The Mid-Atlantic Ridge near 45° N. XI. Seismic velocity, density, and layering of the crust[1]

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Compressional wave velocities at pressures to 1000 kg/cm² and densities are given for a representative suite of rocks selected from 42 dredge hauls on the Mid-Atlantic Ridge near 45° N. The spectrum of rocks studied includes vesicular and massive basalts, metabasalts, meta-gabbros, and serpentinites. Evidence is presented for block faulting of an originally continuously layered crust of vesicular basalt and massive basalt underlain by a metamorphosed basalt and gabbro sequence. A study of the velocities of these layers at *in situ* pressures shows that they correlate well with the seismic layering of the oceanic crust. The velocity of the massive basalt layer is in agreement with that reported for layer 2. The underlying layer, consisting of low-to-medium grade metamorphosed basalts and gabbros (greenstones and greenschists) exhibits higher velocities. None of these exceed 6 km/s but it is suspected that these rocks at greater burial depths will exhibit velocities comparable to those of layer 3. The occurrence of serpentinites at all elevations on the slopes of seamounts implies that they do not form a continuous layer and because of their relatively low velocity, it is unlikely that layer 3 is composed of these rocks. Their presence as hydrated diapiric intrusives is more plausible.

Introduction

Seismic structural information can be used more effectively if correlation can be made between seismic refraction velocities and laboratory measurements of velocities of rocks representative of the geological section of the area. The correlation will be more meaningful if the laboratory measurements on the rocks are carried out at pressures equal to their *in situ* pressures. During recent cruises to the Mid-Atlantic Ridge near 45° N, over 800 specimens were collected at 42 dredge stations on the slopes of seamounts (Aumento 1968, 1969; Aumento and Loncarevic 1969) providing extensive coverage from 70 km east to 150 km west of the axis of the Ridge (Fig. 1). In addition, a detailed picture of the velocity of crustal layers was obtained from the seismic refraction profiles whose locations are shown in Fig. 1 (Keen and Barrett 1969; Keen and Tramontini 1970). The crustal section consists of five discrete, apparently continuous, layers with a range of velocities from 2.2 km/s to 6.8 km/s underlain by a normal mantle velocity of 7.8–8.0 km/s; the whole crustal thickness ranges from 3.5–4 km.

A number of continuous seismic reflection profiles were obtained in the area extending westward from the Median Valley for approximately 160 km (Keen 1969; Keen and Manchester 1970). The latter showed an increase in the thickness of unconsolidated sediment with distance from the Median Valley. The steep and abrupt slopes in basement topography suggest that many of the seamounts are block-faulted features, and one can infer that it may be possible to sample at least the upper 1.5 km of the geological section by dredging the slopes of seamounts at various elevations. A regional geological picture may thus be inferred by correlating the resultant velocities measured in the laboratory with the seismic refraction velocities and structure.

In addition to the specimens from the dredge hauls and the seismic information, detailed bathymetric, gravimetric, and total magnetic field surveys were made over the whole area shown in Fig. 1 (Loncarevic *et al.* 1966; Aumento *et al.* 1970). It was shown that the

