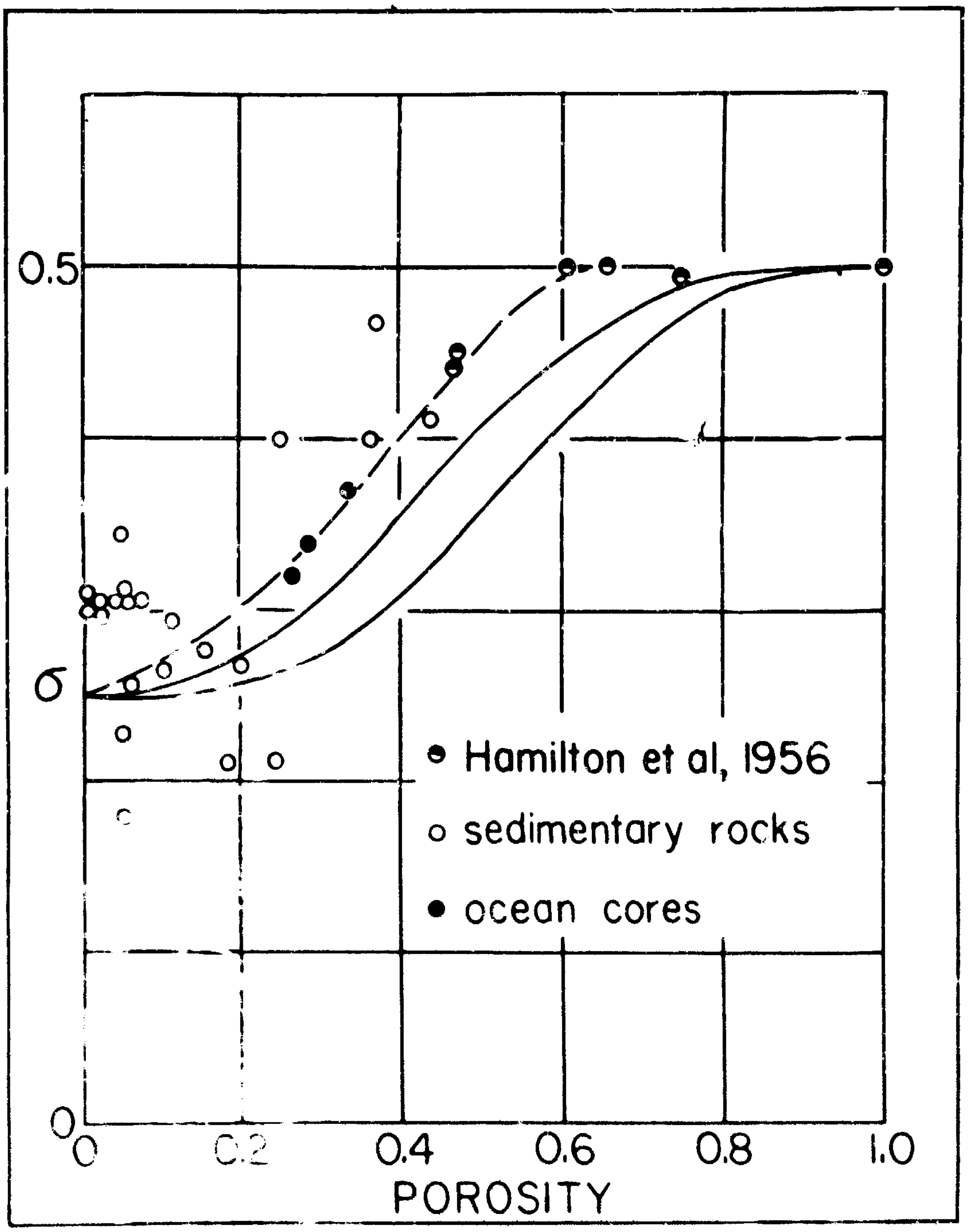


Figure 5 Poisson's ratio plotted against porosity. Hamilton's values are upper limits to the ratio. The solid points are those reported by Laughton (1957) for artificially compacted globigerina ooze. The region outlined by solid curves represents an estimate by Nafe and Drake (1957) of the range of values for ocean sediments.

