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COMPOSITION AND EVOLUTION OF THE OCEANIC CRUST

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SUMMARY

Laboratory measurements of compressional wave velocities in rocks are consistent with an oceanic crust composed of an assemblage of hornblende and plagioclase (layer 3) overlain by tholeiitic basalt (layer 2). Partial melting of mantle peridotite under mid-oceanic ridges marks the initial stages in the development of the oceanic crust. Tholeiitic magma, which escapes to the ocean floor, forms layer 2. Layer 3, composed of amphibolite and hornblende gabbro, originates beneath the ocean floor by crystallization of tholeiitic magma under hydrous conditions and subsequent metamorphism in the vicinity of ridge crests. Once formed, the oceanic crust is transported laterally by horizontally spreading upper mantle. Disposal of the oceanic crust is accomplished by downward movement in the vicinity of descending limbs of convection cells. Partial melting of the oceanic crust in regions of downward convection results in the formation of the calc-alkaline suite of rocks and eclogite.

INTRODUCTION

As a result of a variety of new findings in studies of marine geology and geophysics the hypotheses of continental drift and sea-floor spreading have gained new impetus and many new adherents over the past decade. The mechanism by which continental drift is initiated and maintained is generally attributed to sea-floor spreading. In its most general form, sea-floor spreading invokes convection in the upper mantle. Upwelling currents along mid-oceanic ridges are envisaged as producing new oceanic crust which is carried away from the ridge crests by horizontally spreading currents. Even though detailed geophysical investigations have provided a wealth of data on the structure of these ridges and the oceanic crust, little is known, at present, about the mechanism which is responsible for the generation of oceanic crust. Of fundamental importance to an understanding of this mechanism is the composition of the oceanic crust.

This paper deals primarily with the problem of oceanic crustal composition. The seismic characteristics of the oceanic crust and mid-oceanic ridges are compared with laboratory investigations of the elastic properties of rocks and current hypo-

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thesis bearing on the composition of the oceanic crust. Composition of the oceanic crust is then examined in light of evolution of the crust by sea-floor spreading.

STRUCTURE OF THE OCEANIC CRUST

First arrival times obtained from seismic refraction studies have shown that thicknesses and velocities of layers in normal oceanic regions possess world wide similarities. Average oceanic crustal structures summarized by HILL (1957) and RAITT (1963) are listed in Table I.

TABLE I

AVERAGE OCEANIC CRUSTAL STRUCTURES (AFTER HILL, 1957, AND RAITT, 1963)

Layer	*Thickness (km)*		*Velocity (km/sec)*	
	HILL (1957)	RAITT (1963)	HILL (1957)	RAITT (1963)
1	0.45		2.00	
2	1.75	1.71 ± 0.75	5.00	5.07 ± 0.63
3	4.70	4.86 ± 1.42	6.71	6.69 ± 0.26

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Layer 1 has been sampled extensively and consists of unconsolidated or semi-consolidated sediments. Its thickness has been found to vary from zero on topographic highs to several kilometers in some oceanic deeps and continental rise areas (EWING and NAFE, 1963). Detailed seismic studies of layer 1 are complicated by poor acoustic properties of the sediments. Quite often appreciable velocity gradients are present in the upper portion of the sediments and velocities of the upper sediments are equal or lower than water velocity (NAFE and DRAKE, 1957; EWING and NAFE, 1963).

As with layer 1, many problems arise in obtaining velocities and thickness of layer 2 from seismic refraction measurements. This is primarily due to the short distance over which layer 2 can be observed as a first arrival (RAITT, 1963). Further errors result from variations in thickness of the overlying layer 1 and irregularities in the interface between layers 1 and 2.

The composition of layer 2 is more controversial than that of layer 1. Observed velocities of layer 2 are in the range of low pressure laboratory measurements of a variety of rocks, including most sedimentary rocks, volcanics, and granites. Apart from actual errors in measurement, the range of velocities reported for this layer can be attributed either to variability in composition or variations in porosity of rock with a relatively uniform composition. HAMILTON (1959) using data on sediment compaction interpreted layer 2 as being composed of well consolidated or lithified sediment. The interface between layers 1 and 2 could therefore result

from a decrease in porosity within a sedimentary column. EWING and EWING (1959), on the other hand, suggested that layer 2 is the upper part of layer 3. The boundary between these layers could possibly reflect a closure of pore spaces. Thickening of layer 2 near Pacific volcanic islands suggests an igneous origin (RAITT, 1957).

