

## FORMATION VELOCITY AND DENSITY—THE DIAGNOSTIC BASICS FOR STRATIGRAPHIC TRAPS

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A multiplicity of factors influence seismic reflection coefficients and the observed gravity of typical sedimentary rocks. Rock velocity and density depend upon the mineral composition and the granular nature of the rock matrix, cementation, porosity, fluid content, and environmental pressure. Depth of burial and geologic age also have an effect.

Lithology and porosity can be related empirically to velocity by the time-average equation. This equation is most reliable when the rock is under substantial pressure, is saturated with brine, and contains well-cemented grains. For very low porosity rocks under large pressures, the mineral composition can be related to velocity by the theories of Voigt and Reuss.

One effect of pressure variation on velocity results from the opening or closing of microcracks. For porous sedimentary rocks, only the difference between overburden and fluid pressure affects the

microcrack system. Existing theory does not take into account the effect of microcrack closure on the elastic behavior of rocks under pressure or the chemical interaction between water and clay particles.

The theory of Gassmann can be used to calculate the effect of different saturating fluids on the *P*-wave velocity of porous rocks. The effect may be large enough in shallow, recent sediments to permit gas sands to be distinguished from water sands on seismic records. At depths greater than about 6000 ft, however, the reflection coefficient becomes essentially independent of the nature of the fluid.

Data show the systematic relationship between velocity and density in sedimentary rocks. As a result, reflection coefficients can often be estimated satisfactorily from velocity information alone.

### INTRODUCTION

The purpose of this paper is to set forth certain relationships between rock physical properties, rock composition, and environmental conditions which have been established through extensive laboratory and field experimentation together with theoretical considerations. The literature on the subject is vast. We are concerned primarily with seismic *P*-wave velocity and density of different types of sedimentary rocks in different environments. These properties govern occur-

rences of seismic reflections and variations of observed gravity. They thus have significant bearing upon the manner of use of these geophysical methods and their effectiveness in finding or delineating stratigraphic traps.

A stratigraphic trap connotes a porous and permeable reservoir rock which alters laterally on one or more sides into a nonpermeable rock by facies changes or a pinch-out. A particularly important example is a reef surrounded by rock with different properties. Reflection seismic represen-

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tation of the convergence of seismic horizons bracketing pinch-outs may comprise very useful information. Also, relatively small lateral convergences in seismic transit times may reveal differential compaction, and lateral variation in transit time may indicate lateral change in the physical characteristics of the bracketed rock. Direct seismic reflection evidence, however, typically involves lateral variation of interference patterns of reflections from the top and bottom of a reservoir rock. Evidence furnished by gravity observations is even more subtle but can be applicable in some cases.

Many laboratory studies have been made to show how  $P$ -wave velocity in rocks is affected by pressure and fluid saturation; these studies principally use ultrasonic techniques. When the samples are cores recovered from wells, conditions that exist in the earth can be reproduced with a fair degree of realism. As a result, a number of significant relationships for  $P$ -wave velocities of rocks under different conditions of stress and fluid containment have been established.

A change in rock lithology or composition however cannot be simulated very satisfactorily in the laboratory. Consequently, no relevant relationships between different specific parameters are well established. Commonly then, we must resort to empirical correlations based on field data. Such correlations generally entail some unknowns, so they are satisfactorily applicable only for particular formations and environments.

An illustration of the wide range of  $P$ -wave velocities and lesser range of bulk densities for the more prevalent sedimentary rock types through a wide range of basins, geologic ages, and depths (to 25,000 ft) is given in Figure 1.

An additional consideration is a general tendency for velocity and density to increase with increase in depth of burial and with increase in age of formations as verified by Faust (1953). Successively deeper layers may differ materially in composition and porosity with accompanying marked local departures in velocity and density from progressive increase with depth.

An understanding of interrelationships between different rock properties and environmental conditions, however, requires recognition and consideration of the nature of rocks in general as being granular with interconnected fluid-filled

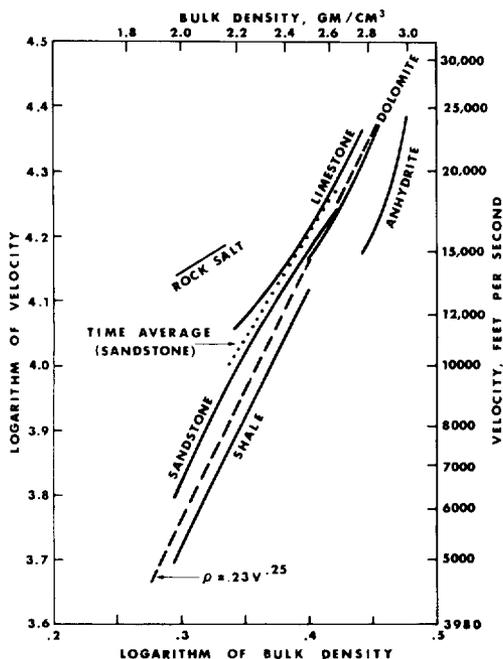


FIG. 1. Velocity-density relationships in rocks of different lithology.

interstices. As a consequence, porosity, mineral composition, intergranular elastic behavior, and fluid properties are primary factors. These factors are dependent upon overburden pressure, fluid pressure, microcracks, age, and depth of burial.

