THE ABUNDANCE OF SERPENTINITES IN THE OCEANIC CRUST[1]

NIKOLAS I. CHRISTENSEN
Department of Geological Sciences and Geophysics Program,
University of Washington, Seattle, Washington 98195

ABSTRACT

New measurements of compressional- and shear-wave velocities in serpentinites from the Mid-Atlantic Ridge at 43° N, 1° N, and 4° S are reported at pressures to 10 kbar. At oceanic crustal pressures, compressional-wave velocities (V_p) in serpentinites are between 4.2 and 5.2 km/sec and shear-wave velocities (V_s) range from 2.1 to 2.5 km/sec. The ratios of V_p to V_s for serpentinites are much higher than those measured for other rocks believed to be abundant in the oceanic crust. Poisson's ratios of serpentinites vary between 0.34 and 0.38. Comparisons of velocities and Poisson's ratios of serpentinites and partially serpentinized peridotites with similar data determined by seismic refraction studies demonstrate that serpentine is not a major constituent of the oceanic crust. Consequently, oceanic serpentinites are interpreted as forming in fracture zones by the hydration of peridotites which have been emplaced into the fracture zones by tectonic processes. An oceanic crust composed primarily of mafic igneous rocks and their metamorphosed equivalents is in better agreement with comparisons of laboratory measured elastic properties with seismic refraction data.

INTRODUCTION

Serpentinites are common constituents of tectonically active areas including mountain belts, island areas, and mid-ocean ridges. The geologic significance of serpentinites in these regions has been emphasized by many (e.g., Hess 1955; Dietz 1963; Raleigh and Paterson 1965; Coleman 1970*a*). Serpentinites from oceanic areas are often associated with fracture zones and trenches. Quon and Ehlers (1963), Nicholls et al. (1964), Muir and Tilley (1966), and Miyashiro et al. (1969) have described serpentinites recovered near fracture zones which cut the Mid-Atlantic Ridge at latitudes between 24° and 30°N. The Vema Fracture Zone in the North Atlantic (9°–11°N) contains exposures of serpentinized peridotites (Melson and Thompson 1971). Serpentinites have been collected north of a complex fracture zone which interrupts the Mid-Atlantic Ridge crest near 43°N lat (Phillips et al. 1969). Bonatti (1968), Bonatti et al. (1971), and Melson and Thompson (1970) have reported serpentinites collected along the St. Paul, Romanche, and Chain fracture zones in the vicinity of the Mid-Atlantic Ridge at latitudes 2°N–4°S. Serpentinites also have been collected near the intersection of the Rodriguez Fracture Zone and the Mid-Indian Ridge (Hekinian 1968; Engel and Fisher 1969), and from the crest of the Mid-Indian Ridge (Chernysheva and Bezrukov 1966). Bowin et al. (1966) and Chase and Hersey (1968) have described serpentinites from the Puerto Rico Trench.

Textural studies show that many of the oceanic serpentinites were derived from peridotite. Often the rocks are completely serpentinized, with lizardite being the most common serpentine mineral. Although small differences in CaO, Al_2O_3, TiO_2, K_2O, and FeO contents have been reported which may be of petrologic significance, the overall chemistry of oceanic serpentinites is relatively uniform (Miyashiro et al. 1969).

Alternate interpretations are possible to explain the presence of oceanic serpentinites. It has been proposed by Hess (1962, 1965) and Dietz (1963) that serpentine is a major constituent of the lower oceanic crust (layer 3). According to this theory, large volumes of partially serpentinized peridotite are envisioned as being generated under oceanic ridges by hydration of mantle peridotite. Layer 3 is believed to originate by lateral movement of this partially ser-

[1] Manuscript received February 18, 1972; revised April 24, 1972.

[JOURNAL OF GEOLOGY, 1972, Vol. 80, p. 709–719]

pentinized peridotite away from the oceanic ridges. Melson and Thompson (1971) have suggested that fracture zones expose sections of oceanic crust, and the abundance of serpentinite along fracture zones indicates that it is a major constituent of the lower oceanic crust. Bailey et al. (1970) interpret serpentinite and partially serpentinized peridotite above the Coast Range thrust in California as the basal part of Mesozoic oceanic crust. An alternate explanation is that the serpentinites are not major constituents of the oceanic crust, but represent slices of hydrated peridotite which are locally abundant along major fracture zones (Oxburgh 1967; Cann 1968; Christensen 1970*b;* Melson and Thompson 1971; Miyashiro et al. 1970). The question is thus one of abundance of serpentine in the oceanic crust.

With the advent of recent theories in global tectonics, the composition of the oceanic crust has become critical to an understanding of the geologic processes operating at ridge crests. Structure of the oceanic crust is largely defined on the basis of seismic velocity. Petrologic models of ridge crests and the oceanic crust must satisfy restrictions imposed by comparisons of oceanic refraction seismic velocities with laboratory measurements of velocities in rocks. In the absence of information on velocities of oceanic rocks, it has been difficult to evaluate critically the many compositional models that have been proposed for oceanic crustal regions.

In this paper, laboratory data on the elasticity of oceanic serpentinites are presented to pressures of 10 kbar. The data, when compared with marine seismic refraction data, show that serpentine is not an abundant constituent of the oceanic crust. It therefore appears that some process other than serpentinization of mantle peridotite must account for the formation of oceanic crust.

ELASTICITY OF OCEANIC SERPENTINITES

Compressional- and shear-wave velocities and densities are given in table 1 for three serpentinites dredged from the Mid-Atlantic Ridge. To determine the degree of anisotropy, velocities were measured in three perpendicular directions for each sample. The cores used for the measurements were 2.5 cm in diameter and between 5 and 7 cm in length. Densities were calculated from the dimensions and weights of the cores. For comparisons with previous measurements, the velocities and densities in table 1 are uncorrected for changes in length due to compression at high pressures.

