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The Geology of the Oceanic Crust: Compressional Wave Velocities of Oceanic Rocks[1]

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A diverse suite of rocks has been recovered from escarpments in the North Atlantic: basalts, dolerites, meta-basalts (zeolite and greenschist-facies), meta-dolerites (greenschist facies), gabbros, meta-gabbros (greenschist- and amphibolite-facies), serpentinized peridotite, and actinolitic rocks. Since these rocks have been sampled from tectonic escarpments, it is difficult, based on petrology alone, to separate those rock types resulting from localized tectonization from those rocks reflecting regional geologic trends. In an attempt to resolve this ambiguity, the compressional wave velocities (confining pressure range 1 bar to 7 kb) of representative samples were measured as a function of pressure. Compressional wave velocity measurements at confining pressures equivalent to the in situ conditions were compared with published seismic refraction results. These comparisons are compatible with an oceanic basement (3.7–6.0 km/sec) composed of basalt, dolerite, and meta-basalt. The oceanic layer (6.7–6.9 km/sec) is composed of gabbro. The measured velocities of meta-gabbro (greenschist- and amphibolite-facies), serpentinized peridotite, and actinolite-rich rocks are not compatible with the generalized velocity structure of the ocean basins, thus suggesting that these rock types do not occur in abundance within the oceanic crust but instead reflect tectonic processes associated with the escarpments on which they are sampled.

During the past twenty years, seismic refraction studies at sea have yielded solutions for the depth, thickness, and compressional wave velocity of oceanic layers under the mid-oceanic ridge and under ocean basins. Results have been summarized by *Raitt* [1963] and *Ludwig et al.* [1971]. On a travel time-distance plot, the recorded arrivals of waves refracted by the oceanic crust generally have four segments. These arrivals represent three oceanic layers and the upper mantle (Figure 1, left). The uppermost crustal layer (layer 1) varies in thickness (negligible at the ridge crest, up to 5 km in the ocean basins bordering some continents), and consists of sediments ranging from unconsolidated (1.5 km/sec) to consolidated (3.5 km/sec). Oceanic basement (layer 2) underlies the sedimentary blanket. The compressional wave velocities measured for layer 2 over the mid-oceanic ridge and ocean basins range from 3.7 to 6.0 km/sec; the vast majority of these measurements fall within the range of 4.4 to 5.6 km/sec. The thickness of layer 2 varies from 0.70 to 4 km and averages at 2 km. Below layer 2 is the oceanic layer (layer 3), which is observed throughout the ocean basins. Most measurements indicate that layer 3 has a narrow velocity range (6.7 to 6.9 km/sec) and a thickness of 4 to 7 km. Recent, more detailed seismic refraction studies [*Maynard*, 1970; *Sutton et al.*, 1971; Buhl, personal communication], however, indicate that, at least in some areas of the ocean basin, layer 3 is divisible into a 6.0- to 6.8-km/sec upper section and a 7.0- to 7.5-km/sec lower part. The upper mantle underlies the oceanic crust. The characteristic velocity of compressional waves in the upper mantle varies from 7.8 to 8.2 km/sec.

From seismic refraction results alone it is difficult to infer the composition of the two oceanic crustal layers. Studies of the compressional wave velocities of a diverse suite of rock types exposed on the continents show that several rock types have measured velocities compatible with the velocities measured for the oceanic crustal layers. At low pressures, laboratory measurements of sedimentary rocks, vol-

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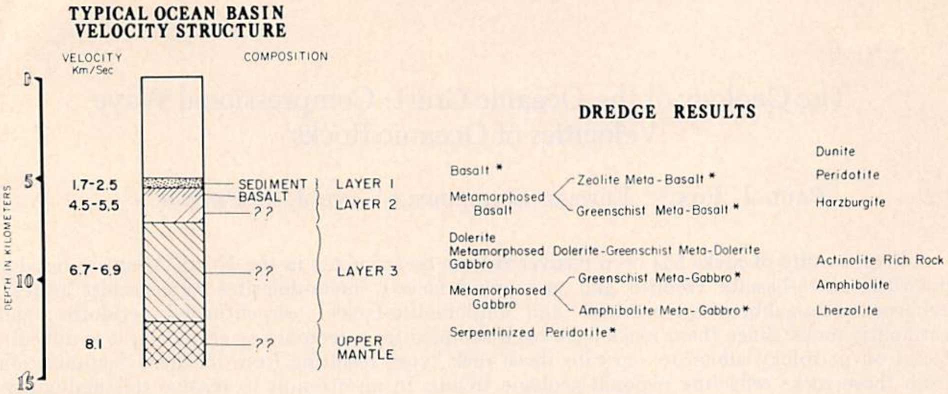

Fig. 1. Left: generalized velocity structure of the oceanic crust. Right: a list of rock types recovered from oceanic escarpments. Asterisks denote abundance.

canic rocks, and low-grade meta-sediments have compressional wave velocities in the 3.7- to 6.0-km/sec range of layer 2 [*Anderson and Liebermann*, 1968]. The measured velocities of layer 3 are in the range of laboratory measurements of limestones, gabbroic rocks, some metamorphic rocks, and some serpentinized peridotites [*Anderson and Liebermann*, 1968]. The analysis of rocks dredged from escarpments in the ocean basins partly resolves the compositional ambiguity, in that some portion of the rocks recovered probably reflect the composition of the oceanic crust. A diverse and complex suite of mafic and ultramafic rocks has been recovered from these escarpments (Figure 1, right), and it is difficult to relate these various rock types to the regional geology of the oceanic crust. The dredges containing a diverse suite of rocks are recovered from steep, tectonic escarpments of the mid-oceanic ridge system. These escarpments either are parallel to the ridge crest and are located at or near the axis of accretion or are transverse to the ridge crest and are associated with fracture zones. From petrology alone, it is difficult to determine which rock types, if any, reflect the regional geology of the oceanic crust and which are a function of localized tectonics and/or dislocation metamorphism associated with the escarpment on which the rocks are found. By defining the velocity characteristics of samples representative of the diverse assemblage of rocks collected during surveys of the fracture zones in the North Atlantic, the velocity envelopes for a given rock type can be compared with the seismic refraction data. Comparisons of seismic refraction results with laboratory measurements at confining pressures equivalent to in situ lithostatic pressures provide information as to the possible composition of velocity layers. Rock types with velocities compatible with measured velocity horizons (seismic refraction) may reflect the composition of the velocity horizons, and the rock types with velocity characteristics different from the measured velocity of the oceanic layers probably reflect processes associated with the tectonic escarpments on which they were sampled.

METHOD

The specimens selected for measurement were cored from representative samples using a diamond drill with an inside diameter of 13 mm. After coring, the ends of the cylindrical specimens were cut so that they were mutually parallel. The lengths of the finished cores range from 2 to 5 cm. The specimens were jacketed in thin copper foil (2 mil thick), which kept the pressure fluid from filling pore spaces and microfractures of the specimens. To generate compressional waves, transducers with natural resonant frequencies of about 1 MHz were cemented to the ends of the specimens. Leads attached to the transducers passed through the closure of the pressure vessel. The pressure system was capable of generating pressures to 7 kb. A mixture of kerosene and petroleum ether was used as the pressurizing fluid. Pressures were read on a Heise gauge accurate to 0.1% of full scale (i.e., 0.1% of 7000 bars).

The apparatus used to measure the compressional wave velocities of the samples was that described by *Mattaboni and Schreiber* [1967]. An electric pulse applied to a piezo-electric transducer bonded to the end of the specimen initiated a compressional pulse at one end of the sample. The mechanical pulse was received by a transducer affixed to the other end of the specimen, where the pulse was converted into an electric signal, which was amplified and displayed on a dual channel oscilloscope. The time interval between the initial pulse and the received signal was determined by setting a variable frequency oscillator so that the period of one wavelength was an exact submultiple of the time delay to be measured.