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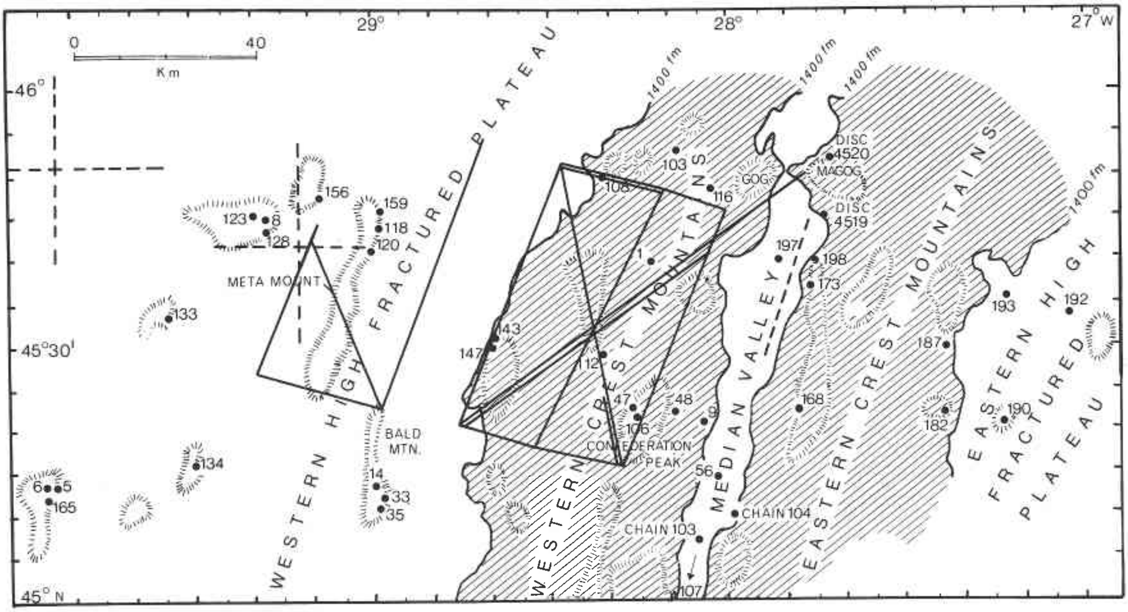

FIG. 1. Schematic bathymetry of the area at 45° N showing the major morphological features; the locations of the 42 sampling stations (solid dots), sonobuoy refraction profiles using explosives (solid lines), and two-ship refraction profiles using an airgun (broken lines) are superimposed.

residual Bouguer gravity anomaly persists regardless of the choice of a single replacement density, suggesting geological inhomogeneities within the area. Density measurements for the geological section sampled and the correlated seismic information will undoubtedly improve the quality of the Bouguer map and provide more information on the deepest crustal layer and upper mantle.

This paper gives compressional wave velocities at pressures in excess of 1000 kg/cm^2 of a representative suite of rock samples from the Mid-Atlantic Ridge at the locations shown in Fig. 1.[2] Velocity as a function of pressure is discussed for the various rock types and velocity versus density of the samples is compared with the empirical curves derived from the laboratory measurement of velocities by other workers. Evidence for layering and block faulting in the area is presented to show that the dredge material is representative of the geological section.

Sample Descriptions

Various rock types with differing degrees of metamorphism, hydration, porosity, and weathering were obtained from the area near 45° N; in selecting specimens for the velocity measurement, an attempt was made to cover the spectrum of these properties. A brief petrographic description of the specimens studied is given in Table I. The degree of porosity and weathering is also indicated, as are the lengths, densities, and velocities at *in situ* lithostatic pressures (V_p) of the cores studied.

The associated range of densities (from 2.12 to 3.01 g/cm^3) is indicative of a wide range of velocities in the rock suite. Complete petrographic and chemical descriptions of these rocks together with observed variations with topography have been reported elsewhere (Aumento 1968, 1969; Aumento and Loncarevic 1969).

Curves indicating the variation in velocity with pressure for all samples are presented in Fig. 2. No correction has been made for the change of length of the specimen under pressure. The error in velocity due to the latter is less than a few parts per thousand at the highest pressure (Birch 1960) and can therefore be neglected. Similarly, the change in density at the highest pressure is small and may be neglected (Woollard 1959).

The velocities of most of the samples rise

[2]See Appendix I for experimental procedure.

TABLE I

Petrographic description, density, length, and velocity at *in situ* pressure of cores drilled from the dredged rock sample from the Mid-Atlantic Ridge 45° N