Average velocities and densities at selected pressures corrected for length changes are given in table 2. Corrections were made for compression using an iterative routine and the dynamically determined compressibilities. Poisson's ratio (σ), the seismic parameter (ϕ), the bulk modulus (K), compressibility (β), the shear modulus (μ), Young's modulus (E), and Lamé's constant (λ) given in table 2 were calculated from the corrected velocities and densities assuming isotropic elasticity.

The velocities were measured by a pulse transmission technique similar to that described by Birch (1960) and Christensen and Shaw (1970). Barium titanate and a-c-cut quartz transducers of 1–3 MHz frequencies were used to generate the compressional and shear waves. Pressure was generated in a cavity 3.4 cm in diameter and 15.2 cm long by a two-stage pumping technique. Pressure was measured by means of a calibrated manganin coil. Temperatures of all runs were between 20° and 30°C.

PETROLOGY AND CHEMISTRY

Sample AII-32-8-4 was obtained from the Mid-Atlantic Ridge from cruise 32 of the research vessel R. V. *Atlantis II* near 43°N lat. The serpentinites from this area have been previously described by Phillips et al. (1969). Although minor antigorite is present, lizardite is the most abundant serpentine mineral in this rock. The X-ray diffraction patterns described by Page and Coleman (1967) have been used for identification of the serpentine minerals. Pseudomorphs of lizardite after pyroxene are common. The only other minerals identified in this specimen are magnetite and olivine, which form less than 5% of the rock.

Sample AII-20-26-118 was dredged near the Mid-Atlantic Ridge at 1°N in the vicinity of St. Paul's Rocks. The sample is extremely fresh; lizardite is the abundant serpentine mineral. In addition, the sample contains approximately 6% olivine occurring as fine-grained mylonitized aggregates completely enclosed in the serpentine matrix. Fine-grained magnetite and chromite are also present.

Sample AII-42-2-4 was obtained from the Mid-Atlantic Ridge at 4°S. The rock is highly weathered and contains over 95% lizardite. Hydrogarnet, magnetite, chrysotile, and olivine are also present.

Chemical analyses of the three serpentinites are given in table 3 along with their mean atomic weights, which have been shown by Birch (1961) to be important parameters in velocity-density relationships. The analyses are similar to previously reported serpentinite analyses from oceanic regions (Miyashiro et al. 1969; Hekinian 1968; Bowin et al. 1966; Bonatti 1968; Melson and Thompson 1971).

DISCUSSION

Barrett and Aumento (1970) have reported compressional-wave velocities to 1 kbar for a serpentinite from the Mid-Atlantic Ridge. Compressional-wave velocities for this rock were approximately 2 km/sec. The rock studied was highly weathered with a density of only 2.1 g/cm³. Previous measurements of velocities in fresh serpentinites at elevated pressures are limited to specimens from continental regions (Birch 1960; Christensen 1966*a*), and from the AMSOC core hole near Mayaguez, Puerto Rico (Birch 1964). Since the latter

TABLE 1

COMPRESSIONAL (P)- AND SHEAR (S)-WAVE VELOCITIES (KM/SEC)

Identification Number	Density (g/cc)	Mode	$p=$ 0.2 kbar	$p=$ 0.4 kbar	$p=$ 0.6 kbar	$p=$ 0.8 kbar	$p=$ 1.0 kbar	$p=$ 2.0 kbar	$p=$ 4.0 kbar	$p=$ 6.0 kbar	$p=$ 8.0 kbar	$p=$ 10.0 kbar
AII-42-2-4-A	2.458	P	4.19	4.37	4.49	4.56	4.626	4.810	5.077	5.253	5.377	5.455
AII-42-2-4-B	2.458	P	4.20	4.34	4.45	4.54	4.593	4.788	5.052	5.230	5.352	5.430
AII-42-2-4-C	2.486	P	4.29	4.44	4.57	4.63	4.689	4.881	5.125	5.286	5.397	5.470
Mean	2.467	P	4.23	4.38	4.50	4.58	4.636	4.826	5.085	5.256	5.375	5.452
AII-42-2-4-A	2.458	S	2.15	2.19	2.22	2.25	2.285	2.378	2.492	2.551	2.585	2.603
AII-42-2-4-B	2.458	S	2.11	2.14	2.21	2.24	2.267	2.374	2.485	2.546	2.581	2.601
AII-42-2-4-C	2.486	S	2.17	2.20	2.24	2.29	2.309	2.392	2.500	2.553	2.580	2.590
Mean	2.467	S	2.14	2.18	2.22	2.26	2.287	2.381	2.492	2.550	2.582	2.598
AII-32-8-4-A	2.474	P	4.92	5.01	5.07	5.12	5.148	5.263	5.429	5.538	5.613	5.658
AII-32-8-4-B	2.453	P	4.71	4.83	4.88	4.91	4.930	5.090	5.281	5.409	5.497	5.555
AII-32-8-4-C	2.478	P	4.65	4.78	4.87	4.91	4.930	5.068	5.254	5.375	5.457	5.503
Mean	2.468	P	4.76	4.87	4.94	4.98	5.003	5.140	5.321	5.441	5.522	5.572
AII-32-8-4-A	2.474	S	2.43	2.44	2.45	2.46	2.471	2.511	2.563	2.595	2.611	2.619
AII-32-8-4-B	2.453	S	2.25	2.30	2.32	2.35	2.369	2.433	2.485	2.506	2.516	2.520
AII-32-8-4-C	2.478	S	2.30	2.33	2.35	2.35	2.377	2.440	2.495	2.518	2.528	2.532
Mean	2.468	S	2.33	2.36	2.37	2.39	2.450	2.461	2.514	2.539	2.551	2.557
AII-20-26-118-A	2.555	P	5.06	5.09	5.13	5.17	5.185	5.291	5.480	5.611	5.688	5.715
AII-20-26-118-B	2.507	P	4.47	4.53	4.56	4.62	4.666	4.811	5.069	5.248	5.358	5.414
AII-20-26-118-C	2.544	P	4.44	4.51	4.58	4.62	4.659	4.766	4.981	5.135	5.230	5.276
Mean	2.546	P	4.66	4.71	4.76	4.80	4.837	4.956	5.177	5.331	5.425	5.468
AII-20-26-118-A	2.555	S	2.08	2.12	2.15	2.18	2.201	2.245	2.308	2.350	2.385	2.412
AII-20-26-118-B	2.507	S	2.15	2.18	2.20	2.21	2.219	2.262	2.322	2.364	2.390	2.410
AII-20-26-118-C	2.544	S	2.22	2.24	2.26	2.27	2.279	2.335	2.387	2.417	2.432	2.440
Mean	2.546	S	2.15	2.18	2.20	2.22	2.233	2.281	2.339	2.377	2.402	2.421