Recent investigations of magnetic properties of rocks and magnetic surveys of oceanic crustal areas support a basaltic composition for layer 2. Steep local magnetic gradients on the flanks of ridges where the sediment cover is thin or absent require that the source of magnetic anomalies in the oceanic crust originates for the most part in layer 2 (HEIRTZLER and LE PICHON, 1965; VINE, 1966). The iron oxide content of metamorphic and sedimentary rocks are usually too low to produce the observed anomalies, whereas basalt is generally rich in magnetic iron oxides. HESS (1964) also concluded, on the basis of low magnetic susceptibilities measured for serpentinites, that layer 2 is, at least in part, basalt.

Layer 3 constitutes over two thirds by volume of the oceanic crust and hence the composition of this layer is fundamental to an understanding of the processes responsible for the generation of oceanic crust. It is almost universally observed in oceanic seismic refraction measurements and is characterized by a relatively uniform compressional wave velocity and thickness (RAITT, 1963). This uniformity in thickness and compressional wave velocity suggests that layer 3 is relatively homogeneous. Composition of this layer will be discussed in detail in the following sections.

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LAYER 3 AS ULTRAMAFIC ROCK

HESS (1955, 1962, 1964, 1965) has advocated that layer 3 of the oceanic crust is composed of serpentinized peridotite. The reaction olivine + $H_2O \rightleftharpoons$ serpentine + heat was originally proposed as an explanation for epeirogenic movements of the sea floor and the Colorado Plateau, and later developed into an explanation of the Mohorovičić discontinuity. According to this theory, partially serpentinized peridotite is envisaged as being generated on the crests of oceanic ridges and spreading laterally to form layer 3 of the oceanic crust. Serpentinite is formed by hydration of peridotite along oceanic ridge crests above the 500° C isotherm.[1] The present position of the oceanic Mohorovičić discontinuity is determined by the level of the 500° C isotherm along the ridge crests. Thus, away from the oceanic ridges the discontinuity is fossilized at lower temperatures of 150°-200° C, and beneath the discontinuity there exists a thin shell where water and peridotite coexist metastably.

The presence of serpentinite in the oceanic crust is supported by dredge hauls

[1] Recent work by SCARFE and WYLLIE (1967) indicates that the upper stability limit of serpentine is about 430°C at 1 kbar.

which have recovered serpentinite near the Azores (SHAND, 1949), from the crest of the Mid-Atlantic Ridge between 29° and 31° N (QUON and EHLERS, 1963; NICHOLLS et al., 1964), from the Carlsberg Ridge (CANN and VINE, 1965), and the Vema fracture (VAN ANDEL and BOWIN, 1968).

LAYER 3 AS BASALTIC ROCK

Several investigators have interpreted the seismic velocities of layer 3 as evidence for basaltic or gabbroic composition (e.g., EWING and EWING, 1959; GUTENBERG, 1959; ENGEL et al., 1965). Although the exact connotation of gabbroic or basaltic is not clear in many papers, the terms will be used here to designate rocks chemically similar to basalt or gabbro. Thus, basaltic rock may be interpreted as basalt, gabbro, diabase, greenstone, amphibolite, or eclogite, whereas specific rock names (e.g., basalt, amphibolite, etc.) will denote texture and mineralogy as well as chemistry.

Recent support for a basaltic layer 3 is based primarily on the rejection of serpentinite as a major constituent of the oceanic crust. CANN (1968) has proposed a model of mid-ocean ridge structure in which basalt is produced by partial melting of the mantle at a depth of about 30 km beneath ridge crests; development of the crust is initiated by this generation of basalt. Lower levels of layers of extruded basalt are metamorphosed at ridge crests (in the presence of high thermal gradients) to rocks of the greenschist and amphibolite facies. Layer 2, composed of fresh basalts and greenschists, and an amphibolite layer 3 are carried away from the crest of the mid-ocean ridges by horizontally spreading upper mantle.

In support of a basaltic oceanic crust, CANN (1968) poses several objections to the concept of a partly serpentinized peridotite oceanic crust:

(*1*) The uniform velocities observed for layer 3 imply a remarkably uniform degree of serpentinization throughout the oceanic crust.

(*2*) The observed inverse relationship between the thicknesses of layers 2 and 3 and the uniform total crustal thickness over the Mid-Atlantic Ridge suggest that the two layers are of similar composition.

(*3*) The absence of layer 3 over the Mid-Atlantic Ridge and its presence over the East Pacific Rise suggest that, if layer 3 is composed of serpentinite, heat flow should be greater over the Mid-Atlantic Ridge (i.e., the 500° C isotherm should be within the crust). Observed heat flow appears to be higher over the East Pacific Rise.