When studying the velocity characteristics of rocks, it is best, if possible, to core the specimen in three mutually perpendicular directions to evaluate possible anisotropic velocity effects due to preferred orientation of mineral components or structures. Unfortunately, many of the samples recovered in the dredges were too small to be cored two or three times. Also, the emphasis of the research reported here, which represents results obtained during the initial phase of a long-range project, was to measure the compressional wave velocity of as many different rocks as possible, in an effort to establish the range of velocities associated with a given rock type. Measuring the compressional wave velocity in only one direction is probably not a serious limitation, because analysis of the hand specimens and thin sections showed that none of the samples studied had discernible penetrative anisotropic fabric. It is possible, however, that the geometry of microscopic pores and fractures could produce small (<5%) velocity variations within a sample, particularly at the lowest confining pressures (<1 kb).

The percentage of interstitial water in the deeply buried oceanic rocks of layer 2 and 3 is not known. It is important to know the amount of interstitial water because studies have shown that below confining pressures of approximately 1 kb, water saturation increases the measured compressional wave velocities, when compared with air- or oven-dried measurements, by as much as 25% [*Simmons and Nur*, 1968; *Dortman and Magid*, 1969; *Christensen*, 1970*b*]. Some of the basement rocks recovered during Legs XIII, XIV, and XV of the Deep Sea Drilling Project were recovered at localities for which seismic refraction data were available. We found during the study of these DSDP basement samples (all samples were air dried) that the measured velocities of the samples at the confining pressures equivalent to the calculated in situ pressure correlated with the refraction data [*Schreiber et al.*, 1972*b*; *Fox et al.*, 1972; *Fox and Schreiber*, in press]. These results suggested to us that, to a first approximation, air-dried laboratory samples have interstitial water contents similar to in situ samples. The samples analyzed during this study were all air-dried specimens.

Forty-six representative samples were selected from the diverse suite of rocks recovered in the dredge hauls from oceanic escarpments (Table 1). These rocks were all collected during a survey of the Kane fracture zone; the petrology of the several rock types collected have been presented in detail by *Miyashiro et al.* [1969*a*, *b*, 1970*b*, 1971]. For each sample, the compressional wave velocity was measured at 12 settings of confining pressure between

TABLE 1. Locations, Depths, and Morphologic Provinces of Samples Selected for Compressional Wave Velocity Analysis

Dredge No.	Location	Depth, fathoms	Depth, meters	Morphologic Setting
V25-D1	25°41'N 45°18'W	1725 to 1790	3155 to 3274	East wall of the rift valley
V25-D4	23°33.8'N 44°49.8'W	2300 to 2200	4206 to 4023	East wall of the rift valley at the junction of the south wall of the Kane fracture zone
V25-D5	23°31.7'N 45°07.0'W	2000 to 1850	3658 to 3383	West wall of the rift valley at the junction of the south wall of the Kane fracture zone
V25-D6	23°44.7'N 45°33.6'W	2300 to 2100	4206 to 3840	Near the base of the south wall of the Kane fracture zone between the displaced rift valley
V25-D8	23°46.7'N 46°04.2'W	2242 to 2100	4100 to 3840	Center section of the north wall of the Kane fracture zone between the displaced rift valley
V25-D9	23°46.1'N 46°37.0'W	1600 to 1450	2926 to 2652	Near the top of the south wall of the Kane fracture zone opposite the junction with the rift valley
V25-D13	24°46.7'N 50°26.1'N	3020 to 2800	5523 to 5121	Near the base of the north wall of the Kane fracture zone

atmospheric pressure (0.001 kb) and 7.00 kb. The results are presented in Table 2; these values are averages of the measurements taken at increasing pressure and decreasing pressure. With many of the specimens, the measured velocities under confining pressures exhibited a hysteresis between increasing and decreasing pressure measurements. This hysteresis is probably due to nonelastic adjustments of grain boundaries and/or pore or fracture geometry that occur under the influence of high hystrostatic pressure [*Birch*, 1961; *Christensen*, 1965]. The true velocity lies somewhere between the up-run and the down-run. With most samples studied (Table 2, Figures 1 to 5), the velocity of the compressional waves undergoes an initial rapid increase as confining pressure is increased (0.001 to 1.00 kb). This is because, at low pressures, the velocities of elastic waves are affected by the volume and geometry of the pore space within the sample. As the confining pressure increases, the space between grain boundaries closes, porosity is reduced, and the velocity increases rapidly [*Birch*, 1960, 1961; *Walsh*, 1965]. Generally, above 1 kb the voids close, and with increasing pressure, the velocities show a relatively small increase.

RESULTS

Eight basalt samples were chosen for analysis. The basalts ranged from extremely fresh basalts recovered from the rift valley of the mid-Atlantic ridge to strongly weathered basalts recovered from the western extension of the Kane fracture zone (Table 1). These samples are typical of abyssal tholeiitic basalts and are composed of phenocrysts of plagioclase and/or microphenocrysts of plagioclase and olivine set in a matrix of plagioclase, olivine, clinopyroxene, opaques, microlites, and, sometimes, glass. In the weathered samples, hydrous and oxidized phases replace the primary constituents of the specimen: glass alters to palagonite, serpentine pseudomorphs olivine, and brown alteration products pervade the groundmass. The measured velocities of the basalts range from 4.85 to 5.66 km/sec at 0.50 kb to 5.5 to 6.24 km/sec at 7.0 kb (Table 2, Figure 2). The weathered basalts have slightly lower compressional wave velocities than has the unweathered basalt. We have also measured the compressional wave velocities of 24 representative basalt samples recovered in the eastern and western North Atlantic during Legs XI and XIV of the Deep Sea Drilling Program [*Schreiber et al.*, 1972*a*; *Fox et al.*, 1972]. The compressional wave velocity envelope defined for the dredged basalt samples compares well with the velocities measured for the basalts recovered from the top of layer 2 during DSDP. At 0.50 kb confining pressure, the velocities of the DSDP basalt samples range from 3.50 to 5.73 km/sec; at 7.00 kb confining pressure the measured velocities range from 4.16 to 6.00 km/sec. Also, the compressional wave velocity envelope measured from the dredged basalts reported on here is in agreement with the velocity range defined for basalts recovered by dredging from the Juan de Fuca Ridge [*Christensen*, 1970*b*], and the crestal provinces of the mid-Atlantic ridge at 45°N [*Barrett and Aumento*, 1971] and at 22°N [*Christensen and Shaw*, 1970].

Three samples of moderately to strongly weathered dolerites were selected for analysis. These samples have granular to subophitic textures and are composed of plagioclase, pyroxene, serpentine pseudomorphs after olivine, opaques, and alteration products. The velocities of these samples range from 4.85 to 5.65 km/sec at 0.5 kb to 5.76 to 6.35 km/sec at 7.00 kb (Table 2). During Leg XV of DSDP, dolerite was sampled at sites 146 and 150 in the Caribbean [*Edgar et al.*, 1971]. At both sites the fine- to medium-grained dolerite is composed of equigranular, subophitic intergrowths of plagioclase and augite with oxides and interstitial alteration products. Thirty-four representative samples of the dolerite from sites 146 and 150 have been measured [*Fox and Schreiber*, in press, 1973]. The compressional wave velocities measured for the DSDP dolerites range from 4.08 to 5.64 km/sec at 0.5 kb; at 7.0 kb the velocities of the samples range from 5.06 to 6.03 km/sec. The range in velocities measured for the three dredged dolerite samples are similar to the measured range of velocities for the DSDP dolerites. All the dolerites measurements, however, have experienced varying degrees of alteration, and it is most likely that if these samples were unaltered, they would transmit compressional waves at higher velocities. Measurements made on fresh continental dolerites [*Anderson and Liebermann*, 1968] and one fresh dolerite