TABLE 2

AVERAGE VELOCITIES, DENSITIES, AND ELASTIC CONSTANTS CORRECTED FOR DIMENSION CHANGES

Pressure (kbar)	V_p (km/sec)	V_s (km/sec)	ρ (g/cm^3)	σ	ϕ ([km/sec]^2)	K (Mb)	β (Mb^{-1})	μ (Mb)	E (Mb)	λ (Mb)
AII-20-26-118										
0.4	4.709	2.180	2.548	0.36	15.84	0.40	2.48	0.12	0.33	0.32
1.0	4.833	2.231	2.551	0.36	16.72	0.42	2.34	0.13	0.35	0.34
2.0	4.949	2.278	2.557	0.37	17.57	0.45	2.23	0.13	0.36	0.36
6.0	5.311	2.368	2.574	0.38	20.73	0.53	1.87	0.14	0.40	0.44
10.0	5.436	2.401	2.591	0.38	21.82	0.57	1.77	0.15	0.41	0.47
AII-32-8-4										
0.4	4.872	2.354	2.470	0.35	16.35	0.40	2.48	0.14	0.37	0.31
1.0	4.999	2.403	2.473	0.35	17.29	0.43	2.34	0.14	0.39	0.33
2.0	5.132	2.457	2.478	0.35	18.29	0.45	2.21	0.15	0.40	0.35
6.0	5.420	2.529	2.496	0.36	20.85	0.52	1.92	0.16	0.43	0.41
10.0	5.538	2.542	2.512	0.37	22.06	0.55	1.80	0.16	0.44	0.45
AII-42-2-4										
0.4	4.384	2.176	2.470	0.34	12.91	0.32	3.14	0.12	0.31	0.24
1.0	4.632	2.285	2.473	0.34	14.49	0.36	2.79	0.13	0.35	0.27
2.0	4.818	2.377	2.479	0.34	15.68	0.39	2.57	0.14	0.38	0.30
6.0	5.234	2.539	2.498	0.35	18.79	0.47	2.13	0.16	0.43	0.36
10.0	5.417	2.581	2.514	0.35	20.46	0.51	1.94	0.17	0.45	0.40

TABLE 3

CHEMICAL ANALYSES*

Specimen	AII-42-2-4	AII-20-26-118	AII-32-8-4
SiO_2	39.30	42.70	41.30
Al_2O_3	1.83	1.06	0.79
TiO_2	0.04	0.03	0.03
FeO	2.23	2.37	1.57
Fe_2O_3	6.19	6.58	4.32
MgO	37.60	33.50	37.90
CaO	0.94	1.40	0.64
K_2O	0.01	0.02	0.01
Na_2O	0.05	0.13	0.08
H_2O^+	11.90	13.60	13.30
Total	100.09	101.39	99.94
Mean atomic weights	20.60	20.60	20.30

* K. V. Campbell, analyst.

velocities were reported for only one direction (parallel to the core axis), the isotropic nature of these rocks was not established. Many of the samples reported to be serpentinite contain appreciable amounts of olivine, magnetite, and chromite and therefore have velocities and densities higher than relatively pure serpentinites. This is illustrated in figure 1. The new data for oceanic serpentinites (fig. 1, *open circles*) are similar to velocities reported by Birch (1964) and Christensen (1966*a*) for serpentinites with similar densities.

Measurements by Birch (1961) suggest that antigorite has a higher compressional-wave velocity than chrysotile. This is based on measurements for a serpentinite from Ludlow, Vermont, identified as antigorite (V_p at 2 kbar = 6.59 km/sec, ρ = 2.60

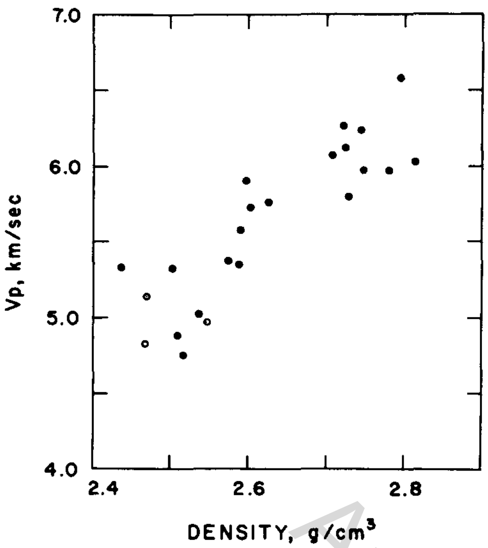

FIG. 1.—Compressional-wave velocity at 2 kbar versus density for serpentinites reported by Birch (1960, 1964), Christensen (1966*a*), and this paper.

g/cm³). The high velocity for the antigorite specimen may be due in part to the rather large anisotropy reported for this specimen. However, based on new information on the chemical differences of the serpentine minerals (Page 1968) and other physical properties, Coleman (1971*b*) concludes that antigorite may have higher seismic velocities.