#### ROCK COMPOSITION AND GASSMANN'S THEORY

The three components which characterize the composition of rocks and the superscripts which indicate symbols of their properties are:

- (1) the solid matter of which the skeleton or frame is built (index  $\hat{\quad}$ );
- (2) the frame or skeleton (index  $\bar{\quad}$ ); and
- (3) the fluid filling of the pores (index  $\sim$ ).

Properties of the whole rock are indicated by symbols without superscripts.

In a classic paper, Gassmann (1951) showed that when a rock with its fluid is a closed system, grossly isotropic and homogeneous, the use of elementary elastic theory yields the following interrelationship between the rock parameters:

$$\bar{k} = \hat{k}(\bar{k} + Q)/(\hat{k} + Q), \quad (1)$$

where

$$Q = \hat{k}(\hat{k} - \bar{k})/\phi(\hat{k} - \bar{k}).$$

Definitions of the symbols are shown below.

NOMENCLATURE	
Symbol	Quantity
$M$	Space modulus (or $P$ -wave modulus)
$\mu$	Rigidity modulus (or $S$ -wave modulus)
$k$	Bulk modulus (or reciprocal of the compressibility)
$\nu$	Poisson's ratio
$\rho$	Bulk density
$\phi$	Fractional porosity
$\bar{P}$	Pressure on skeleton = Total external pressure less the internal fluid pore pressure = Net overburden pressure
<i>Superscripts</i>	
—	Properties of the frame or skeleton (empty porous rock)
^	Properties of the solid matter (grains or crystals) of which the skeleton is built
~	Properties of fluid occupying the pore space of the skeleton
<i>Parameters of Time-Average Equation:</i>	
$V$	Velocity of fluid-saturated rock
$V_F$	Velocity of fluid occupying the pore space of the rock
$V_M$	Velocity of solid mineral of which the matrix of rock is built

He also noted that  $\mu = \bar{\mu}$  and that  $M = k + 4/3 \mu$ . White (1965) gives these relationships in the form:

$$M = \bar{M} + (1 - \bar{k}/\hat{k})^2 / (\phi/\bar{k}) + (1 - \phi) / \bar{k} - \bar{k}/\hat{k}^2. \quad (1a)$$

It is also known that:

$$\rho = (1 - \phi)\hat{\rho} + \phi\bar{\rho}, \quad (1b)$$

and

$$P\text{-wave velocity} = \sqrt{M/\rho}. \quad (1c)$$

There is general agreement among experimenters that equations (1a) to (1c) satisfactorily predict the effect of different saturating fluids on  $P$ -wave velocity in most porous rocks, in spite of the obvious simplifications of the theory.

Gassmann's theory was extended by Biot (1956) to include the dynamic effects of relative motion between the fluid and the frame and also to take into account viscoelastic effects in part associated with the presence of microcracks. These refinements are unimportant for waves at seismic prospecting frequencies, and even at well-logging frequencies Gassmann's simple formula often is adequate.

Neither Gassmann's nor Biot's theory treats the effect of microcrack closure on the elastic behavior of rocks under pressure or the chemical effects such as the interaction between water and clay particles.

The parameters of the solid matter of the frame that enter equations (1), (1a), and (1b) are  $\hat{k}$  and  $\hat{\rho}$ . Some typical values are listed in Table 1; additional data are available in Clark (1966).

The parameters of typical fluids are also known, and some are listed in Table 2.

Thus, characteristics of the solid matter and the pore fluid of many rocks can be assigned numerical values without much difficulty.

If values of  $\bar{k}/\hat{k}$  are 0.5 or greater, with  $\phi = 0.2$  and with  $\hat{k}/\bar{k} = 18$ , the magnitude of the second term on the right of equation (1a) becomes about .06  $\hat{k}$  or less. This means that if the frame or skeleton has relatively high elastic constants, its characteristics essentially govern the properties of the whole rock, regardless of the fluid filling. Those characteristics depend upon the elastic interactions between the grains, their bonding, and the presence of microcracks.