TABLE 2. Compressional Wave Velocities in Km/Sec

			Speed of Compressional Seismic Waves at Pressures of, km/sec											
Sample Number	Rock Type	Density, gm/cm^3	0.001 kb	0.25 kb	0.50 kb	0.75 kb	1.00 kb	1.50 kb	2.00 kb	3.00 kb	4.00 kb	5.00 kb	6.00 kb	7.00 kb
V25-D1-4	Basalt	2.83	5.35	5.58	5.66	5.72	5.75	5.81	5.86	5.95	6.00	6.02	6.05	6.08
V25-D4-14	Basalt	2.77	5.15	5.44	5.65	5.75	5.81	5.92	6.00	6.09	6.15	6.17	6.20	6.24
V25-D6-29	Basalt	2.75	4.75	5.00	5.08	5.14	5.16	5.22	5.28	5.35	5.44	5.50	5.55	5.55
V25-D8-6	Basalt	2.78	4.90	5.10	5.16	5.20	5.22	5.25	5.30	5.40	5.45	5.50	5.55	5.60
V25-D8-46	Basalt	2.84	5.35	5.45	5.50	5.55	5.60	5.65	5.70	5.76	5.81	5.83	5.85	5.86
V25-D13-6	Weathered basalt	2.72	4.45	4.70	4.85	4.95	5.00	5.15	5.20	5.30	5.40	5.50	5.60	5.66
V25-D13-33	Weathered basalt	2.75	4.75	4.90	4.95	5.00	5.05	5.11	5.16	5.25	5.35	5.40	5.45	5.50
V25-D6-40	Strongly weathered basalt	2.43	4.75	4.95	5.05	5.10	5.15	5.20	5.26	5.35	5.40	5.45	5.46	5.50
V25-D6-9	Moderately weathered dolerite	2.91	5.36	5.50	5.65	5.75	5.80	5.90	6.00	6.11	6.20	6.25	6.30	6.35
V25-D6-23	Strongly weathered dolerite	2.80	4.75	4.90	5.05	5.15	5.20	5.30	5.36	5.50	5.60	5.65	5.70	5.74
V25-D6-62	Strongly weathered dolerite	2.80	4.40	4.65	4.85	4.85	4.95	5.05	5.16	5.35	5.50	5.63	5.70	5.76
V25-D6-30	Zeolite meta-basalt	2.51	4.05	4.31	4.41	4.50	4.56	4.69	4.75	4.83	4.90	4.97	5.05	5.11
V25-D6-95	Zeolite meta-basalt	2.63	4.10	4.36	4.46	4.53	4.60	4.65	4.67	4.74	4.80	4.85	4.90	4.94
V25-D6-105	Zeolite meta-basalt	2.62	4.30	4.56	4.70	4.79	4.85	4.95	5.05	5.11	5.16	5.21	5.25	5.30
V25-D8-10	Zeolite meta-basalt	2.60	4.50	4.72	4.90	4.96	5.03	5.05	5.13	5.15	5.17	5.20	5.22	5.25
V25-D8-27	Zeolite meta-basalt	2.66	4.10	4.55	4.77	4.90	5.00	5.07	5.10	5.16	5.23	5.28	5.34	5.40
V25-D5-37	Chlorite-quartz meta-basalt	2.61	4.63	4.95	5.10	5.20	5.28	5.36	5.41	5.45	5.49	5.52	5.55	5.58

TABLE 2. (continued)

Sample Number	Rock Type	Density, gm/cm^3	Speed of Compressional Seismic Waves at Pressures of, km/sec											
			0.001 kb	0.25 kb	0.50 kb	0.75 kb	1.00 kb	1.50 kb	2.00 kb	3.00 kb	4.00 kb	5.00 kb	6.00 kb	7.00 kb
V25-D5-10	Actinolite-chlorite meta-basalt	2.68	4.25	4.80	5.05	5.20	5.28	5.35	5.40	5.50	5.55	5.60	5.65	5.66
V25-D6-36	Chlorite-epidote meta-basalt	2.80	5.60	5.85	5.95	6.00	6.05	6.10	6.15	6.22	6.25	6.28	6.30	6.30
V25-D6-63	Chlorite-epidote meta-basalt	2.86	5.70	5.87	6.00	6.10	6.20	6.27	6.35	6.46	6.55	6.60	6.65	6.66
V25-D6-71	Chlorite-epidote meta-basalt	2.87	5.60	5.80	5.96	6.02	6.08	6.12	6.14	6.15	6.20	6.21	6.25	6.26
V25-D5-36	Chlorite-quartz rock	2.65	4.50	5.12	5.30	5.40	5.46	5.57	5.61	5.65	5.66	5.68	5.70	5.76
V25-D9-9	Actinolite rock	2.77	4.18	5.22	5.41	5.53	5.62	5.75	5.80	5.91	6.01	6.11	6.21	6.31
V25-D9-10A	Actinolite rock	2.96	4.50	5.30	5.47	5.60	5.70	5.85	5.94	6.05	6.18	6.29	6.40	6.50
V25-D9-10B	Actinolite rock	2.97	3.80	5.45	5.62	5.76	5.86	6.05	6.19	6.45	6.65	6.76	6.87	6.98
V25-D6-6	Gabbro	2.96	6.20	6.75	6.85	6.90	6.95	7.00	7.06	7.15	7.20	7.25	7.30	7.31
V25-D6-39A	Gabbro	2.90	6.05	6.50	6.60	6.70	6.75	6.80	6.84	6.93	7.00	7.02	7.06	7.10
V25-D6-39B	Gabbro	2.89	5.88	6.60	6.72	6.81	6.86	6.92	6.95	7.00	7.05	7.13	7.18	7.21
V25-D6-41	Slightly weathered gabbro	2.90	5.15	6.05	6.22	6.32	6.40	6.48	6.55	6.62	6.70	6.75	6.80	6.85
V25-D6-38	Slightly weathered gabbro	2.89	5.15	5.65	5.80	5.90	5.98	6.10	6.15	6.25	6.33	6.42	6.51	6.60
V25-D6-163	Slightly weathered gabbro	2.84	5.10	5.65	5.95	6.05	6.15	6.26	6.32	6.45	6.56	6.65	6.70	6.75
V25-D6-28	Strongly weathered gabbro	2.66	3.75	4.35	4.55	4.67	4.75	4.90	5.00	5.15	5.25	5.30	5.40	5.45

TABLE 2. (continued)

Sample Number	Rock Type	Density, gm/cm^3	Speed of Compressional Seismic Waves at Pressures of, km/sec											
			0.001 kb	0.25 kb	0.50 kb	0.75 kb	1.00 kb	1.50 kb	2.00 kb	3.00 kb	4.00 kb	5.00 kb	6.00 kb	7.00 kb
V25-D5-3	Very slightly metamorphosed gabbro	2.96	4.35	5.58	5.68	5.78	5.85	5.95	6.05	6.20	6.33	6.45	6.53	6.61
V25-D5-52	Very slightly metamorphosed gabbro	2.93	5.10	6.10	6.30	6.40	6.44	6.50	6.57	6.65	6.72	6.79	6.83	6.86
V25-D5-32	Slightly metamorphosed gabbro	2.89	5.81	6.30	6.42	6.50	6.54	6.60	6.65	6.70	6.75	6.80	6.88	6.95
V25-D5-26	Slightly metamorphosed gabbro	2.78	5.53	6.16	6.31	6.39	6.44	6.50	6.54	6.60	6.65	6.70	6.73	6.75
V25-D5-11	Chlorite metagabbro	2.91	5.75	5.95	6.10	6.20	6.26	6.31	6.36	6.40	6.50	6.55	6.60	6.66
V25-D5-51	Chlorite metagabbro	2.84	4.06	5.20	5.45	5.57	5.60	5.65	5.71	5.80	5.90	5.98	6.06	6.16
V25-D5-14	Actinolite-chlorite metagabbro	2.82	5.00	5.50	5.66	5.81	5.92	6.04	6.10	6.20	6.25	6.31	6.40	6.45
V25-D5-18	Hornblende-actinolite-chlorite metagabbro	2.72	4.95	5.38	5.50	5.63	5.70	5.80	5.87	6.00	6.10	6.20	6.27	6.32
V25-D6-43	Actinolite metagabbro	2.87	5.76	6.25	6.40	6.45	6.50	6.58	6.60	6.65	6.70	6.75	6.80	6.82
V25-D6-58	Hornblende metagabbro	2.78	4.75	5.50	5.61	5.70	5.75	5.85	5.95	6.01	6.11	6.20	6.26	6.31
V25-D6-17	Hornblende metagabbro	2.90	5.95	6.15	6.35	6.44	6.48	6.55	6.60	6.65	6.70	6.75	6.80	6.84