Specimen No.	Description	Density (g/cm³) ± 0.01	Length (cm) ± 0.003	V_p (km/s) ± 1%
198-1C	Massive Gabbro showing incipient alteration to Greenstone	2.91	5.035	5.70
198-2	Fine grained Gabbro showing incipient alteration to Greenstone	2.86	5.423	5.62
6-4	Metagabbro	2.72	5.994	5.56
187-1C	Metabasalt (Greenstone)	2.86	4.595	5.45
187-1B	Metabasalt (Greenstone)	2.86	6.162	5.14
47A-1	Fresh Zenocrystic Basalt	2.70	8.646	4.30
197-7	Very fresh massive Zenocrystic Basalt	2.89	6.863	4.74
173-2	Fresh Basalt	2.78	4.651	4.75
173-1C	Fresh fine-grained Basalt	2.84	4.417	4.70
173-1B	Vesicular Pillow Basalt	2.81	3.425	4.44
106-1	Vesicular Pillow Basalt	2.37	5.793	3.64
165-8	Fresh Serpentinized Gabbro	2.59	6.151	4.52
123-1	Partially weathered Basalt with secondary veins	2.25	7.181	3.84
165-12	Weathered Basalt showing incipient alteration to Greenstone	2.68	3.425	3.97
187-2B	Weathered Basalt	2.17	3.261	3.40
165-1	Partially weathered Serpentinized Peridotite	2.12	4.049	2.05
165-2	Partially weathered Serpentinized Peridotite	2.15	2.019	2.24
165-3	Partially weathered Serpentinized Peridotite	2.14	6.099	2.14
106-3	Fresh Basalt (possibly erratic)	3.01	6.704	—
5-3	Amphibolite (possibly erratic)	2.77	6.482	—

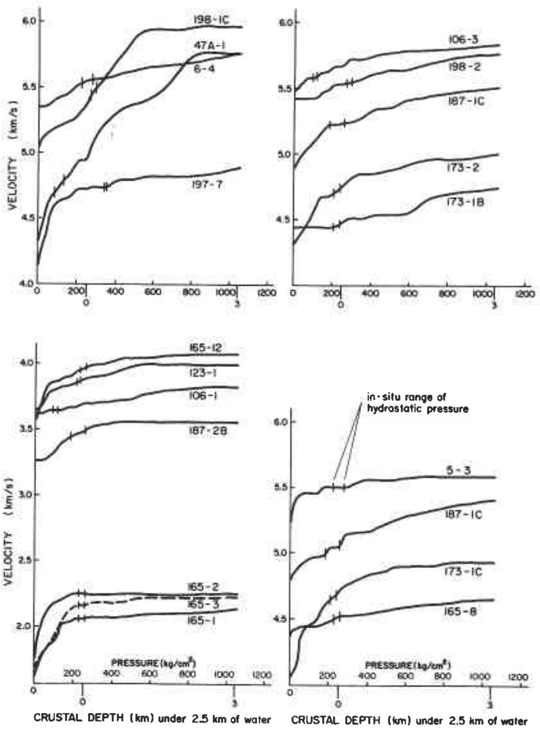

FIG. 2. Experimental relationship between compressional wave velocity and pressure for all the core samples. Equivalent crustal depths are calculated using an average rock density of 2.67 g/cm³ and assuming a depth of water of 2.5 km.

sharply as the pressure is increased from atmospheric to about 300 kg/cm². This may reflect the fact that most of the samples were collected in a depth of water where the hydrostatic pressure approximates this value. Also in many cases, a sharp increase in velocity with pressure occurs near the *in situ* hydrostatic pressure shown in Fig. 2. It was observed that many of the samples had microscopic fractures that probably developed when the rocks were relieved of their hydrostatic head upon recovery. In fact, the glassy margins of some of the fresher, more porous rocks exploded on the ship's deck.

The highest velocities measured at *in situ* pressures (>5.14 km/s) are characteristic of the metamorphosed basic rocks, the highest values being those of gabbros showing incipient alteration to greenstones (>5.56 km/s). This can be seen more clearly in Fig. 3, a plot of velocity at *in situ* pressure versus density. The velocities of the metabasalts are slightly lower than the metagabbros although both rock types have been altered to greenstones of similar densities. The *in situ* velocities of all the fresh massive basalts are very similar with a mean average of about 4.74 km/s. The density differences between all the basalts and metabasalts are negligible. A wide spread in the velocities (3.64 to 4.44 km/s) and densities (2.37