If, however,  $\bar{k}$  and  $\bar{\mu}$  are zero,  $\mu$  and  $\bar{M}$  also become zero, and equation (1a) becomes:

$$\frac{1}{M} = \frac{1}{k} = \frac{\phi}{\bar{k}} + \frac{(1 - \phi)}{\hat{k}}. \quad (1d)$$

**Table 1. Bulk modulus and density of some minerals**

Solid	Bulk modulus, $\hat{k}$ dynes/cm <sup>2</sup> × 10 <sup>10</sup>	Density, $\hat{\rho}$ gm/cm <sup>3</sup>
α-Quartz	38	2.65
Calcite	67	2.71
Anhydrite	54	2.96
Dolomite	82	2.87
Corundum	294	3.99
Halite	23	2.16
Gypsum	40	2.32

**Table 2. Bulk modulus and density of some fluids**

Fluid	Bulk modulus, $\bar{k}$ dynes/cm <sup>2</sup> × 10 <sup>10</sup>	Density, $\bar{\rho}$ gm/cm <sup>3</sup>
Water, 25°C (distilled)	2.239	0.998
Sea water, 25°C	2.402	1.025
Brine, 25°C (100,000 mg/L)	2.752	1.0686
Crude oil (1)	0.862	0.85
(2)	1.740	0.80
Air, 0°C (dry, 76 cm Hg)	0.000142	0.001293
Methane, 0°C (76 cm Hg)	0.001325	0.007168

**Table 3. Upper and lower theoretical bounds for velocity in aggregates of quartz crystals, with no porosity and no microcracks**

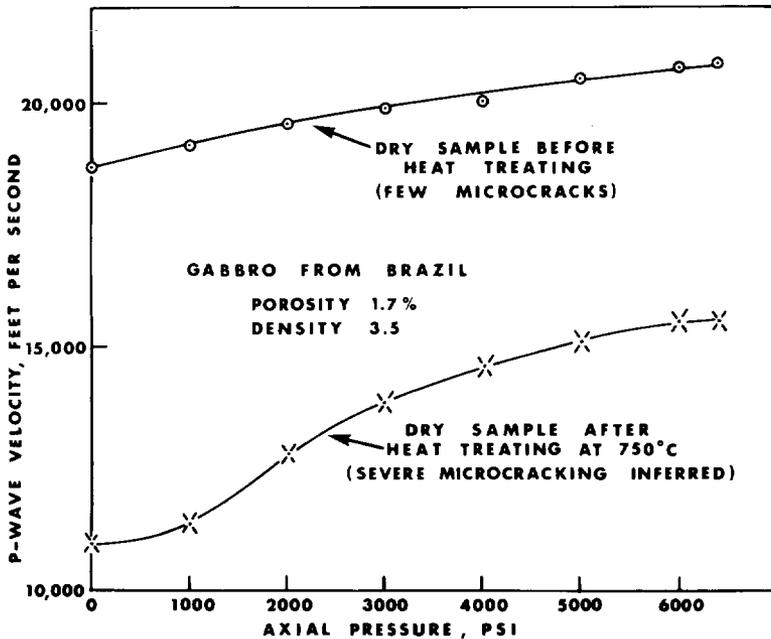
	Voigt	Reuss
P-wave velocity, ft/sec	20,300	19,300
S-wave velocity, ft/sec	13,900	12,900

**EFFECT OF MICROCRACKS**

Some sedimentary rocks such as quartzites and most igneous rocks have almost no porosity. For these rocks the velocity is determined by the mineral composition, provided an extensive system of microcracks is not present. In the absence of microcracks, the elastic parameters can be accurately estimated by use of the theories of Voigt (1928) and Reuss (1929) and the known elastic constants of the crystals. It has been demonstrated by Hill (1952) that the Voigt and Reuss methods give upper and lower bounds to velocities for aggregates of crystals that are randomly oriented. A relevant example is provided by quartz crystals, and the results are given in Table 3.

The effect of microcracks can be illustrated by the behavior of gabbros upon heating, as demonstrated by Ide (1937). Also, our experimental data for a gabbro as depicted in Figure 2 shows that the untreated, dried rock has a P-wave

This is the relationship applicable for mixtures of any two fluids. For clays and shales having very high water content,  $\bar{k}$  and  $\bar{\mu}$  approach zero, and the velocity of the formation approaches that of the fluid. It is noteworthy that for clays and shales the water is "bound" to the fine grained microstructure so that water content is a more pertinent term than porosity for use in describing the rock structure. Again, the properties of the rock frame are difficult to characterize. Permeabilities of clays and shales, of course, are very much lower than those of reservoir sands.