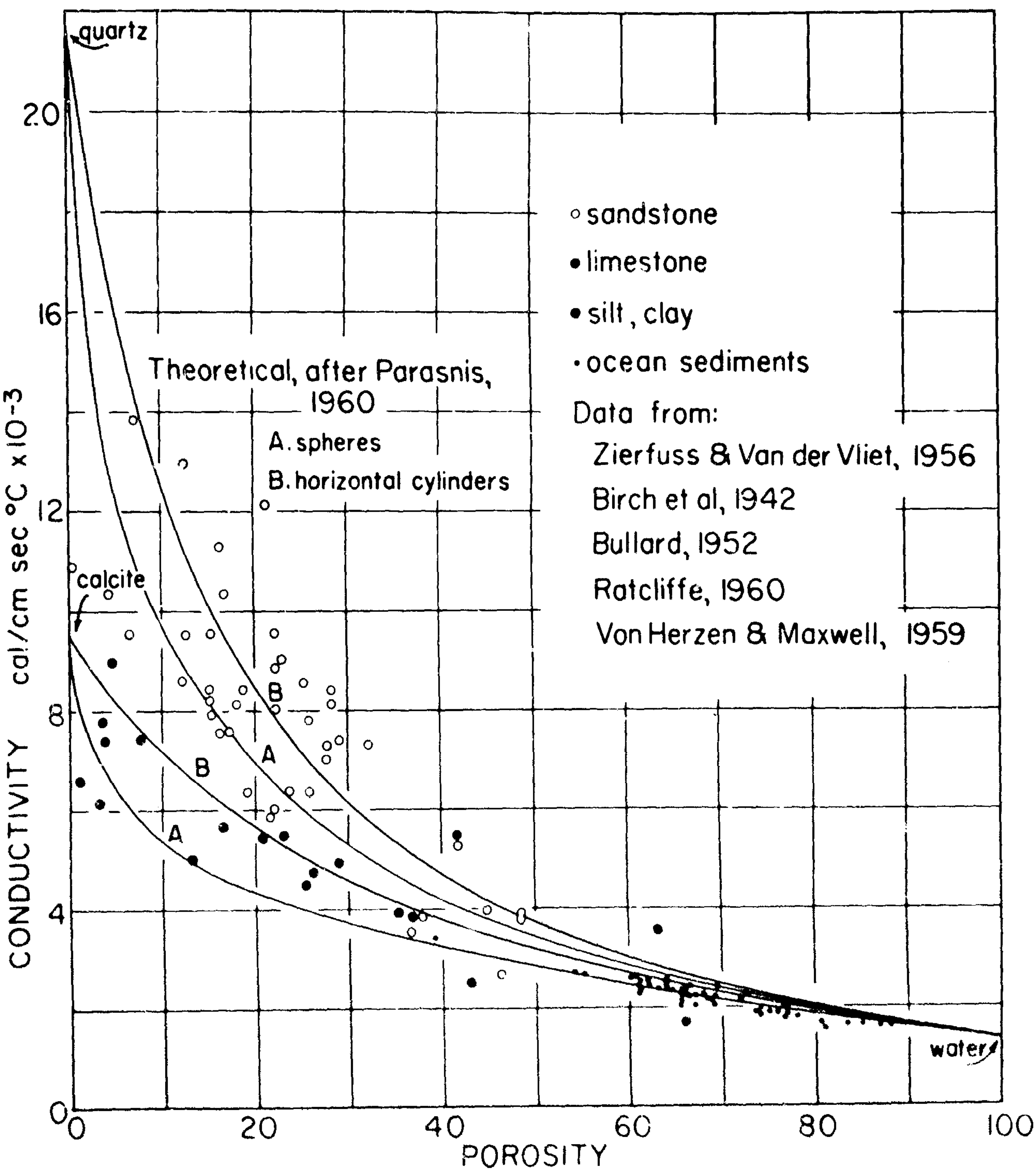


Figure 6 Thermal conductivity-porosity data for ocean sediments, sandstones and limestones compared with theoretical curves adapted from Parasnis (1960).