TABLE 2. (continued)

Sample Number	Rock Type	Density, gm/cm^3	Speed of Compressional Seismic Waves at Pressures of, km/sec											
			0.001 kb	0.25 kb	0.50 kb	0.75 kb	1.00 kb	1.50 kb	2.00 kb	3.00 kb	4.00 kb	5.00 kb	6.00 kb	7.00 kb
V25-D5-17	Cataclastic meta-gabbro	2.64	4.40	5.10	5.20	5.28	5.32	5.43	5.50	5.57	5.61	5.66	5.73	5.78
V25-D9-5	Serpentinized peridotite	2.42	3.40	3.80	3.95	4.05	4.12	4.20	4.25	4.36	4.47	4.56	4.66	4.76
V25-D9-8	Serpentinized peridotite	2.32	3.00	3.46	3.55	3.58	3.60	3.65	3.70	3.76	3.86	3.95	4.02	4.11

sample from the Aves ridge in the Caribbean [*Fox et al.*, 1971] have compressional wave velocities 0.5 to 1.0 km/sec faster at equivalent confining pressures than the measurements on the altered dolerites.

Ten meta-basalts with the effects of zeolite-facies metamorphism and of varying degrees of greenschist metamorphism were selected for analysis. Zeolite-facies meta-basalts are characteristically cut by a network of veins composed of zeolites. Zeolites occur in the matrix of the meta-basalt in some samples. These samples are also weathered, with hydrous and oxidized phases developing at the expense of the primary silicate components. The meta-basalts in the greenschist-facies are characterized by a mineral assemblage consisting of all, or a combination of some, of the following minerals; plagioclase (sometimes Na-enriched), relict clinopyroxene, actinolite, chlorite, epidote, and quartz. The zeolite-facies meta-basalt range in velocity from 4.41 to 4.90 km/sec at 0.5 kb confining pressure to 5.66 to 6.66 km/sec at 7.0 kb (Table 2; Figure 3). *Miyashiro et al.* [1971] showed, on the basis of characteristic mineralogic and chemical composition of the greenschist meta-basalts, that they could be separated into two categories. The first group of greenschist meta-basalts is characterized by primary igneous plagioclase that has resisted recrystallization, relict primary mafic minerals if recrystallization has not been intense, and secondary actinolite and chlorite. The meta-basalts in group 1 approximately preserve their original igneous composition except for H_2O content. Meta-basalts in this category contain no epidote or sodium enriched plagioclase. The second group of the greenschist meta-basalts in the greenschist-facies is characterized by the recrystallization of the plagioclase to albite and the occurrence of all, or a combination of some, of the following minerals: chlorite, actinolite, quartz and epidote. These group 2 meta-basalts have experienced intense chemical migration during metamorphism. Greenschist meta-basalt samples representative of both groups 1 and 2 were chosen for velocity study. The velocity results show that group 1 greenschist meta-basalts (calcic plagioclase-chlorite-actinolite) and some group 2 greenschist meta-basalts (characterized by chlorite-quartz) have similar measured velocities from 1 bar to 7.0 kb con-

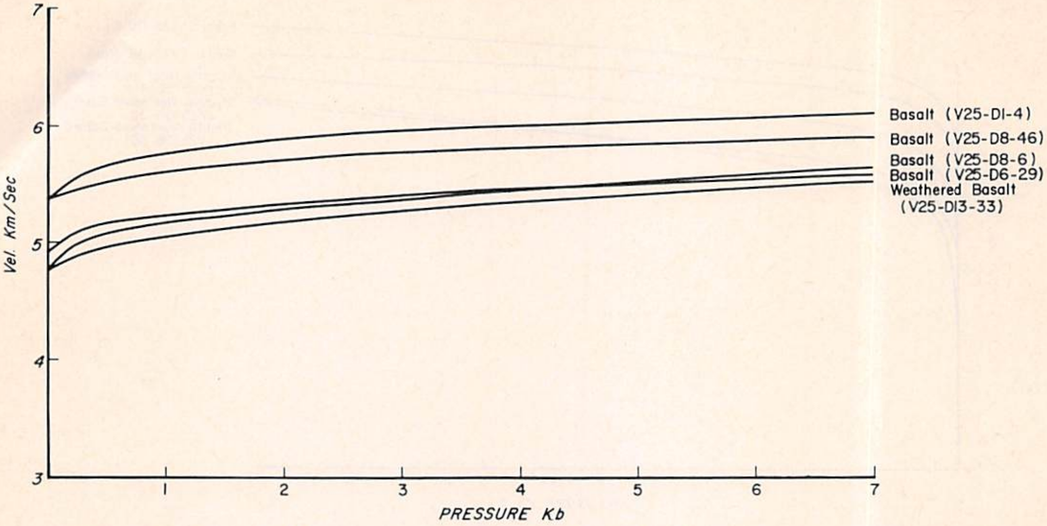

Fig. 2. The change in the compressional wave velocity as a function of pressure for basalt and weathered basalt samples.

fining pressure (see Table 2; Figure 3). These greenschist meta-basalts have a velocity range similar to the range measured for basalts and dolerites. Some group 2 greenschist meta-basalts (characterized by albitized plagioclase-epidote-chlorite-quartz), however, have a measured velocity range which is 0.2 to 0.8 km/sec higher than basalt or the other meta-basalt types (see Table 2; Figure 3). *Christensen and Shaw* [1970] and *Barrett and Aumento* [1970] have measured the compressional wave velocities of several greenschist-facies meta-basalts from the mid-Atlantic ridge, and these samples had the same velocity range as defined for all the greenschist meta-basalts from the Kane fracture zone. Unfortunately the complete petrographic descriptions are not presented in these two papers, and it is not possible to separate the meta-basalts into the groups 1 and 2 of *Miyashiro et al.* [1971].

The velocity characteristics of two relatively fresh and unaltered gabbros and five weathered gabbros were defined. The gabbros are composed of a medium- to coarse-grained assem-

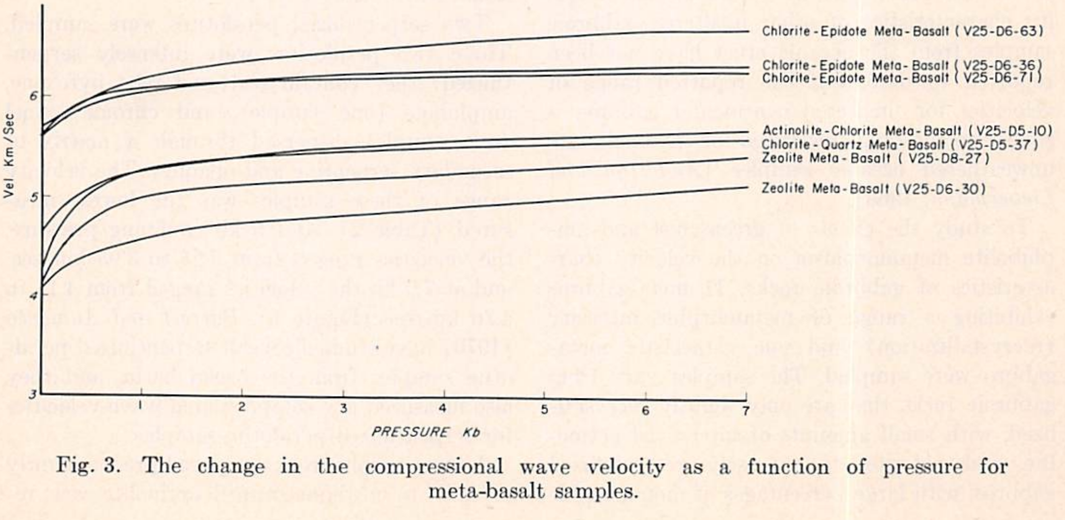

Fig. 3. The change in the compressional wave velocity as a function of pressure for meta-basalt samples.