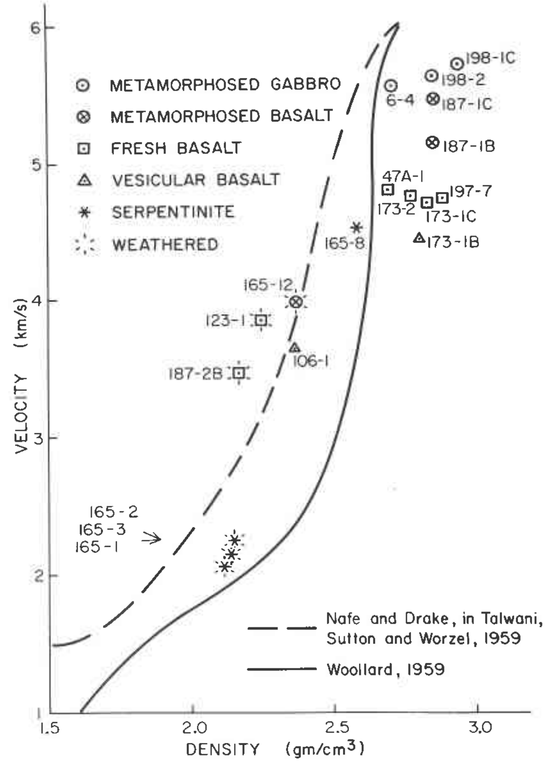

FIG. 3. Experimental relationship between compressional wave velocity (at *in situ* lithostatic pressures) and density for all core samples. Empirical curves published by others are presented for comparison. The two samples of possible erratic origin have not been plotted.

to 2.81 g/cm^3) of pillow basalts is evident from the two samples measured. The velocities in these rocks undoubtedly vary widely with the degree of vesicularity and subsequent infilling of vesicles by zeolites. The measured density and velocity of fresh serpentinite were lower than those of the fresh basalts and metabasalts. The remainder of the material showed varying degrees of weathering and associated low velocities and densities.

Figure 3 also includes curves showing the relation of compressional wave velocity to density from the results published by Woollard (1959) and Nafe and Drake (in Talwani *et al.* 1959). It is evident that all the basalts and metabasalts have a significantly lower velocity to density ratio than either of the two curves. However, these lower velocities are consistent with the refraction seismic results of Keen and Barrett (1969).

Evidence for Layering and Block Faulting

As pointed out earlier, if it can be shown that the seamounts comprising the Mid-Atlantic Ridge are large uplifted blocks of an originally uniformly stratified, flat-lying volcanic sequence, then the slopes of the seamounts provide an exposed geological section from which samples can be obtained. The steep and abrupt slopes observed are not inconsistent with the above suggestion. Further, Keen (1969) showed from continuous seismic profiling that the evidence for near-vertical block faulting from basement topography underlying the sedimentary cover is even more convincing than that from bottom topography.

If block faulting of a systematically layered crust has occurred, then one might expect that the rock type found at a given depth below the top of a seamount, D, would be the same as that found at equal D's all over the area. However, using the dredging technique, there are two factors which may complicate an examination of the above concept. First of all, the dredge material is collected over a range of depths and therefore may sample more than one layer. Also, some of the material collected could have rolled down from higher elevations. Therefore, to assess this concept we must only consider the maximum depths of the dredge hauls below the tops of the seamounts, E, as a function of the type of material collected.

Figure 4 demonstrates that certain rock types are found within a discrete range of depths. The abscissa represents the number of dredge hauls, N, containing a particular rock type whose E is greater than a given depth, D, the ordinate. This number has been normalized by dividing by the total number of dredge hauls containing the same rock type, N_T, and converting to a percentage. Figure 4 shows that vesicular basalt most frequently occurs near the tops of seamounts whereas the metamorphics form the deepest part of the section. The massive basalts are most commonly found to lie between the vesicular basalt and the metamorphics. It should be noted that the separation of these curves will be less obvious if the layering is not horizontal with respect to the plane of zero depth, D.

To obtain the best layered model, the ranges of depth, D, is shown as a function of distance from the axis of the ridge in Fig. 5. The rock