The elastic properties of serpentine are quite different from those of the common rock-forming silicates (Christensen 1966*a*). This is illustrated in table 4, where the average elastic constants at 10 kbar of oceanic serpentinites are summarized along with the elastic properties of Mid-Atlantic Ridge mafic rocks and peridotite nodules. The high Poisson's ratio (σ) for serpentinite is particularly significant. Because of this, σ should be a useful parameter for identifying regions which are underlain by serpentinite. Furthermore, since the effects of both pressure and temperature on σ appear to be minimal (e.g., Anderson et al. 1968; Hughes and Maurette 1956), seismic determination of σ can be readily compared with laboratory data.

If isotropic elasticity is assumed, Poisson's ratio is related to compressional- and shear-wave velocities by the relationship

TABLE 4

AVERAGE ELASTIC CONSTANTS FOR SELECTED ROCKS

Elastic Constant	Average Serpentinite	Dolerite, Mid-Atlantic Ridge*	Basalt, Mid-Atlantic Ridge*	Peridotite Nodule†
σ	0.37	0.25	0.28	0.28
ϕ ([km/sec]²)	21.90	23.40	25.40	40.30
K (Mb)	0.53	0.68	0.74	1.34
β (Mb⁻¹)	1.88	1.46	1.36	0.75
μ (Mb)	0.16	0.42	0.39	0.69
E (Mb)	0.43	1.05	0.99	1.77
λ (Mb)	0.43	0.40	0.48	0.88

* Christensen and Shaw 1970.
† Christensen 1966*a*.

$$\sigma = \frac{1}{2}\left[1 - \frac{1}{(V_p/V_s)^2 - 1}\right].$$

This is illustrated in figure 2. This relationship can be used to calculate σ from marine seismic refraction studies in which both compressional- and shear-wave velocities have been reported.

A limited amount of data have been reported for shear-wave velocities in the oceanic crust, but what data are available are extremely significant. Recent refraction shear velocities are summarized in table 5 along with corresponding compressional-wave velocities and calculated values of σ. Poisson's ratio for oceanic crustal rocks appears to be approximately 0.28. Much of the scatter is probably due to problems in accurately identifying the secondary arrivals of the seismic energy. These observed values of σ are clearly less than the average σ (0.37) measured for oceanic serpentinites.

The hypothesis of a serpentine-rich oceanic crust proposed by Hess (1962, 1965) assumes that layer 3 is partially serpentinized peridotite. The average layer 3 compressional-wave velocity of 6.7 km/sec (Raitt 1963) corresponds to laboratory measurements of compressional-wave velocities at 2 kbar in peridotites approxi-

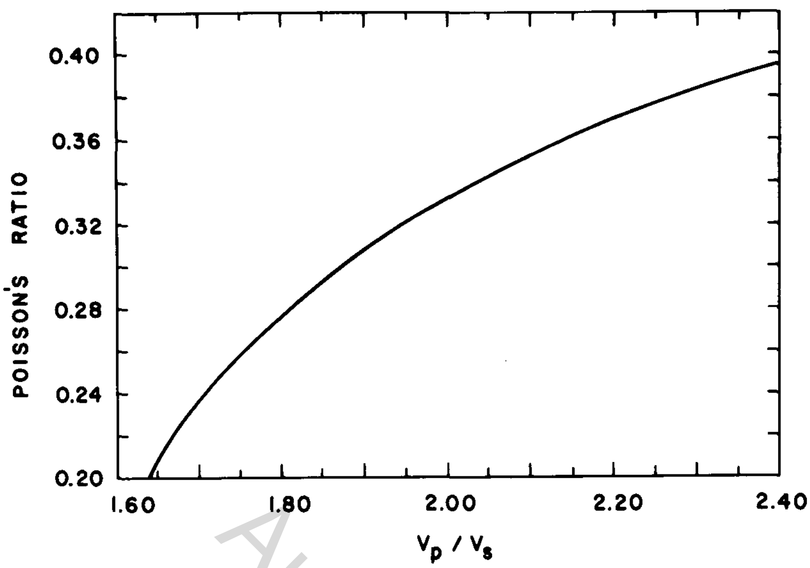

FIG. 2.—Poisson's ratio versus the ratio of compressional- to shear-wave velocity

TABLE 5

COMPRESSIONAL- AND SHEAR-WAVE VELOCITIES AND POISSON'S RATIOS FOR THE OCEANIC CRUST

Location	V_p (km/sec)	V_s (km/sec)	σ
North of Puerto Rico Trench*	6.72	3.79	0.27
	6.54	3.88	0.21
	6.77	4.10	0.21
	6.65	3.84	0.25
	6.65	3.88	0.24
Indian Ocean†	6.69	3.61	0.29
	6.42	3.51	0.29
	5.41	3.01	0.28
	6.11	3.36	0.28
	5.97	3.47	0.25
	5.54	3.08	0.28
Indian Ocean‡	6.55	3.60	0.28
Philippine Sea§	6.82	3.81	0.27
	6.83	3.78	0.28
	6.60	3.68	0.28
Northwest Pacific Basin‖	6.90	3.74	0.29
	6.90	3.89	0.27
North Pacific#	6.35	3.55	0.27
	6.94	3.85	0.28

* Data from Bunce and Fahlquist 1962.
† Data from Francis and Shor 1966.
‡ Data from Francis and Raitt 1967.
§ Data from Murauchi et al. 1968.
‖ Data from Den et al. 1969.
\# Data from Helmberger and Morris 1969.

mately 40%–50% serpentinized (fig. 3). Peridotite 50% serpentinized has a density of 2.9 g/cm³ and a σ of 0.33. This value of σ is also much higher than σ for the lower oceanic crust. Layer 2, if composed of partially serpentinized peridotite under a thin covering of basalt, should have a higher degree of serpentinization and correspondingly higher values of σ. This clearly does not agree with the seismic observations of Francis and Shor (1966), which suggest that σ increases with depth in oceanic crust.