**FIG. 2. Effect of microcracks on velocity of gabbro.**

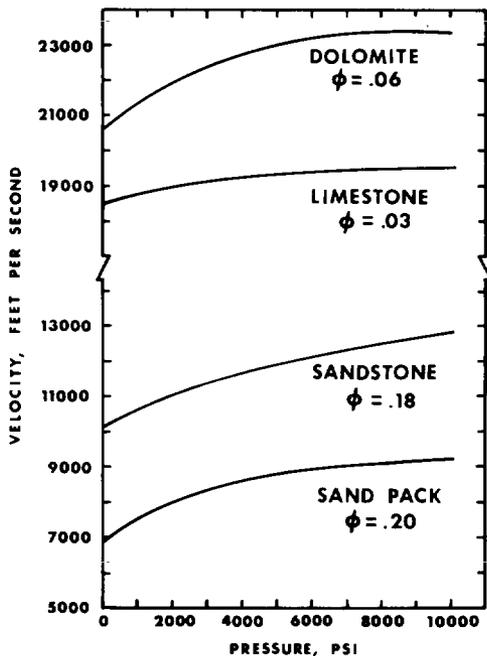


FIG. 3. *P*-wave velocity versus skeleton pressure for brine-saturated carbonates, sandstone, and sand pack.

velocity of 18,700 ft/sec which increases to 20,800 ft/sec with axial stress. In contrast, the treated rock (heated to 750°C and cooled) has a *P*-wave velocity of 11,000 ft/sec which increases with pressure but even at 6000 psi does not exceed 16,000 ft/sec. Heating is presumed to introduce a system of microcracks caused by the unequal expansion of the minerals. Pressure tends to close the cracks, but a very large pressure would be required to reestablish the original framework. The typical observation that nonporous rock velocity increases with pressure thus is attributable in substantial measure to the presence at low pressures of microcracks which are diminished at higher pressures. A similar relationship also holds true for the more porous rocks. Figure 3 shows examples of velocity versus pressure for rocks spanning a range of porosity and mineral composition. In the case of the packing of quartz grains, the "microcracks" are, presumably, the contacts between grains.

#### TIME-AVERAGE RELATIONSHIP

At pressures corresponding to those of deeper sediments in situ, the influence of variation in pressure on velocity becomes small, and then

porosity and mineral composition alone determine velocity. Under this condition, a time-average relationship has been found empirically to interrelate velocity and rock parameters for a fairly wide range of porosities:

$$\frac{1}{V} = \frac{\phi}{V_F} + \frac{1-\phi}{V_M} \quad (2)$$

The parameters  $V_F$  and  $V_M$  give the dependence on fluid velocity and mineral composition;  $\phi$  is the fractional porosity. Wide experience with both in-situ determinations and laboratory experimentation supports the general applicability of this relationship for most sedimentary rocks, particularly when the fluid content is brine. This relationship is compatible with Gassmann's theory in recognizing that the elastic moduli of the frame increase as porosity decreases.

The parameter  $V_M$  is equal to the value of  $V$  as  $\phi$  approaches zero. We have noted earlier that the theories of Voigt and Reuss can be used to estimate the velocity for this extreme case. For example, as given in Table 3, the velocity  $V_M$  for aggregates of quartz crystals should lie between 19,300 and 20,300 ft/sec. Such a value has been found to be very satisfactory for deep clean sandstones. A value of  $V_M$  between 22,000 and 23,000 ft/sec similarly is found to be applicable for deep limestones.

For formations at relatively shallow depths the influence of microcracks, and therefore pressure, cannot be ignored without introducing significant errors. The time-average equation can be retained if the parameter  $V_M$  is regarded as an empirical constant with a value less than the Voigt-Reuss values. In other words, we can assume that the traveltime (which is the quantity usually measured) is a linear function of porosity at any depth with the coefficients of the linear form to be chosen by consideration of suitable data.

As a special illustration of the use of the time-average equation, in Figure 4 we have plotted some laboratory data for cores from a depth of about 5000 ft. The cores were confined at a skeleton pressure of 3000 psi with brine in the pores to simulate the original environment. The two principal minerals in the rock (calcite and quartz in the form of tripolite) are intimately mixed in relative proportions varying from approximately 50 percent calcite-50 percent quartz to 80 percent calcite-20 percent quartz. A petro-

graphic analysis of the cores indicated that the lower porosity samples (also from the upper part of the formation) appeared to have a continuous calcite matrix, whereas the higher porosity samples appeared to have a continuous quartz (tripolite) matrix. It can be seen that the lower porosity data points can be approximated by a time-average line with  $V_M = 22,500$  ft/sec, which is a velocity suitable for a calcite matrix; the higher porosity data points can be approximated by a time-average line with  $V_M = 19,200$  ft/sec, which is a velocity suitable for a quartz matrix. It is interesting that the data appear to separate on these two lines according to the mineral that is predominantly the continuous phase. In this investigation we found no correlation between velocity and the concentration by volume of the minerals.