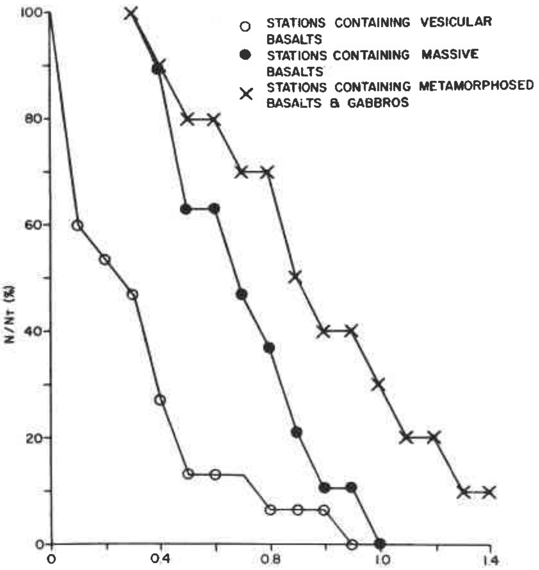

FIG. 4. A plot of the number of dredge hauls, N, containing a particular rock type whose E is greater than a given depth, D, the ordinate, divided by the total number of dredge hauls containing the same rock type, N_T.

types collected at each station are placed within the range of depth to obtain smooth horizons. The most striking feature of Fig. 5 is the consistent occurrence of metamorphosed basalts in those stations whose depth, D, are greater than 0.4 km and their absence at lesser depths. It is therefore proposed that these basalts and gabbros (now greenstones and greenschists) once formed semi-continuous layers prior to block faulting and that metamorphism to the greenschist facies occurred at burial depths of about 1 km. Above this horizon, basalts are common to most of the dredge hauls and apparently outline a layer of about 0.4 km in thickness. Shallower depths yield mostly vesicular pillow basalts, some with zeolitic replacement. On both the Crest Mountains and High Fractured Plateau volcanic agglomerates are common at all depths.

Neglecting the weathered vesicular basalt rocks, the layering shown in Fig. 5 can be inferred without considering that some of the material came from higher elevations, that is, all of the rocks collected could have been *in situ*. Serpentinites, however, occur at all elevations and are randomly distributed throughout the crustal section. It is thus suggested that these rocks are local diapiric intrusive masses, and do not outcrop as undisturbed sections of the layered ultramafic bodies.

Apparent thinning of the uppermost vesicular basalt layer with increasing distance from the axis is consistent with the ocean floor spreading hypothesis. Longer periods of weathering and erosion of the older rocks would explain the observed thinning of the uppermost layer and also would contribute more sedimentary material at greater distances from the ridge axis as observed by Keen and Manchester (1970).

While the details of the seismic refraction results will not be discussed here, the velocities and thicknesses of layers agree well with the model presented in Fig. 5. On all profiles shot in areas of low relief, a 4.7 km/s layer with a thickness of about 0.5 km was underlain by a layer with a range of velocity of 5.8 to 6.0 km/s. This provides further evidence to support the systematically layered and block faulted model.

Summary

The observations can be summarized as follows:

(1) There is considerable evidence to show that near vertical (Keen 1969) block faulting is responsible for much of the topographic relief near the axis of the Mid-Atlantic Ridge. This implies that at least 1.5 km of the geological section is represented in the dredge material.

(2) The consistence of rock type in the material dredged from a given depth below the tops of seamounts is indicative of a systematically layered crust prior to these major vertical movements.

(3) The uppermost vesicular pillow basalt layer thins with increasing distance from the axis. We attribute this to mechanical and chemical denudation from the tops of seamounts.

(4) The massive basalt layer apparently thins with increasing distance from the axis.

(5) As a result of (3) and (4), there is an apparent progressive shallowing of the metamorphosed basalt and gabbro layer away from the axis of the ridge.