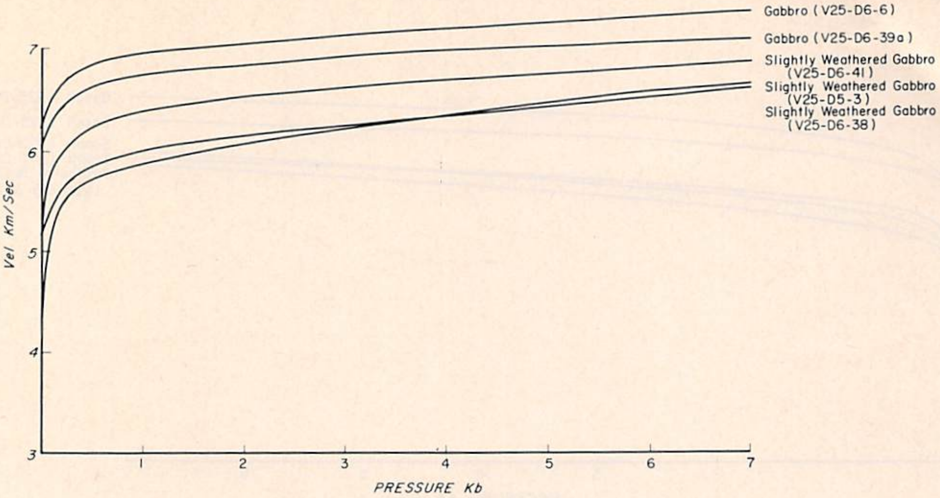

Fig. 4. The change in the compressional wave velocity as a function of pressure for gabbro samples.

blage of plagioclase, clinopyroxene, olivine in some samples, and, depending on the intensity of weathering, a varying percentage of alteration products. The fresh gabbros have a velocity range of 6.60 to 6.85 km/sec at 0.5 kb confining pressure and have a velocity range of 7.10 km/sec to 7.31 km/sec at 7 kb (Table 2; Figure 4). The development of hydrous and oxidized phases, as a result of weathering, markedly affects the compressional wave velocity characteristics of gabbros; the weathered gabbros range in velocity at 0.5 kb confining pressure from 4.55 to 6.22 km/sec and at 7 kb from 5.45 to 6.85 km/sec (Table 2; Figure 4). The velocity characteristics of other unaltered gabbroic samples from the oceanic crust have not been reported to date, but the reported range of velocities for unaltered continental gabbros is similar to the range measured for the unaltered, unweathered oceanic samples [*Anderson and Liebermann*, 1968].

To study the effects of greenschist and amphibolite metamorphism on the velocity characteristics of gabbroic rocks, 11 meta-gabbros exhibiting a range of metamorphic intensity (recrystallization) and one cataclastic meta-gabbro were sampled. The samples vary from gabbroic rocks that are only slightly recrystallized, with small amounts of interstitial actinolite and chlorite, to intensely recrystallized gabbros, with large percentages of metamorphic minerals of the greenschist-(chlorite-actinolite) or amphibolite-(hornblende) facies (Table 2). At 0.5 kb confining pressure, the measured velocities of the meta-gabbros range from 5.45 to 6.42 km/sec, and at 7 kb, the velocities range from 6.31 to 6.95 km/sec (Table 2; Figure 5). When the results are compared with the velocity characteristics defined for unweathered-unaltered gabbro samples, it is clear that the development of the greenschist- and amphibolite-facies minerals at the expense of the original mafic minerals (plagioclase, pyroxene and olivine) markedly reduces the compressional wave velocity.

Two serpentinized peridotites were sampled. These two peridotites were intensely serpentinized; they contain scattered relic pyroxene, amphibole (one sample), and chrome spinel (one sample) dispersed through a matrix of secondary serpentine and opaques. The velocity range of these samples was the lowest measured (Table 2). At 0.5 kb confining pressure, the velocities ranged from 3.55 to 3.95 km/sec, and at 7.0 kb the velocities ranged from 4.11 to 4.76 km/sec (Figure 6). *Barrett and Aumento* [1970] have studied several serpentinized peridotite samples from the ocean basin, and they also measured low compressional wave velocities for serpentinized peridotite samples.

A metamorphic rock composed predominantly of fine- to medium-grained actinolite was re-

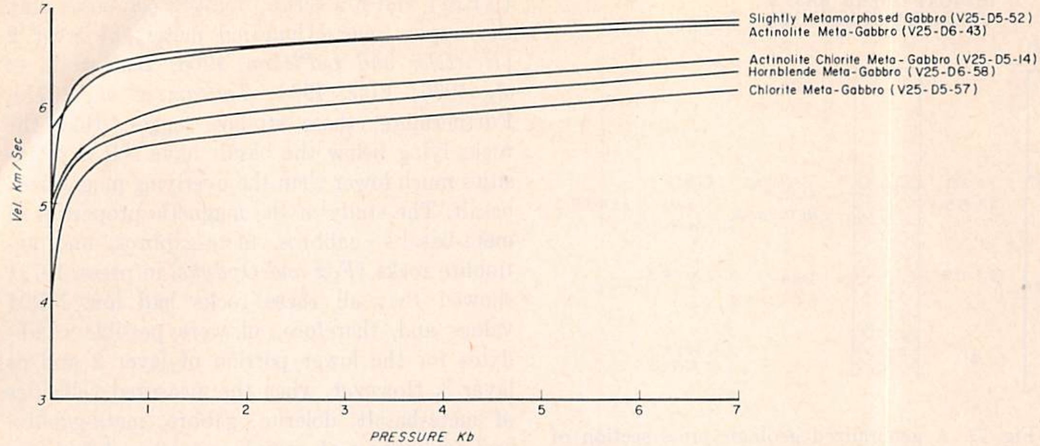

Fig. 5. The change in the compressional wave velocity as a function of pressure for meta-gabbro samples.

covered in one dredge. Three samples from two specimens were studied. Although in hand specimens and in thin section the actinolitic specimens appeared to be devoid of a penetrative fabric, two samples were taken from the largest specimen to see if the actinolite specimen was anisotropic. The velocities of these samples ranged from 5.41 to 5.62 km/sec at 0.5 kb confining pressure; at 7 kb confining pressure the velocities ranged from 6.31 to 6.98 km/sec (Figure 6).

DISCUSSION

The compressional wave velocity results presented in the preceding section define velocity ranges or velocity envelopes for basalt, meta-basalt (zeolite- and greenschist-facies), dolerite, gabbro, meta-gabbro (greenschist- and amphibolite-facies), and serpentinized peridotite. These compressional wave velocity results, when coupled with the generalized velocity structure of the oceanic crust (see Figure 1*a*), which is based on a synthesis of many published seismic refraction profiles, provide sufficient data to draw a generalized geologic cross section of the oceanic crust (Figure 7).

Layer 2 (oceanic basement) is characterized by a wide velocity range (4.0 to 6.0 km/sec), but most measurements fall within 4.4 to 5.6 km/sec [*Raitt,* 1963]. The calculated in situ

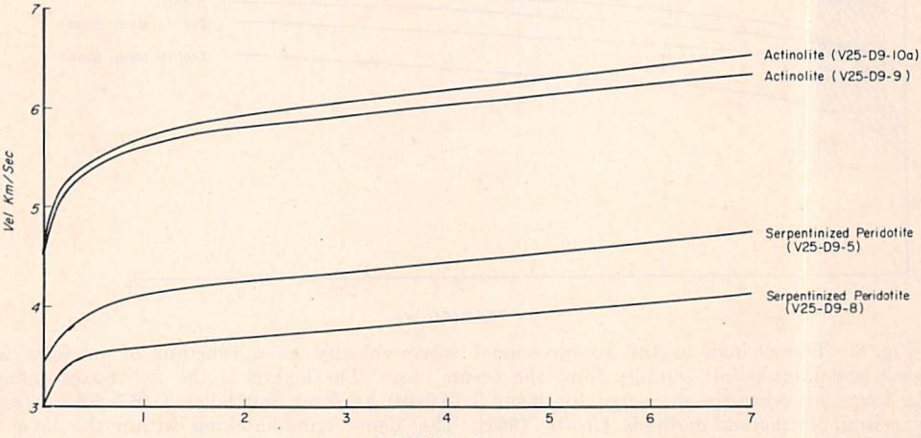

Fig. 6. The change in the compressional wave velocity as a function of pressure for serpentinized peridotite and actinolite samples.