It is also of interest to note that when the traveltimes at high pressure (10,000 psi) were plotted against porosity, no separation of the data along two lines could be detected. This may indicate that velocity measurements on cores at pressures appropriate to the depth of the formation contain more useful information than measurements at an arbitrarily high pressure.

#### OVERBURDEN AND FLUID PRESSURE

The overburden pressure is usually defined as

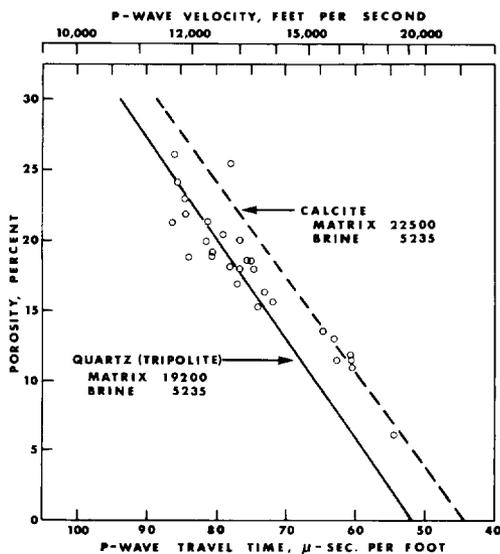


FIG. 4. Velocity versus porosity for samples of quartz-calcite rocks under 3000 psi confining pressure.

the vertical stress caused by all the material, both solid and fluid, above the formation. An average value is 1.0 psi for each foot of depth, although small departures from this average have been noted. The fluid pressure is usually defined as the pressure exerted by a column of free solution that would be in equilibrium with the formation. The reference to a free solution is significant when dealing with clays or shales with which other pressures such as osmotic, swelling, etc., can be associated. The normal fluid pressure gradient is frequently assumed to be .465 psi for each foot of depth, although large departures from this value occur in high-pressure shales.

The skeleton or frame pressure of a rock is the total external pressure less the fluid pressure. The elastic parameters of the skeleton increase as the skeleton pressure increases, and a corresponding increase in velocity is observed. The increase in elastic parameters is attributable to the reactions at the intergranular contacts and the closure of microcracks as the skeleton pressure increases. Hence, when both overburden pressure and formation fluid pressure are varied, only the difference between the two has a significant influence on velocity.

A set of our velocity data that confirms this assertion is given in Figure 5. When the skeleton pressure  $\bar{P}$  or the difference between overburden and fluid pressure is increased, the velocity increases; when the difference between overburden and fluid pressure remains constant, the velocity remains constant.

#### RECENT BASINS

In this section we consider a sedimentary basin which illustrates how  $P$ -wave velocity is affected by many of the factors discussed above. Wells in a number of young basins typically will penetrate successive layers of sand and shale that may range in age from recent to lower Eocene. This provides us an opportunity to study the effects of age, pressure, depth, porosity, and fluid content for a fairly constant matrix material.

The uppermost sedimentary layers are unconsolidated, and the porosity varies mainly with the grain size distribution and clay content. The velocity is only slightly greater than that of sea water. With increasing depth the velocity increases partly because the pressure increases and partly because cementation occurs at the grain-

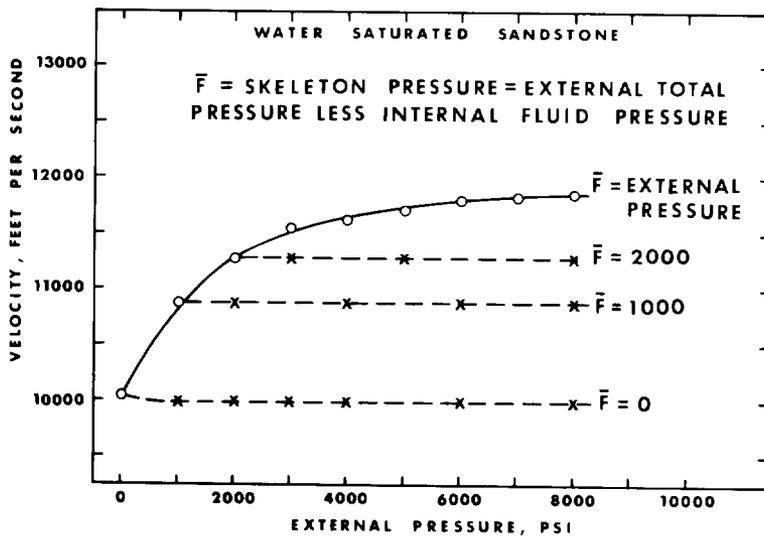


FIG. 5. Velocity through a water-saturated sandstone core as a function of skeleton pressure.

to-grain contacts. Cementation is the more important factor. It has been shown by Maxwell (1960) that the rate of cementation depends also on the rate of flow and the composition of fluids flowing through the pores as well as on temperature.