Helmberger and Morris (1969) have recently presented a detailed seismic model of the oceanic crust north of the Hawaiian Islands. Their velocity distributions, shown in figure 4, were obtained from observations of both travel times and amplitudes of closely spaced records along a 120-km profile. Shear-wave velocities in this model were directly observed for the lower 5 km of the crust. The model presented by Helmberger and Morris shows several important features on velocity distributions within the oceanic crust. Rather than the simple three-layer model commonly found in oceanic refraction studies, they found a strong, positive velocity gradient in the

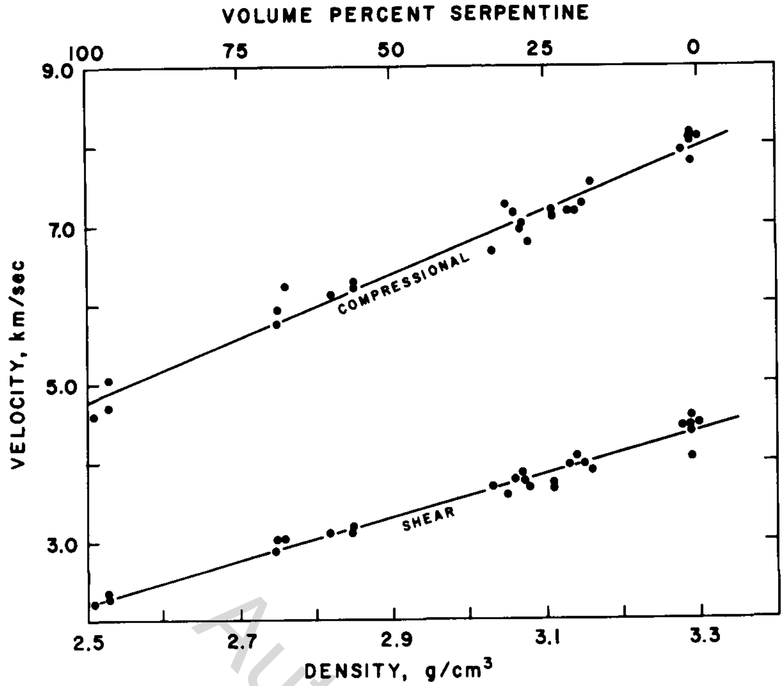

FIG. 3.—Velocity at 10 kbar versus density and vol % serpentine for peridotites and serpentinized peridotites (after Christensen 1966*a*).

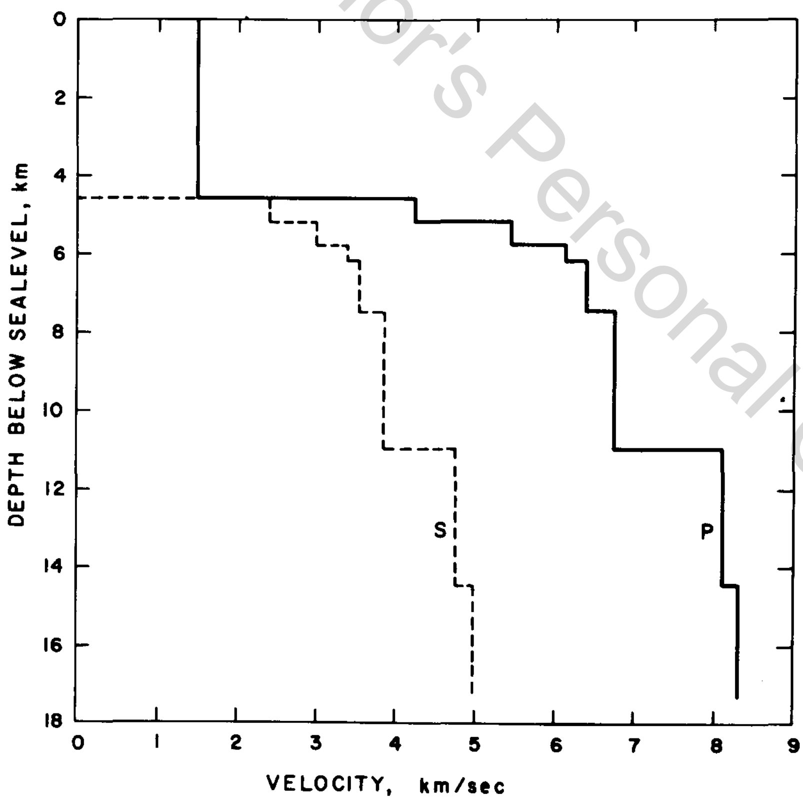

FIG. 4.—Crustal and upper-mantle velocities north of the Hawaiian Islands (after Helmberger and Morris 1969).

depth interval of layer 2. This is underlain by crustal rock with relatively uniform seismic velocities. A similar compressional-wave velocity distribution for the upper oceanic crust has been reported in the Norwegian Sea by Hinz and Moe (1971).