OXBURGH (1967) and OXBURGH and TURCOTTE (1968), using a boundary-layer solution for steady cellular convection in the upper mantle, have predicted geotherm distribution for ascending convection currents under mid-oceanic ridges. Data for the melting of olivine tholeiite were used to define a zone of partial melting of mantle peridotite. Layer 3 is envisaged as originating from basalt which rises to the surface from the fusion zone. Once part of the oceanic crust, the basaltic

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rock is laterally transported in a passive role by the convection cells. Low-grade metamorphism of the basaltic layer near ridge crests is suggested as an explanation for the low N.R.M. of layer 3. OXBURGH (1967) and OXBURGH and TURCOTTE (1968) further conclude that a serpentinite crust is not compatible with their thermal model.

Support for a basaltic composition of the oceanic crust comes from dredgings of the oceanic ridges (e.g., QUON and EHLERS, 1963; NICHOLLS et al., 1964; ENGEL et al., 1965; CANN and VINE, 1965; MELSON and VAN ANDEL, 1966; MUIR and TILLEY, 1966; MELSON et al., 1968). The dredged mafic rock samples consist of basalt, greenstone, amphibolite, basaltic tuff, gabbro, and dolerite. Even though basalts with alkaline affinities have been reported from mid-oceanic ridge areas, tholeiites are by far the dominant oceanic basalt type (ENGEL et al., 1965; MELSON et al., 1968).

COMPOSITION OF LAYER 3 FROM MEASUREMENTS OF ELASTIC WAVE VELOCITIES IN ROCKS

A large amount of experimental data is presently available which can be applied directly to the interpretation of oceanic crustal composition. In particular, velocities of propagation of compressional waves in many rocks have been measured for pressures up to several kilobars. Average velocities in three mutually perpendicular directions for several pertinent rocks are listed in Table II. Data of

TABLE II

COMPRESSIONAL WAVE VELOCITIES (KM/SEC) IN SOME POSSIBLE OCEANIC CRUSTAL ROCKS

Rock	*Pressure (kbars)*				*Reference*
	0.5	1.0	2.0	4.0	
Chlorite schist	5.8	6.75	6.82	6.92	BIRCH (1960)
Epidote amphibolite	6.8	7.09	7.32	7.52	CHRISTENSEN (1965)
Amphibolite	6.0	6.63	6.87	7.04	CHRISTENSEN (1965)
Eclogite	7.5	7.69	7.81	7.89	BIRCH (1960)
Serpentinite	4.5	4.64	4.75	4.91	CHRISTENSEN (1966)
Tholeiitic basalt	5.9	6.02	6.08	6.16	CHRISTENSEN (1968)
Hornblende gabbro	6.7	6.74	6.78	6.80	HUGHES and MAURETTE (1957)

this type are greatly influenced by porosity at pressures below approximately 1 kbar. At low pressures openings between the mineral grains in igneous and metamorphic rocks lower the velocities, whereas at pressures of the order of 1 kbar solid contact between grains is established, and the recorded velocities increase at a lower rate with increasing pressure (BIRCH, 1961). Because of this influence of

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porosity on velocities, it is extremely tenuous to attempt to correlate elastic wave velocities of rocks measured at pressures below 1 kbar with seismic wave velocities.

The effect of temperature on compressional wave velocity is not as well understood as the influence of pressure; however, a few measurements are available for igneous and metamorphic rocks (HUGHES and JONES, 1950; HUGHES and CROSS, 1951; HUGHES and MAURETTE, 1956, 1957). BIRCH (1955) emphasized that many of these measurements must be used with caution, as heating at low pressures produces irreversible decreases in velocity due to loosening of grain structure. In view of this, only temperature measurements in rocks at several kilobars pressure are reliable.

The available data at elevated pressures for mafic and ultramafic rocks show a decrease in compressional wave velocity of less than 1% with an increase in temperature from 25° to 200° C. This is also supported by measurements on single crystals of gem quality and hot-pressed polycrystalline aggregates (e.g., SOGA and ANDERSON, 1966; SOGA, 1967). Since this decrease in velocity is within the error of measurement of compressional wave velocities for most rocks (BIRCH, 1960; CHRISTENSEN, 1966), the effect of temperature on velocity appears to be of minor significance in oceanic crustal areas with normal temperature gradients.