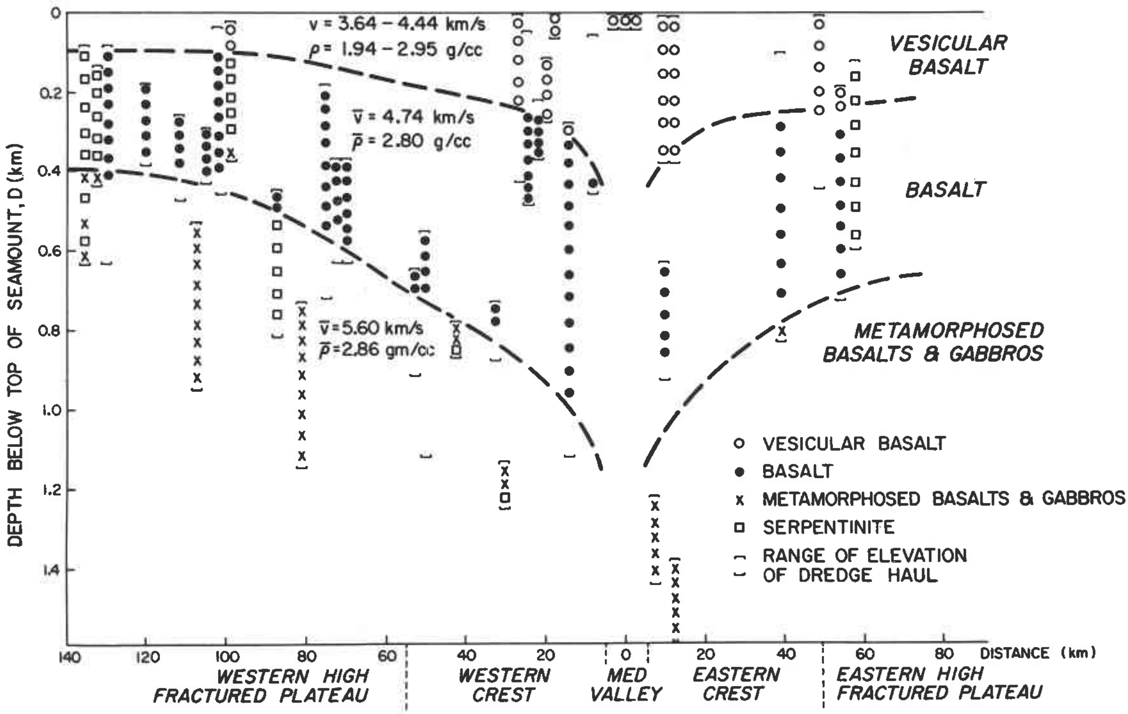

FIG. 5. A plot of the range of depth, D, below the top of the seamount for all dredge samples against distance from the ridge axis. The rock types are plotted within the range of depth of the dredge haul to demonstrate probable layering of the crust prior to block faulting.

(6) The average compressional wave velocity of basalt, metabasalt and metagabbro increases for these rocks in the order given. Basalt velocities compare favorably with that of layer two as given by Raitt (1963) and it is suspected that the metamorphic rocks of higher grade when at greater depths than those observed here, will exhibit velocities comparable to that of layer three. The velocity–density ratio for this rock sequence is somewhat lower than that of other similar sequences in other areas.

(7) A wide range of velocities and densities are observed in the pillow basalts depending upon the degree of vesicularity and amount of vesicle infilling.

(8) The occurrence of serpentinites is random and unrelated to depth below the tops of seamounts. Also the velocity of fresh serpentinite is relatively low (<4.7 km/s) and it is extremely unlikely that this material comprises a deeper higher velocity layer such as oceanic crustal layer three. The presence of serpentinites as hydrated diapiric intrusives along planes of weakness, as pointed out by Van Andel and Bowin (1968) and Van Andel (1968), seems plausible.

In conclusion, we have demonstrated that a systematically layered crust, each layer with its associated velocity, density, and degree of metamorphism existed on the Crest and High Fractured Plateau of the Mid-Atlantic Ridge prior to major vertical block faulting. These results should provide a good model from which gravity data might be used to study the deeper crustal structure on a regional basis. Probable intrusions of serpentinites with low velocity and density are the only anomalies in the simple layered interpretation. In addition, because of the relative non-magnetic properties of the metamorphosed rocks (Irving *et al.* 1970) and the observed thickness of the basalt layer, any structure on the metamorphic horizon will be an important source of magnetic anomaly.

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Appendix I
Experimental Methods

The measurement of velocity consists of the determination of travel time of a compressional pulse through a short cylindrical specimen. It has been demonstrated by others (summarized by Birch 1960) that the true compressional wave velocity can be measured in the above manner independent of length and diameter of sample and of transducer frequency within certain ranges of dimensions and experimental accuracy.