Although results from the Deep-Sea Drilling Project have shown that the upper few meters of layer 2 are basalt, the exact nature of the bulk of layer 2 is still uncertain. The large increase in compressional-wave velocity with depth in layer 2 is an important feature which can be explained by several petrologic models, including metamorphic recrystallization or intrusion of basalt by gabbro. It could be argued that this rapid increase in velocity results from decreasing serpentinization of oceanic crust underlying a thin layer of basalt. In figure 5, the compressional-wave velocities of Helmberger and Morris (1969) have been related to the percentage of serpentine, assuming that the entire oceanic crust is partially serpentinized peridotite. The data of Christensen (1966*a*) for partially serpentinized peridotite at appropriate pressures have been used for this model. Figure 5 illustrates that the oceanic crustal compressional-wave velocity distribution can be readily explained by assuming increasing serpentinization with depth.

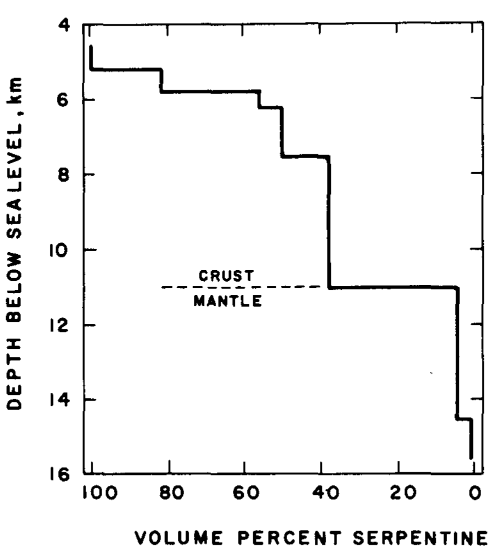

FIG. 5.—Percentage of serpentine in the oceanic crust (see text).

A check of this model is offered by the shear velocities given by Helmberger and Morris (1969). In figure 6, appropriate shear velocities (Christensen 1966*a*) for the degree of serpentinization given in figure 5 are compared with the crustal shear velocities. It is apparent that shear velocities for a serpentine-rich crust are appreciably lower than the velocities of Helmberger and Morris. It is concluded that this crustal velocity distribution is not consistent with a serpentine-rich oceanic crustal model.

Recently, it has been suggested that the oceanic crust is primarily basaltic in composition and the lower oceanic crust, in particular, is composed of gabbro and its metamorphosed equivalents. This model is consistent with laboratory studies of seismic velocities (e.g., Birch 1960; Christensen 1970*a*, 1970*b*), petrologic considerations (e.g., Cann 1968; Coleman 1971*b*), and comparisons of oceanic crustal structure with several ophiolite assemblages (e.g., Davies 1968; Coleman 1971*b*).

In this paper, the argument against an oceanic crust consisting of abundant serpentinite is based primarily on comparisons of ratios of seismic compressional- and shear-wave velocities with laboratory measurements in partially serpentinized peridotites. Laboratory measurements of the ratios of compressional- to shear-wave velocities in basalt, diabase, gabbro, and their metamorphic counterparts are, however, in agreement with the available seismic data for the oceanic crust. Poisson's ratios at 2 kbar for several possible oceanic rocks are summarized in table 6. It is clear that mafic rocks satisfy the seismic refraction data for the oceanic crust (table 5). Therefore, it seems probable that any correct petrologic model of the oceanic crust must consist of basalt overlying gabbro and metabasalt. This sequence contains subordinate serpentinized ultramafic rocks. The relative

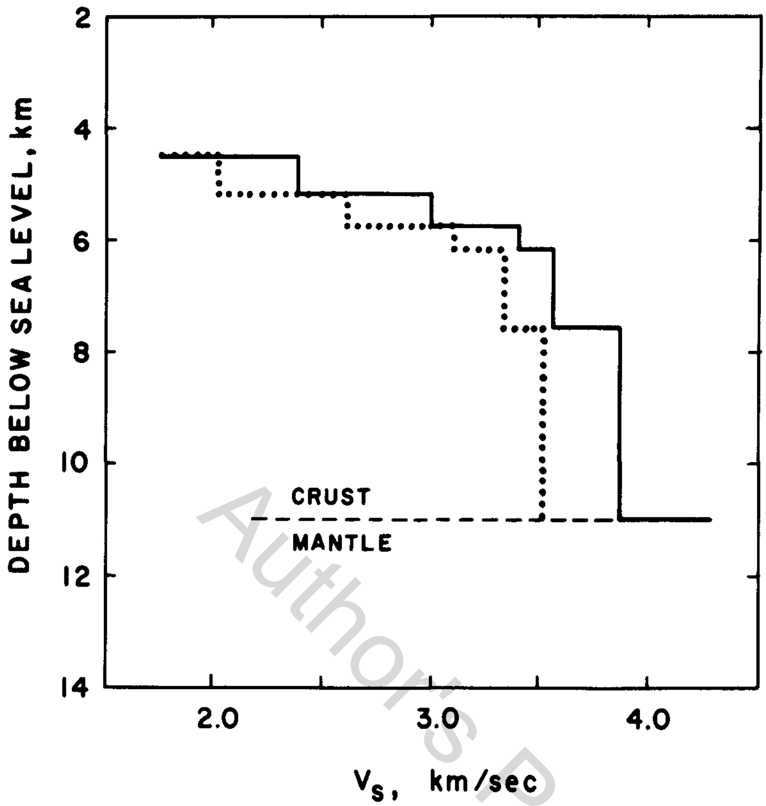

FIG. 6.—Calculated (*dotted lines*) and observed (*solid lines*) shear velocities in the oceanic crust