The effect of serpentinization on compressional wave velocities has been investigated by HESS (1959, 1962), BIRCH (1960, 1961, 1964) and CHRISTENSEN

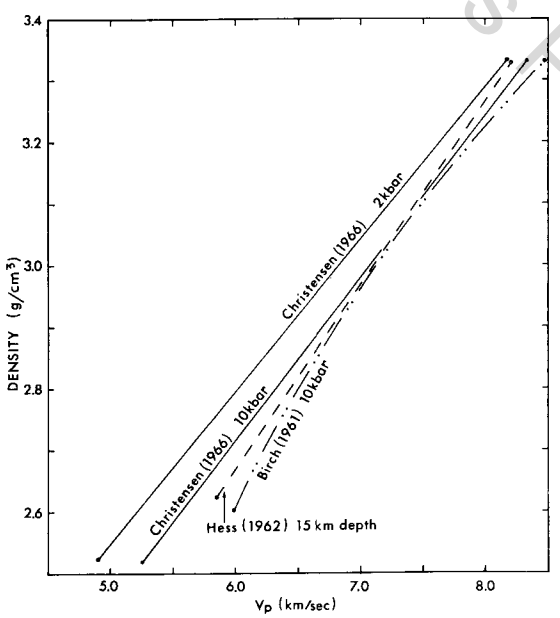

Fig.1. Compressional wave velocity–density relationships for olivine–serpentine and peridotite–serpentine aggregates.

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(1966). These studies have shown that the alteration of dunite or peridotite to serpentinite produces a large decrease in velocity (Fig.1).

HESS' (1962) relationship between compressional wave velocity and density, based on measurements by Birch at Dunbar Laboratory and J. Green of the California Research Laboratory, was estimated for a depth of 15 km; the degree of fit of the data to the curve was not given. BIRCH's (1961) curve represented calculated compressional wave velocities at 10 kbars for pure olivine-serpentine aggregates. Olivine was assumed to have a velocity of 8.5 km/sec and serpentine a velocity of 6.0 km/sec. Compressional and shear wave velocities have also been measured in a suite of rocks ranging from serpentinite and partially serpentinized peridotite to unaltered peridotite (CHRISTENSEN, 1966). The velocity-density relationships obtained from these measurements and given in Fig.1 for pressures of 2 and 10 kbars represent least squares solutions for 27 sets of data. For both curves the correlation coefficients were better than 0.99.

A wide range of velocities have been reported for serpentinite (CHRISTENSEN, 1966, fig.6). Many of the higher velocities can be attributed to the presence of olivine, pyroxene, chromite, magnetite, or brucite in the samples. BIRCH (1960, 1961) found that antigorite has a higher velocity than chrysotile. HESS and OTALORA (1964) postulated that sorbed water in pore spaces of serpentine was responsible for the low densities and velocities reported by BIRCH (1964). HUGGINS and SHELL (1965), on the other hand, have shown that the density of chrysotile increases with water content due to filling of the tubular structure of chrysotile. The relationships between these factors and velocities still remain to be disentangled.

It is clear, however, that the average compressional wave velocity of layer 3 of 6.7 km/sec cannot be attributed to completely serpentinized peridotite. HESS (1962) recognized this and postulated that layer 3 was composed of peridotite 70% serpentinized. The measurements at 2 kbars by CHRISTENSEN (1966) suggest that, if the oceanic crust is composed of a peridotite-serpentinite mix, a velocity of 6.7 km/sec corresponds to about 40% serpentinization (Fig.1). CANN's (1968) objection to a partially serpentinized crust on the basis of the remarkably uniform velocities observed for layer 3 remains a formidable argument against any hypothesis of the development of large volumes of oceanic crust by metamorphism of mantle material. Furthermore, because of their high degree of serpentinization and correspondingly low compressional wave velocities, serpentinites dredged from the ocean floor cannot be samples of layer 3.

A number of measurements of compressional wave velocities in basalts at high pressures have been recently made (CHRISTENSEN, 1968; MANGHNANI and WOOLLARD, 1968) which are useful in evaluating the possible basaltic composition of the oceanic crust. Compressional wave velocities recorded for basalts of tholeiitic composition are shown in Fig.2. Two features characteristic of basalts are apparent from this diagram: (*1*) compressional wave velocities and densities are extremely variable for tholeiites; and (*2*) lower density basalts tend to have lower velocities