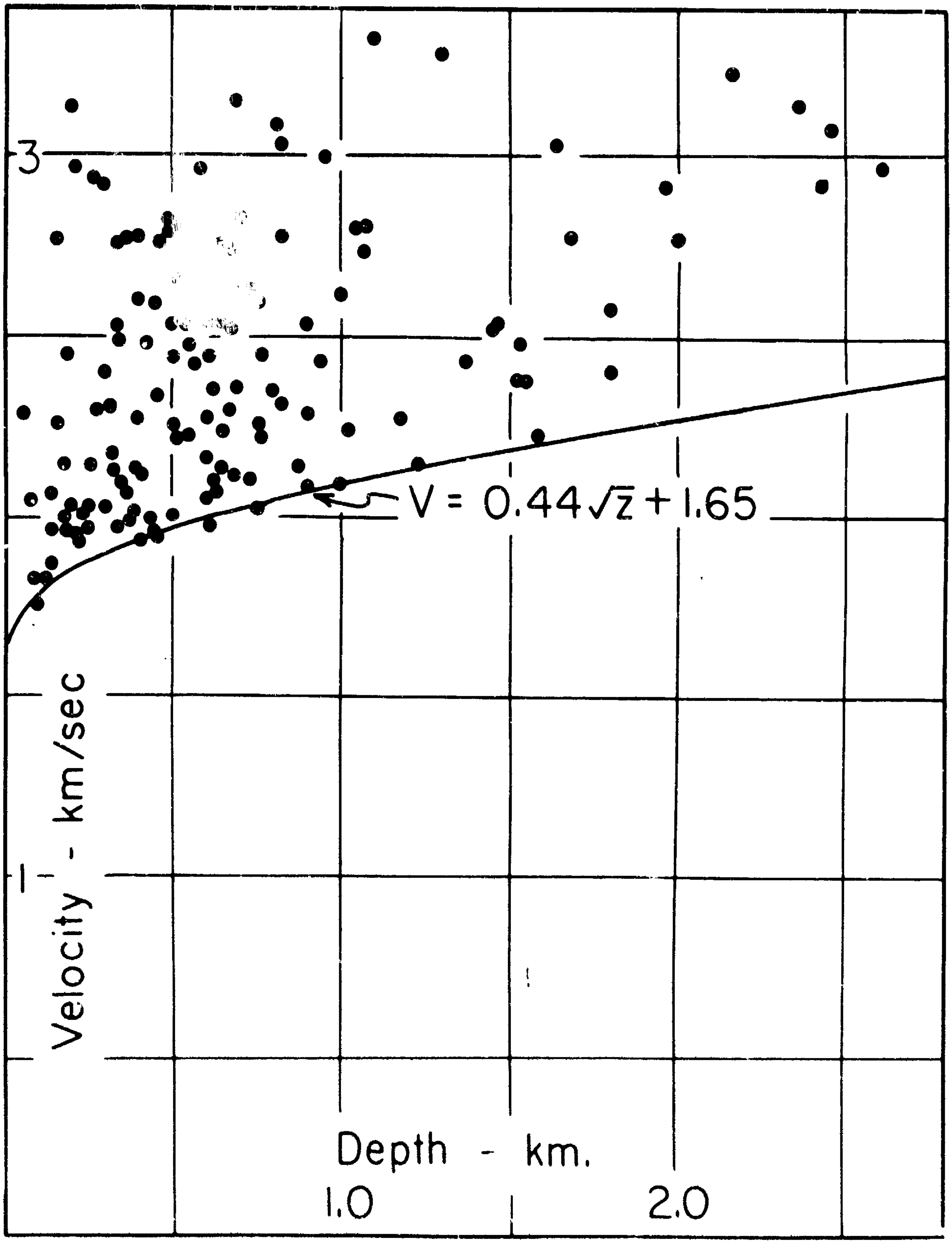


Figure 7 Compressional velocity as a function of depth of burial. Data were obtained from seismic refraction measurements. The solid curve was derived by Laughton (1954) from the artificial compaction of globigerina ooze. It is a rough lower limit to the observations.