TABLE 6

POISSON'S RATIOS AT 2 KBAR

Rock	V_p/V_s	σ
Norite (Transvaal)*	1.84	0.29
Diabase (Virginia)*	1.82	0.28
Diabase (Maryland)*	1.79	0.27
Amphibolite (Montana)*	1.77	0.26
Amphibolite (Connecticut)†	1.71	0.24
Greenstone (Washington)‡	1.83	0.29
Gabbro (British Columbia)‡	1.79	0.27
Greenstone (California)‡	1.78	0.27
Greenstone (Virginia)‡	1.73	0.25
Gabbro (Indian Ocean)‡	1.84	0.29
Gabbro (Minnesota)‡	1.87	0.30
Dolerite (Mid-Atlantic)§	1.73	0.25
Basalt (Mid-Atlantic)§	1.79	0.27
Greenstone (Mid-Atlantic)§	1.88	0.30

* Birch 1960; Simmons 1964.
† Christensen 1965, 1966*b*.
‡ Christensen (unpublished data).
§ Christensen and Shaw 1970.

abundance of the various mafic rocks within the ocean crust is still uncertain. The variety of metamorphics and gabbros which have been reported from oceanic regions suggests that the detailed petrology of the oceanic crust may be rather complex. An investigation of the elasticity of mafic rocks from the ocean ridges is in progress; it is hoped that the results of this study may help in arriving at the distribution of mafic rocks within the oceanic crust.

The common occurrence of serpentinites along major fracture zones strongly suggests that their origin is related to tectonic processes operating at ridge crests. They very likely have originated from serpentinization of diapiric intrusions of mantle into the fracture zones. In addition, a limited number of rocks from dredge hauls suggests the existence of stratiform complexes in some oceanic regions (Engel and Fisher 1969; Melson and Thompson 1970). Thus, some oceanic serpentinites could have originated by hydration of peridotites formed as cumulates from basic magma within the crust or upper mantle.

ACKNOWLEDGMENTS.—G. Thompson generously provided the samples of serpentinite for the velocity measurements. The skill of M. Mulcahey, R. McConaghy, and K. V. Campbell in operating and maintaining the pressure system is gratefully acknowledged. The research was supported by National Science Foundation grant GA-20317.

REFERENCES CITED

ANDERSON, O. L.; SCHREIBER, E.; LIEBERMANN, R. C.; and SOGA, N., 1968, Some elastic constant data on minerals relevant to geophysics: Rev. Geophysics, v. 6, p. 491–524.

BAILEY, E. H.; BLAKE, M. C., JR.; and JONES, D. L., 1970, On-land Mesozoic oceanic crust, *in* California coast ranges: U.S. Geol. Survey Prof. Paper 700-C, p. C70–C81.

BARRETT, D. L., and AUMENTO, F., 1970, The Mid-Atlantic Ridge near 45° N. XI. Seismic velocity, density and layering of the crust: Canadian Jour. Earth Sci., v. 70, p. 1117–1124.

BIRCH, F., 1960, The velocity of compressional waves in rocks to 10 kilobars, 1: Jour. Geophys. Research, v. 65, p. 1083–1102.

——— 1961, The velocity of compressional waves in rocks to 10 kilobars, 2: Jour. Geophys. Research, v. 66, p. 2199–2224.

——— 1964, Velocity of compressional waves in serpentinite from Mayaguez, Puerto Rico, *in* BURK, C. A., ed., A study of serpentinite: Natl. Acad. Sci.-Nat. Research Council, Pub. 1188, p. 132–133.

BONATTI, E., 1968, Ultramafic rocks from the Mid-Atlantic Ridge: Nature, v. 219, p. 363–364.

———; HONNOREZ, J.; and FERRARA, G., 1970, Equatorial Mid-Atlantic Ridge: petrologic and Sr isotopic evidence for an alpine-type rock assemblage: Earth and Planetary Sci. Letters, v. 9, p. 247–256.

BOWIN, C. O.; NALWALK, A. J.; and HERSEY, J. B., 1966, Serpentinized peridotite from the north wall of the Puerto Rico Trench: Geol. Soc. America Bull., v. 77, p. 257–270.

BUNCE, E. T., and FAHLQUIST, D. A., 1962, Geophysical investigation of the Puerto Rico Trench and outer ridge: Jour. Geophys. Research, v. 67, p. 3955–3972.

CANN, J. R., 1968, Geological processes at mid-ocean ridge crests: Geophys. Jour. Royal Astron. Soc., v. 15, p. 331–341.

CHASE, R. L., and HERSEY, J. B., 1968, Geology of the north slope of the Puerto Rico Trench: Deep-Sea Research, v. 15, p. 297–317.

CHERNYSHEVA, V. I., and BEZRUKOV, P. L., 1966, Serpentinite from the crest of the Indo-Arabian Ridge: Doklady Akad. Nauk SSSR, v. 166, p. 207–210.

CHRISTENSEN, N. I., 1965, Compressional wave velocities in metamorphic rocks at pressures to 10 kilobars: Jour. Geophys. Research, v. 70, p. 6147–6164.

——— 1966*a*, Elasticity of ultrabasic rocks: Jour. Geophys. Research, v. 71, p. 5921–5931.

——— 1966*b*, Shear wave velocities in metamorphic rocks at pressures to 10 kilobars: Jour. Geophys. Research, v. 71, p. 3549–3556.

——— 1970*a*, Possible greenschist facies metamorphism of the oceanic crust: Geol. Soc. America Bull., v. 81, p. 905–908.

——— 1970*b*, Composition and evolution of the oceanic crust: Marine Geology, v. 8, p. 139–154.

———, and SHAW, G. H., 1970, Elasticity of mafic rocks from the Mid-Atlantic Ridge: Geophys. Jour., Royal Astron. Soc., v. 20, p. 271–284.

COLEMAN, R. G., 1971*a*, Plate tectonic emplacement of upper mantle peridotites along continental edges: Jour. Geophys. Research, v. 76, p. 1212–1222.