The rapid increase of velocity with depth normally continues until the time-average velocity is approached. Below this depth, the layers behave like other well-consolidated rock and the velocity depends mainly on porosity.

It is for the shallower layers that the fluid content, i.e., water, oil, or gas has an appreciable effect.

The solid curve in Figure 6 shows a representative curve of velocity versus depth for brine-saturated, in-situ sands based on some sonic log and electric log data. The dotted curve is from laboratory data for fresh, unconsolidated, water saturated packings of quartz sand grains at pressures corresponding to the depth. Thus, the dotted curve indicates what would happen to sands if they were buried without consolidation or cementation, and the divergence of these curves is attributable to these effects. The dashed curve shows the time-average velocity calculated using the average porosity read from well logs. At the shallower depths the actual velocity is less than the time-average, but below about 8000 ft the agreement is close.

Figure 7 illustrates the results for sands in more detail. For any depth the traveltime can be approximated by a linear function of porosity; below 8000 ft this linear function coincides with the time-average equation.

#### HIGH-PRESSURE SHALES

In the wells referred to above, the fluid pressure

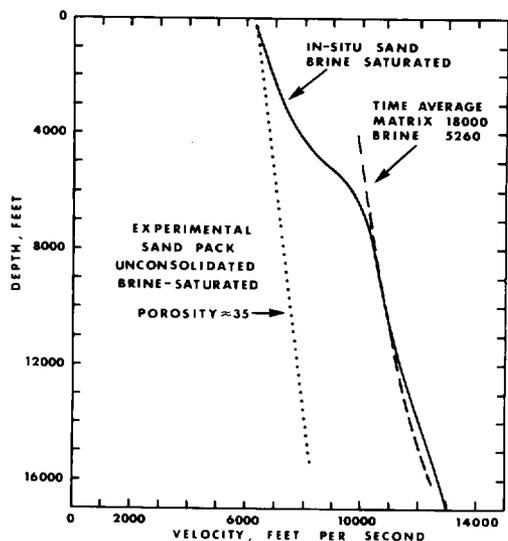


FIG. 6. Velocity as a function of depth showing consolidation effect for in-situ tertiary sands. For comparison, the velocities of experimental sand packs at pressures corresponding to these depths are also shown.

increased at a rate of about 0.5 psi per foot of depth. However, it sometimes happens that high fluid pressure zones are encountered, i.e., zones in which the fluid pressure is well above that given by the normal gradient with depth. Such wells provide an opportunity to study the relation between velocity, the difference between overburden pressure and fluid pressure, and consolidation with depth.

Hottmann and Johnson (1965) presented pertinent data for velocity measured in shales and the corresponding fluid pressure and depth. When there was no excess fluid pressure, they found that the interval traveltime  $\Delta T$ , in microseconds per foot, decreased with depth  $Z$ , in feet, according to the formula

$$Z = A - B \log_e \Delta T, \quad (3)$$

where  $A = 82,776$ , and  $B = 15,695$ . They also give data from wells that penetrated zones with abnormally high fluid pressure. We have found that all these data can be correlated by the equation

$$\left( \frac{P_0 - P_F}{\alpha - \beta} \right)^{1/3} Z^{2/3} = A - B \log_e \Delta T, \quad (4)$$

where  $P_0$  = overburden pressure;  $P_F$  = fluid pres-

sure;  $\alpha$  = normal overburden pressure gradient; and  $\beta$  = normal fluid pressure gradient. Figure 8 illustrates this correlation.

For a normally pressured section equation (4) reduces to equation (3). One interesting feature of equation (4) is that both the pressure difference  $P_0 - P_F$  and the depth  $Z$  are present. The factor  $Z^{2/3}$  may be interpreted as the effect of increased consolidation with depth. For sands the effect of consolidation outweighs the effect of pressure.

#### UNCONSOLIDATED GAS SANDS

In a shale-sand sequence some sands may contain oil or gas, but the overlying shale may not contain either, except for the amount dissolved in water. The reflection coefficient at such an interface may be influenced appreciably by the fluids in the formation, if the depth is not too great. Gassmann's equation can be used to estimate the magnitude of the possible effect of the presence of oil or gas in a sand upon its velocity when the velocity of the brine-saturated sand is known.