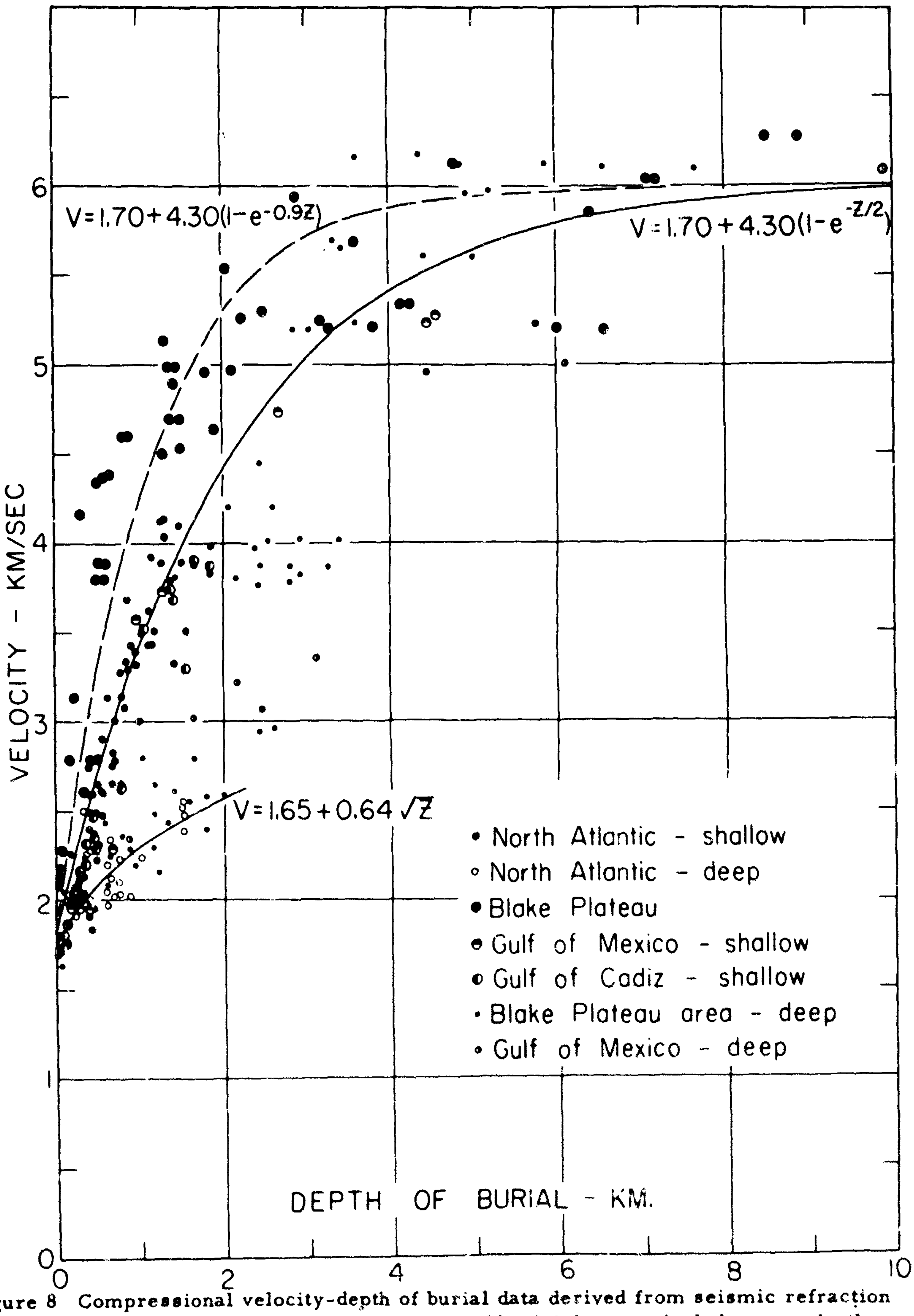


Figure 8 Compressional velocity-depth of burial data derived from seismic refraction measurements. The indicated depth of burial does not include water depth. The slowest rate of increase of velocity with depth occurs in the case of deep water sediments (water depth > 2000 fathoms).