——— 1971*b*, Petrologic and geophysical nature of serpentinites: Geol. Soc. America Bull., v. 82, p. 897–918.

DAVIES, H. L., 1968, Papuan ultramafic belt: Internat. Geol. Cong., 23rd, Prague 1968, Rept. Proc. sec. 1, p. 209–220.

DEN, N.; LUDWIG, W. J.; MURAUCHI, S.; EWING, J. I.; HOTTA, H.; EDGAR, N. T.; YOSHII, T.; ASANUMA, T.; HAGIWARA, K.; SATO, T.; and ANDO, S., 1969, Seismic-refraction measurements in the northwest Pacific basin: Jour. Geophys. Research, v. 74, p. 1421–1434.

DIETZ, R. S., 1963, Alpine serpentines as oceanic rind fragments: Geol. Soc. America Bull., v. 74, p. 947–952.

ENGEL, C. G., and FISHER, R. L., 1969, Lherzolite, anorthosite, gabbro, and basalt dredged from the Mid-Indian Ocean Ridge: Science, v. 166, p. 1136–1141.

FRANCIS, T. G., and RAITT, R. W., 1967, Seismic refraction measurements in the southern Indian Ocean: Jour. Geophys. Research, v. 72, p. 3015–3041.

——— and SHOR, G. G., JR., 1966, Seismic refraction measurements in the northwest Indian Ocean: Jour. Geophys. Research, v. 71, p. 427–449.

HEKINIAN, R., 1968, Rocks from the mid-oceanic ridge in the Indian Ocean: Deep-Sea Research, v. 15, p. 195–213.

HELMBERGER, D. V., and MORRIS, G. B., 1969, A travel time and amplitude interpretation of a marine refraction profile: primary waves: Jour. Geophys. Research, v. 74, p. 483–494.

HESS, H. H., 1955, Serpentinites, orogeny, and epeirogeny: Geol. Soc. America Spec. Paper 62, p. 391–407.

——— 1962, History of ocean basins, *in* Petrologic studies: Buddington volume, Boulder, Colo., Geol. Soc. America, p. 599–620.

——— 1965, Mid-oceanic ridges and tectonics of the sea-floor, *in* WHITTARD, W. F., and BRADSHAW, R., eds., Submarine geology and geophysics: Colston Research Symposium, 17th, Bristol, England, London, Butterworths, p. 317–333.

HINZ, K., and MOE, A., 1971, Crustal structure in the Norwegian Sea: Nature, v. 232, p. 187–190.

HUGHES, D. S., and MAURETTE, 1956, Variation of elastic wave velocities in granites with pressure and temperature: Geophysics, v. 21, p. 277–284.

MELSON, W. G., and THOMPSON, G., 1970, Layered basic complex in oceanic crust, Romanche Fracture, equatorial Atlantic Ocean: Science, v. 168, p. 817–820.

———, ——— 1971, Petrology of a transform fault zone and adjacent ridge segments: Royal Soc. (London) Philos. Trans., ser. A, v. 268, p. 423–441.

MIYASHIRO, A.; SHIDO, F.; and EWING, M., 1969, Composition and origin of serpentinites from the Mid-Atlantic Ridge near 24° and 30° north latitude: Contr. Mineralogy and Petrology, v. 23, p. 117–127.

———; ———; ——— 1970, Petrologic models for the Mid-Atlantic Ridge: Deep-Sea Research, v. 17, p. 109–123.

MUIR, I. D., and TILLEY, C. E., 1966, Basalts from the northern part of the Mid-Atlantic Ridge: Jour. Petrology, v. 7, p. 193–211.

MURAUCHI, S.; DEN, N.; ASANO, S.; HOTTA, H.; YOSHII, T.; ASANUMA, T.; HAGIWARA, K.; ICHIKAWA, K.; SATO, T.; LUDWIG, W. J.; EWING, J. I.; EDGAR, N. T.; and HOUTZ, R. E., 1968, Crustal structure of the Philippine Sea: Jour. Geophys. Research, v. 73, p. 3143–3171.

NICHOLLS, G. D.; NALWALK, A. J.; and HAYS, E. E., 1964, The nature and composition of rock samples dredged from the Mid-Atlantic Ridge between 22°N and 52°N: Marine Geology, v. 1, p. 333–343.

OXBURGH, E. R., 1967, Mantle convection and the thermal requirements of various crustal phenomena: Geophys. Jour. Royal Astron. Soc., v. 14, p. 403–411.

PAGE, N. J., 1968, Chemical differences among the serpentine "polymorphs": Am. Mineralogist, v. 53, p. 201–215.

———, and COLEMAN, R. G., 1967, Serpentine-mineral analyses and physical properties, *in* Geological Survey research, 1967: U. S. Geol. Survey Prof. Paper 575-B, p. B103–B107.

PHILLIPS, J. D.; THOMPSON, G.; VON HERZEN, R. P.; and BOWEN, V. T., 1969, Mid-Atlantic Ridge near 43°N latitude: Jour. Geophys. Research, v. 74, p. 3069–3081.

QUON, S. H., and EHLERS, E. G., 1963, Rocks of the northern part of the Mid-Atlantic Ridge: Geol. Soc. America Bull., v. 74, p. 1–7.

RAITT, R. W., 1963, The crustal rocks, *in* HILL, M. N., ed., The sea: New York, Wiley, v. 3, p. 85–102.

RALEIGH, C. B., and PATERSON, M. S., 1965, Experimental deformation of serpentinite and its tectonic implications: Jour. Geophys. Research, v. 70, p. 3965–3985.

SIMMONS, G., 1964, The velocity of shear waves in rocks to 10 kilobars, 1: Jour. Geophys. Research, v. 69, p. 1123–1130.