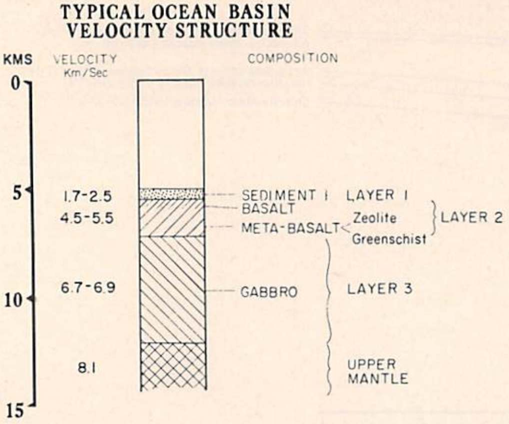

Fig. 7. A generalized geologic cross section of the oceanic crust. The diagram is based on compressional wave velocity measurements of oceanic rocks and published seismic refraction results.

lithostatic confining pressure for layer 2 ranges from 0.25 to 1 kb. The basalt samples chosen for analysis have velocity characteristics at 0.2 to 1.0 kb that are compatible with seismic refraction results. The studies of the wave lengths and amplitudes of the oceanic magnetic anomalies, however, suggest that basalt with high natural remanent magnetization intensities (NRM) and low susceptibilities composes only the upper several hundred meters of layer 2 [*Heirtzler and LePichon,* 1965; *Loncarevic et al.,* 1966; *Vine,* 1968; *Talwani et al.,* 1971]. Furthermore, these studies suggest that the rocks lying below the basalt have NRM intensities much lower than the overlying magnetized basalt. The study of the magnetic properties of meta-basalts, gabbros, meta-gabbros, and actinolite rocks [*Fox and Opdyke,* in press, 1973] showed that all these rocks had low NRM values and, therefore, all were possible candidates for the lower portion of layer 2 and/or layer 3. However, when the measured velocities of meta-basalt, dolerite, gabbro, meta-gabbro, and actinolite rock samples at 0.2 to 1.0 kb are compared with the velocities measured by seismic refraction methods for layer 2, it is noted that only meta-basalts in the zeolite- and greenschist-facies (characterized by chlorite, actinolite, and quartz) and some altered dolerite samples [*Fox and Schreiber,* in press, 1973] have compressional wave velocities that are compatible with velocities measured for layer 2 by seismic refraction methods (Table 2; Figures 8, 9). Although most measured velocity values for layer 2 fall within the range of 4.4

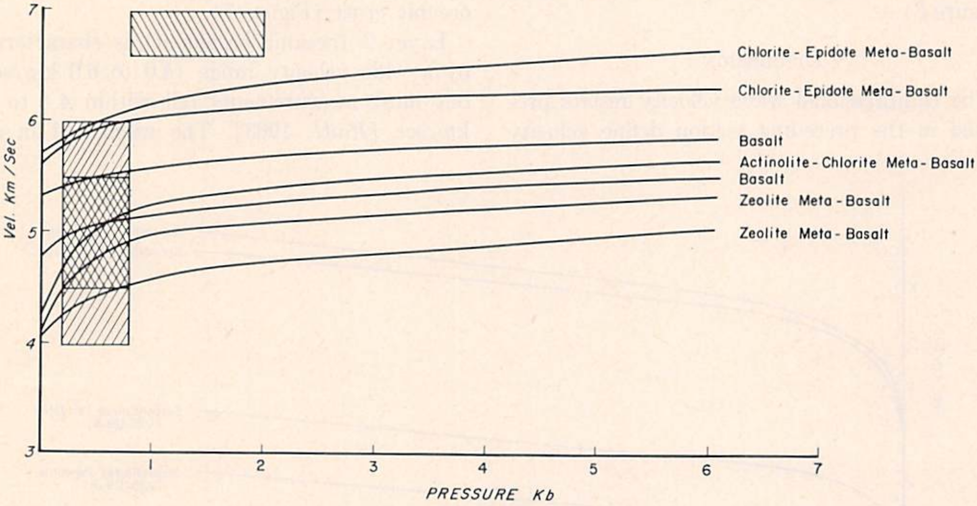

Fig. 8. The change in the compressional wave velocity as a function of pressure for basalt and meta-basalt samples from the ocean basin. The length of the two boxes defines the range of velocities measured for layer 2 (4.0–6.0 km/sec) and layer 3 (6.7–6.9 km/sec) by seismic refraction methods [*Raitt,* 1963]. The dense cross-hatching within the layer 2 velocity envelope indicates the velocity range (4.4–5.6 km/sec) in which a majority of the layer 2 velocity measurements fall. The width of the two boxes indicates the calculated in situ confining pressure range for layers 2 and 3 in the ocean basin.

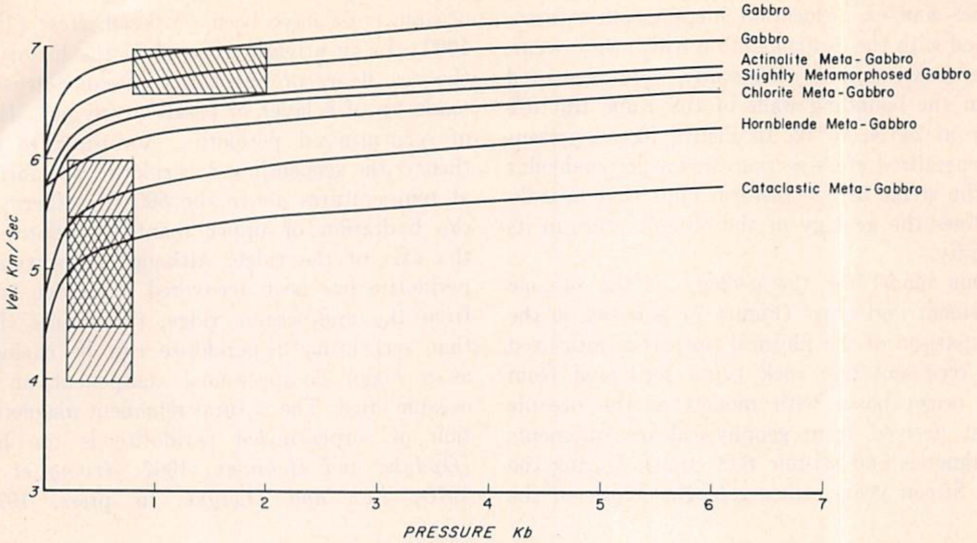

Fig. 9. The change in the compressional wave velocity as a function of pressure for gabbro and meta-gabbro samples from the ocean basin. The length of the two boxes defines the range of velocities measured for layer 2 (4.0–6.0 km/sec) and layer 3 (6.7–6.9 km/sec) by seismic refraction methods [*Raitt,* 1963]. The dense cross-hatching within the layer 2 velocity envelope indicates the velocity range (4.4–5.6 km/sec) in which a majority of the layer 2 velocity measurements fall. The width of the two boxes indicates the calculated in situ confining pressure range for layers 2 and 3 in the ocean basin.

to 5.6 km/sec [*Raitt,* 1963], in some regions (e.g., Bermuda rise, Emperor Seamount chain) velocities in the 5.6- to 6.0-km/sec range are recorded for layer 2. At 0.25 to 1.0 kb confining pressure, meta-basalts in the greenschist-facies that are characterized by Na-enriched plagioclase and epidote, as well as chlorite and actinolite, have measured compressional wave velocities in the 5.6- to 6.0-km/sec range. Also, although the results were not obtained during this study, measurements on dolerites [*Fox and Schreiber,* in press, 1973] and greenschist-facies meta-dolerites [*Christensen and Shaw,* 1970] have velocities at 0.2 to 1 kb confining pressure in the 5.6- to 6.0-km/sec range.