By using appropriate values in equation (1a) for parameters other than the skeleton moduli of a given brine-saturated sand, we can determine the relative skeleton modulus  $\bar{k}/\bar{k}$ . The values of the parameters used are deduced in part from known wave velocities equations, (1b) and (1c), and a specification that  $\bar{M} = S\bar{k}$  with  $S$  having a representative value of 2.0. Suitable values for  $\bar{k}$  or  $\bar{k}/\bar{k}$  can be used for different fluid filling of the pores, and the corresponding values of  $\bar{M}$  and the  $P$ -wave velocity calculated using equations (1).

The results of some such calculations are given in Figure 9, where velocity is plotted versus depth for sands saturated with either brine, oil, or gas. The dashed curve illustrates typical values of velocity versus depth for shales saturated with brine. It is clear that in the first few thousand feet the reflection coefficient at the boundary between a shale and a sand will be significantly greater if the sand contains gas than if the sand contains brine. This observation might possibly be of practical significance when there is a lateral change from a brine-filled sand to a gas-filled sand. However, at considerable depths the reflection coefficient becomes almost independent of the nature of the fluid content.

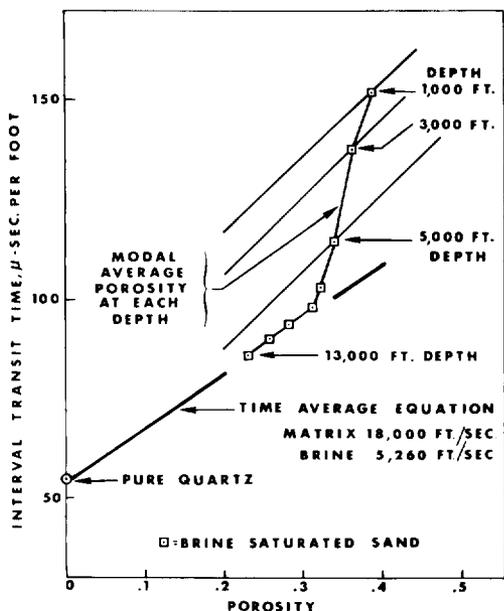


FIG. 7.  $P$ -wave transit times from well logs for sands in young sedimentary basins.

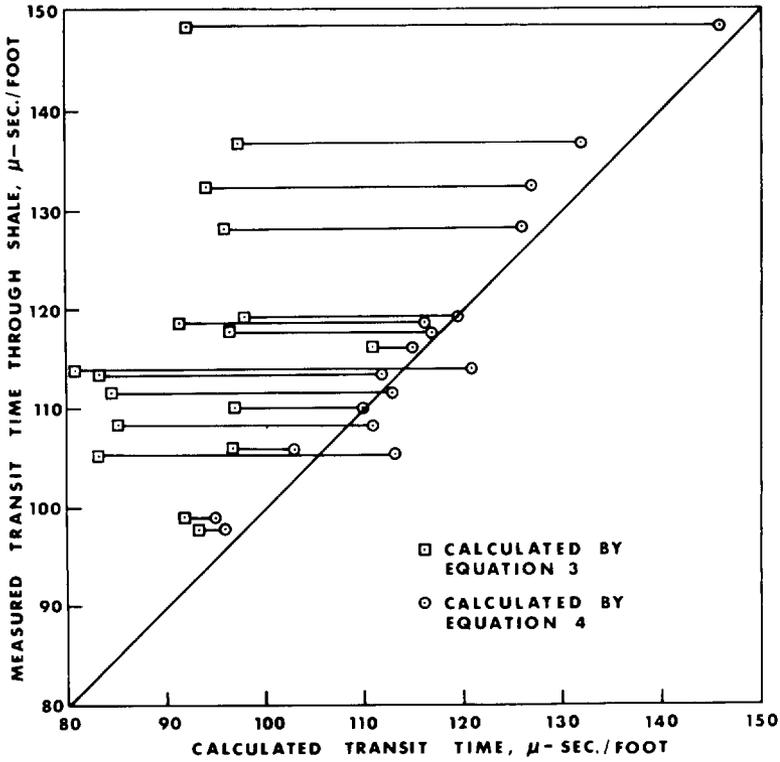


FIG. 8. Transit time relations for high-pressure shales [data from Hottmann and Johnson (1965)].