In this experiment, a simplified method was employed to measure compressional wave velocities in samples subjected to hydrostatic pressures in excess of 1000 kg/cm^2. Cores (2.5 cm diameter) were drilled from the dredged rock samples in lengths ranging from to 2 to 9 cm. The ends of the cores were cut perpendicular to the axis of the core on a diamond saw. Each end of the core was then fitted with a cylindrical plate of polarized barium titanate ceramic with a 2.5 cm diameter and 0.5 cm thickness. A thin film of silicone grease applied to the surfaces between the ceramics and the specimen ensured a proper mechanical contact. The configuration was maintained by placing extended rubber bands over the ends of the ceramics.

A 20V electrical pulse with a rise time of about 0.5 μs was fed to one ceramic (the transducer) from a pulse generator causing a sudden thickness expansion of the transducer. This disturbance was transmitted through the rock specimen to the second ceramic (the receiver) where the mechanical energy was converted to an electrical signal. Both the transmitted and received electrical pulses were monitored on a calibrated two-trace oscilloscope where the travel time could

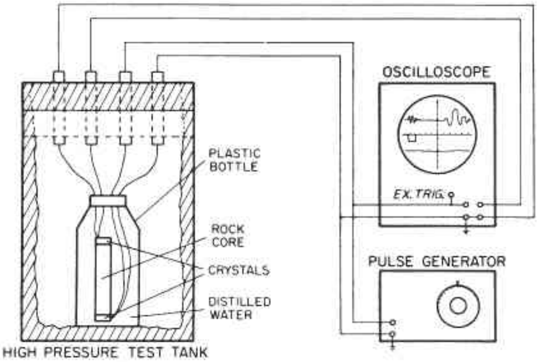

FIG. 6. Experimental configuration used to measure the compressional wave velocities in rock cores.

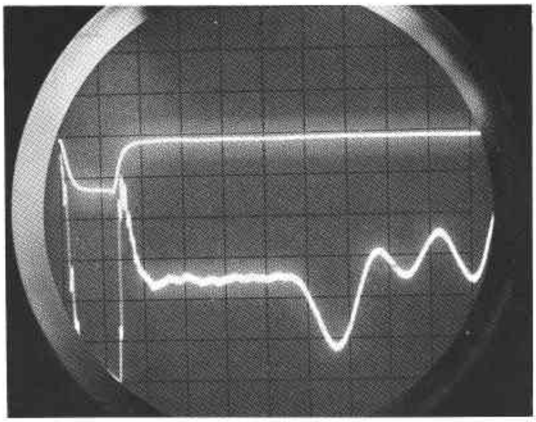

FIG. 7. Typical oscilloscope traces of the transmitted (upper) and received (lower) electrical signals.

be measured directly. In addition, the rock specimens were placed in a high pressure test tank where travel times were measured at pressures up to 4000 psi (285 kg/cm^2) in steps of 500 psi (35 kg/cm^2) and to 15 000 psi (1060 kg/cm^2) in steps 1000 psi (71 kg/cm^2). The compressional wave velocity is the ratio of the length of the specimen to travel time of the first onset. Figure 6 illustrates the experimental configuration and Fig. 7 shows a typical oscilloscope trace of the transmitted and received electrical signals. Densities of the samples were calculated from the measured mass and volume using a calibrated electronic balance and the water displacement method.

The accuracy with which one can determine velocities using the method described above is a function of the length of the specimen, the identification of the first onset in the received signal and the reading accuracy of the length and travel time measuring instruments. The gradual onset of the first arrival received is the principal limitation on the accuracy. As the amplitude of the transmitted signal is gradually increased, the travel time appears to decrease due to the sharpening of the first break. This apparent decrease in travel time was observed until the transmitted signal was increased, in our case, to about 5V where a further increase to 20V produced no measurable change in the first onset time.

In order to identify precisely the actual point on the first onset to be used for travel time measurements, the summed times of two individual samples were compared with the single time measurement for the two samples when placed end to end. Thus a template of the typical first onset could be constructed identifying the point of first break. The estimated error in travel times is 0.05 or 0.1 μs depending upon whether the oscilloscope sweep was set at 2 or 5 μs/cm. Our travel times varied from 8 to 28 μs resulting in an overall error in velocity of less than 1%. It is shown in Fig. 3 that this error is considerably less than the variance between cylinders cored from the same specimen (e.g. 187–1B, 187–1C).