The majority of the published seismic refraction measurements for layer 3 (oceanic layer) fall within the range of 6.7 to 6.9 km/sec [*Raitt,* 1963; *Ludwig et al.,* 1971], thus suggesting that the composition of layer 3 is uniform. The calculated in situ lithostatic confining pressure for layer 3 is 1 to 2 kb. A diverse suite of rocks has been sampled from fracture zone escarpments, but the only rock type whose measured compressional wave velocities compare with the recorded values of layer 3 is unaltered gabbro (Table 2, Figures 8, 9). Only two unaltered gabbro samples were available for laboratory study. The velocities measured for these oceanic samples, however, are similar to velocities measured for unaltered continental gabbros [*Anderson and Liebermann,* 1968]. It is interesting to note that at 1 to 2 kb confining pressure, the unaltered gabbro samples range in velocity from 6.7 to 7.03 km/sec; the measured velocity range of the greenschist- and amphibolite-facies meta-gabbros is 5.87 to 6.6 km/sec. This apparent marked decrease in the velocity of gabbro that has experienced greenschist or amphibolite metamorphism suggests that meta-gabbro can be excluded as a major component of the 6.7 to 6.9 km/sec layer (layer 3).

The measured velocities, at the appropriate confining pressure range, of greenschist meta-basalt characterized by Na-enriched plagioclase and epidote, meta-gabbro (greenschist- and amphibolite-facies), actinolite rock, and serpentinized peridotite are not compatible with the generalized velocity structure of the oceanic crust, thus suggesting that these rock types do not occur in abundance in the oceanic crust. Yet, these rock types are often recovered in large quantities from tectonic escarpments. We suggest that these rocks are related to the tectonics and/or dislocation metamorphism associated with the escarpment on which they occur. The rocks used in this study were recovered from the bounding walls of the Kane fracture zone at 24°N, 46°W. In Figure 10, we present a generalized cross section drawn perpendicular to the strike of a transform fault that broadly outlines the geology of the oceanic crust in its vicinity.

Our model for the geology of the oceanic basement and crust (Figure 7) is based on the comparison of the physical properties measured for representative rock types recovered from the ocean basin with models of the oceanic crust derived from geophysical measurements (magnetics and seismic refraction). During the last fifteen years, other geologic models of the oceanic crust have been proposed. *Hess* [1955, 1962], in an attempt to explain the history of the sea floor, proposed that oceanic crust is made up of a layer of basalt overlying a layer of serpentinized peridotite. According to this theory, the serpentinized peridotite was formed at temperatures above the 500°C isotherm by the hydration of upper mantle peridotite at the axis of the ridge. Although serpentinized peridotite has been recovered in dredge hauls from the mid-oceanic ridge, the results show that serpentinized peridotite can be excluded as a major compositional component of the oceanic crust. The natural remanent magnetization of serpentinized peridotite is too high [*Opdyke and Hekinian*, 1967; *Irving et al.*, 1970; *Fox and Opdyke*, in press, 1973],

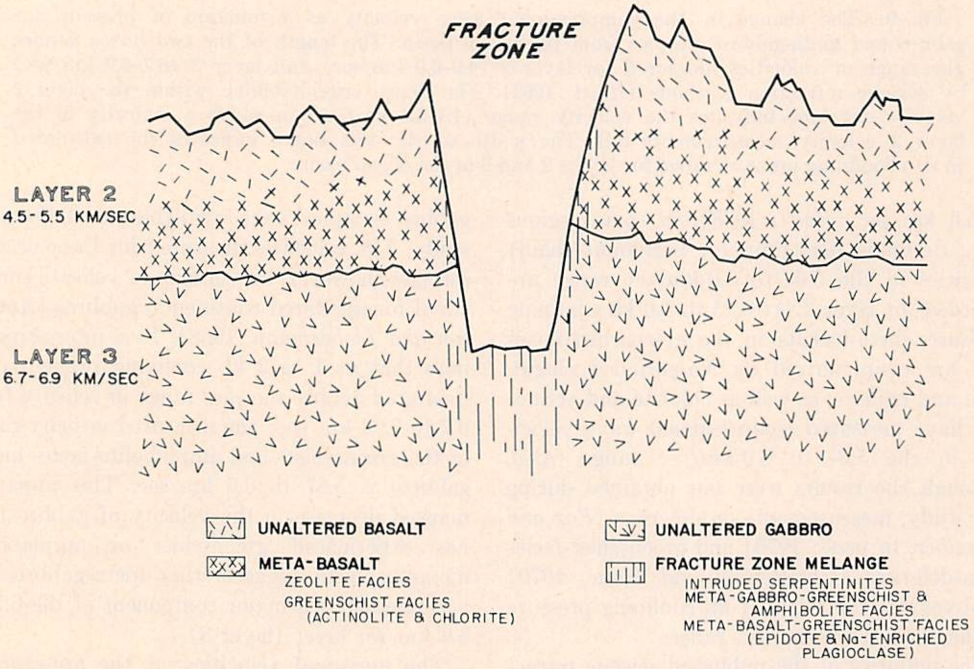

Fig. 10. A generalized geologic cross section of a transform fault based on dredging results and the data derived from magnetic and compressional wave analyses of rocks recovered in dredges. The details of the internal structure of the oceanic crust are not known, but a few suggestions are offered. In all likelihood, the unaltered basalt layer is made up of glassy, fine-grained basalt pillows, fine-grained basaltic dikes, and sills overlaying holocrystalline basaltic dikes and sills that have been metamorphosed in the zeolite- and greenschist-facies. The boundary between layer 2 and 3 must be sharp because it is an efficient refracting horizon. Layer 3 is composed predominantly of gabbro, and although the relationship of the gabbroic masses within layer 3 is undoubtedly complex, in terms of the transmission of compressional waves, the layer is apparently quite homogenous. Recent seismic refraction results (see text), however, indicate that the lower portion of layer 3 in some oceanic regions is composed of rock with a higher velocity perhaps cummulate gabbro or pyroxenite.

and the compressional wave velocities are too low to be compatible with results from the magnetic and seismic refraction surveys.