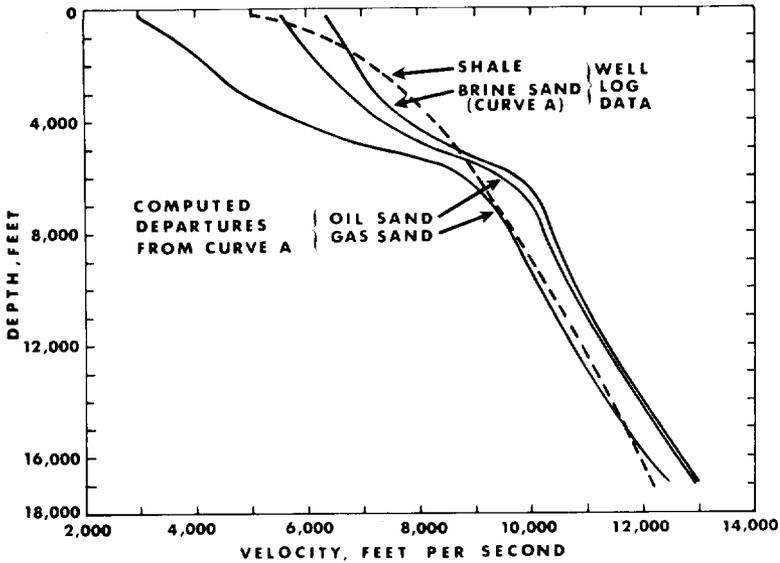


FIG. 9. Velocity versus depth for shale and for in-situ sands containing different fluids.

## REFLECTION COEFFICIENTS

Acoustic impedance contrasts govern seismic reflection coefficients at a plane interface between two media. Thus,

$$R = \frac{\rho_1 V_1 - \rho_2 V_2}{\rho_1 V_1 + \rho_2 V_2} \quad (5)$$

gives the amplitude of the reflected wave when the incident wave has unit amplitude and is perpendicular to the interface.

Peterson et al (1955) showed that for practical purposes this formula can be approximated by

$$R = \frac{1}{2} \ln (\rho_1 V_1 / \rho_2 V_2). \quad (6)$$

The empirical relationship between density and velocity depicted in Figure 1,

$$\rho = .23V^{.25}, \quad (7)$$

is a fair average for a large number of laboratory and field observations of different brine-saturated rock types (excluding evaporites). Combining this with Peterson's relation, we have

$$R = \frac{1}{2} (1.25 \ln V_1 / V_2). \quad (8)$$

In a general way then, density commonly varies with velocity so that its effects upon reflection coefficients is fairly satisfactorily taken into account by multiplying the reflection coefficient due to velocity contrast by a factor 1.25. Departures from this rule exist as evidenced by scatter of observation points and may in some cases be significant. The multiplying factor 1.25 also increases the relative amplitudes of multiple reflections over that estimated using velocity contrasts alone.

## CONCLUSIONS

We have attempted to show that the *P*-wave velocity in the upper layers of the earth (depths less than 25,000 ft) varies systematically with different factors, although an absolute prediction of velocity is seldom possible. The following summary of the effects appears to be generally valid.

(1) Gassmann's theory is typically valid for sedimentary rocks in interrelating elastic constants, densities, and *P*-wave velocities for different rock components and for the consolidated whole rock. An important component, however, is the frame or skeleton, which may

have a wide range of elastic parameters. These parameters ordinarily can be characterized only through Gassmann's theory, and this fact limits the usefulness of the theory for practical applications.

(2) Microcracks can be present in a rock (within the rock skeleton) and materially reduce the *P*-wave velocity of the whole rock. Pressure can close them and cause the velocity to increase. The elastic parameters of rocks without microcracks can be estimated by using the theories of Voigt and Reuss and the known elastic constants of the crystals.

(3) The well known time-average relationship empirically relates velocity and porosity for a moderately wide range of rock types and formation fluids when the rock is under a substantial pressure.

(4) The effective pressure governing the elastic properties of the skeleton of porous sedimentary rocks is the difference between the total external pressure (or overburden pressure) and the internal fluid pressure. Increase in the skeleton pressure increases the elastic reactions at intergranular interfaces and the velocity of the whole rock.

(5) It is shown that in a recent basin, the increase in velocity of sands with depth is substantially greater than the increase in velocity of a sample of comparable sand subjected to pressure in the laboratory. The difference is mainly attributed to in-situ cementation of sand grains with geologic time.

(6) For high-pressure shales, the skeleton pressure is markedly subnormal and may approach zero. The associated velocities are also greatly reduced but have values consistent with the existing skeleton pressure.

(7) For unconsolidated shallow gas sands, some computations using Gassmann's theory indicate a substantial difference in *P*-wave velocity from that of the same sand filled with brine. Some possibility of recognizing this effect in seismic reflections exists.

(8) A simple systematic relationship exists between the velocity and density of many sedimentary rocks in situ. For these rocks the empirical relationship permits estimation of reflection coefficients from velocity information alone.

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