Based on the similarity of velocities measured in the ocean basin with velocities of basaltic rocks, several investigators proposed that the oceanic crust is composed of basalt, dolerite, and gabbro [*Ewing and Ewing*, 1959; *Gutenberg*, 1959; *Birch*, 1960; *Engel et al.*, 1965]. This model fails to specify, however, what type of basaltic rocks occur in the oceanic crust. Also, the presence of metamorphic basalt and metamorphic gabbro could not be inferred from the generalization that the ocean crust is composed of basaltic rock. Initially, the dredge hauls that recovered predominantly basalt and lesser amounts of gabbroic rock [*Shand*, 1949; *Engel and Engel*, 1964; *Nicholls et al.*, 1964; *Muir et al.*, 1964] supported the basaltic crust model, but as more dredge hauls containing meta-basalt (zeolite- and greenschist-facies) and meta-gabbro (greenschist- and amphibolite-facies) were recovered from the ocean basin [*Cann and Vine*, 1966; *Melson and van Andel*, 1968; *Miyashiro et al.*, 1971], it appeared that a large proportion of the oceanic crust had been altered by low-temperature metamorphic effects. An abundance of meta-basalts in the greenschist-facies (characterized by epidote, Na-enriched plagioclase, chlorite, actinolite, and quartz) were found in many dredge hauls, and the occurrence of these rocks suggested to some investigators [*van Andel and Bowin*, 1968; *Cann*, 1968; *Christensen*, 1970*a*] that the lower portion of oceanic basement (layer 2) is composed of greenschist meta-basalt of this type. Although more measurements are needed, the compressional wave velocities measured for this kind of greenschist meta-basalt appear to be too high (5.8 to 6.2 km/sec at 0.25 to 1.0 kb) to be compatible with the majority of the seismic refraction measurements of oceanic basement (layer 2), thus suggesting that this kind of greenschist meta-basalt results from localized metamorphic effects associated with the tectonic escarpments from which the samples were recovered. *Cann* [1968] and *Christensen* [1970] extrapolated downward the metamorphic grade (greenschist) observed in samples recovered by dredging and postulated that the oceanic layer (layer 3) is composed of basaltic rock that has been metamorphosed to amphibolite (hornblende and plagioclase). Few amphibolite samples have been recovered in dredge hauls from the ocean basin, and on this basis alone it is doubtful that amphibolite is an important constituent of the ocean crust. Although amphibolites have not been analyzed during this study, published compressional wave measurements of continental amphibolites range from 6.2 to 7.5 km/sec at 1 to 2 kb [*Christensen*, 1965; *Anderson and Liebermann*, 1968]. A rock type with such strong anisotropy is not a reasonable candidate for a major amount of the oceanic layer (layer 3) because the measured velocity range of the oceanic layer (layer 3) is 6.7 to 6.9 km/sec. After studying the petrology of rocks recovered from the Kane and Atlantis fracture zones, *Miyashiro et al.* [1969*a*, 1970*a*] proposed that the lower part of the oceanic basement (layer 2) and the oceanic layer (layer 3) are composed of metamorphosed basalt and gabbro in the zeolite-, greenschist-, and amphibolite-facies. Based on compressional wave velocity measurements of meta-basalts and meta-gabbros made during this study, it seems unlikely that these metamorphic rocks are volumetrically an important constituent of the oceanic layer (layer 3) because the measured velocities are as much as 1.0 km/sec below the average velocities recorded for the oceanic layer (layer 3). Rather, the compressional wave velocity measurements suggest that zeolite and greenschist (chlorite, actinolite, and quartz) meta-basalt make up the lower portion of the oceanic basement (layer 2) and that the occurrence of strongly metamorphosed gabbro (greenschist- and amphibolite-facies) is a function of dislocation metamorphism associated with fracture zone tectonics. *Cann* [1970] has recently revised his model of the oceanic crust and has proposed that oceanic basement (layer 2) is composed of pillow basalts and feeder dikes that overlie a gabbroic oceanic layer (layer 3). In this model, zeolite, greenschist, and amphibolite metamorphic facies are restricted to the lower portion of the oceanic basement (layer 2). The model proposed by Cann is similar to the geologic model of the oceanic crust arrived at in this study.

It may, however, be a gross oversimplification to constrain the measured compressional wave velocity results of oceanic rocks to the standard or averaged velocity profile of the oceanic crust.

If, for example, a frequency histogram of recorded velocities is made from published seismic refraction results from the world's oceans (Figure 11), it is clear that there is, in fact, a broad range of recorded velocities for layers 2 and 3. Some of the scatter may indeed be due to instrumentational error, but the range of velocities and the number of measurements for a given velocity interval is large enough that the distribution is probably meaningful, and therefore, it may not be wise to enthusiastically embrace the simplistic standard velocity profile of the oceanic crust.

Several recent studies document that the velocity structure of at least some regions of the oceanic crust may be more complex than expected. For example, the results of a detailed sonobuoy refraction study of the Reykjanes ridge show that layer 2 can be divided into two layers [*Talwani et al.*, 1971]. The upper portion, designated as layer 2A, corresponds to the highly magnetized layer that produces the magnetic anomalies and has a low recorded velocity range (2.33 to 3.84 km/sec). The measured velocities of basalts are much higher than the velocities recorded for layer 2A, but when one considers that layer 2A is the top of a broken and fractured volcanic pile composed of extruded, glassy, pillow basalts, shallow dikes and sills, and interbedded sediments, it is not difficult to understand why the recorded velocities are so low. The lower portion, designated as layer 2B, is characterized by a velocity range (4.05 to 5.92 km/sec) that is similar to the velocities normally recorded for layer 2. As discussed previously, laboratory measurements show that meta-basalts and meta-dolerites in the zeolite- and greenschist-facies have velocities in this range. The velocities recorded for layer 3 ranged from 6.0 to 6.5 km/sec. These recorded velocities are similar to the velocities measured for meta-gabbro in the greenschist- and amphibolite-facies (Table 2). Consequently, we would suggest that in the area of the Reykjanes ridge, layer 3 is not composed of unaltered gabbro, but rather of meta-gabbro in the greenschist- and/or amphibolite-facies.

The complexities of layers 2 and 3 have been documented in other areas as well. Two sonobuoy studies of the East Pacific rise north and south of the equator show that in these areas the crust of the ocean basin is characterized by a velocity structure similar to the Reykjanes ridge structure (P. Buhl, personal communica-

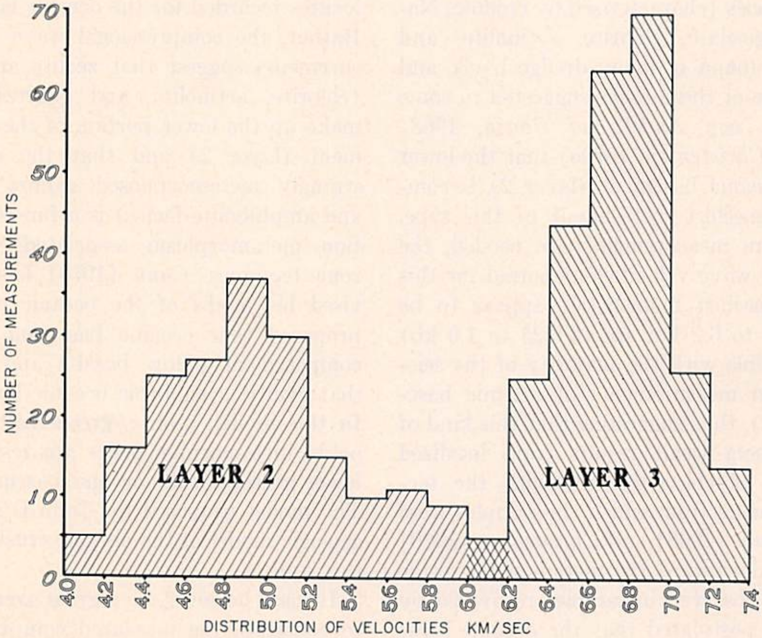

Fig. 11. Histogram of measured velocities for layer 2 (oceanic basement) and layer 3 (oceanic layer) in the main ocean basins.

tion. Sonobuoys measurements by *Maynard* [1970] and *Sutton et al.* [1971] in the central Pacific identify a two-layer structure for the layer 3 (oceanic layer). Lying below a 2- to 3-km-thick upper layer with a recorded velocity range of 6.5–6.8 km/sec lies a higher velocity 2- to 5-km-thick basal layer with a recorded velocity range of 7.0–7.7 km/sec (average 7.4 km/sec). To date we have not analyzed a rock type that transmits compressional waves under confining pressures of 1–2 kb in the 7.0- to 7.7-km/sec range. Accepting measurements on continental rocks of equivalent composition as viable analogues, this high-velocity component of the oceanic layer may represent coarsely crystalline, cummulate gabbro or perhaps a more mafic rock type (pyroxenite). Rocks of these types have been recovered from Equatorial Atlantic fracture zones [*Bonatti et al.,* 1971].

These recent sonobuoy refraction results show that the velocity structure of the ocean basin is in some regions more complex than originally thought, thus suggesting that in the future it would be best to couple geophysical studies with dredging and/or drilling programs. This would enable an investigator to correlate the physical properties of the recovered rock types with geophysical profiles attained in the same area.

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