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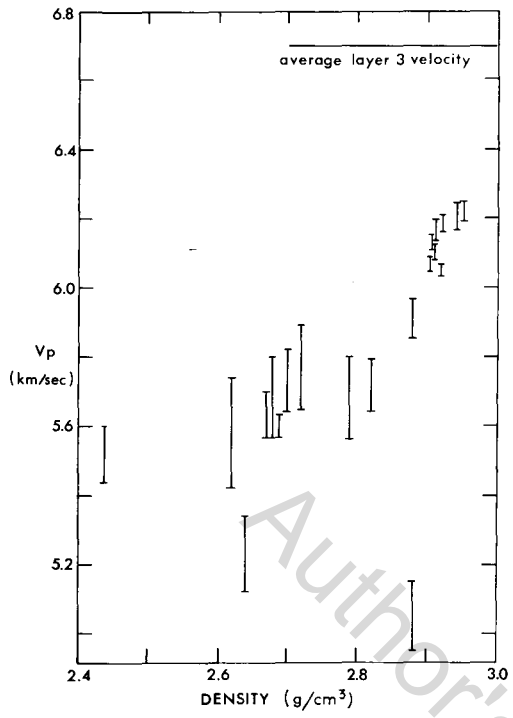

Fig.2. Compressional wave velocities at pressures of 1–2 kbars for tholeiitic basalts.

and also show a greater change in velocity between pressures of 1 and 2 kbars. The higher velocity and density basalts are free of vesicles and glass (CHRISTENSEN, 1968), whereas the basalts with lower velocities are vesicular with accompanying high porosities and may contain appreciable amounts of glass (MANGHNANI and WOOLLARD, 1968).

The average layer 3 velocity is also shown in Fig.2. It is clear from this diagram that layer 3 velocities are not consistent with laboratory measurements on tholeiitic basalt. Compressional wave velocities of layer 2, on the other hand, are within the range of laboratory measurements on vesicular basalt at pressures below 1 kbar.

At present, a limited amount of data is available on the elastic properties of metamorphosed basaltic rocks. BIRCH (1960) reported compressional wave velocities to 10 kbars for a chlorite schist, an amphibolite, and several eclogites. Average velocities of the chlorite schist are given in Table II. The modal analysis reported for the amphibolite (75% tremolite–actinolite–anthophyllite, 6% serpentine, 8% ore) is more consistent with a classification of hornblendite. The eclogite velocities (Table II) were considerably higher than observed velocities of layer 3, and hence eclogite can be eliminated as a major constituent of the oceanic crust. CHRISTENSEN (1965) investigated compressional wave velocities at pressures to

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10 kbars in a variety of metamorphic rocks including two amphibolites and two epidote-rich amphibolites. Representative velocities are given in Table II.

Clearly, further data on the elastic properties of metamorphosed basic rocks of known mineralogical and chemical composition are desirable before

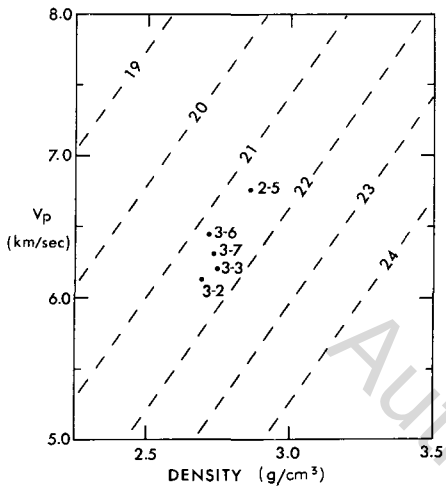

Fig.3. Compressional wave velocities at 10 kbars for greenstones collected from the Mid-Atlantic Ridge (calculated from chemical analyses and densities reported by MELSON and VAN ANDEL, 1966). Dashed lines are lines of constant mean atomic weight (BIRCH, 1961).

a firm conclusion can be reached regarding possible metamorphic rocks which satisfy the seismic velocities of the ocean crust. This is especially true for metamorphosed basaltic rocks of the greenschist facies. However, velocities of greenstones can be estimated using BIRCH's (1961) velocity–density–mean atomic weight relationship. This is shown in Fig.3 where densities and mean atomic weights calculated from the chemical analyses of MELSON and VAN ANDEL (1966) have been used to obtain compressional wave velocities for five greenstones collected from the Mid-Atlantic Ridge. The range in velocities (6.1–6.8 km/sec) obtained by this method are for pressures of 10 kbars. Velocities at 1–2 kbars would be approximately 5.7–6.5 km/sec and are lower than most seismic velocities reported for layer 3.

The laboratory measurements of compressional wave velocities in basalts and metamorphosed basic rocks and the calculated velocities for greenstones allow an estimate to be made of the changes in seismic velocity accompanying progressive metamorphism of basaltic rocks to the amphibolite facies (Fig.4). Because of limited velocity data, little can be inferred about the velocities of the various subfacies of regional metamorphism which have been discussed in detail by several metamorphic petrologists (e.g., FYFE et al., 1958; MIYASHIRO, 1961; HIETANEN, 1967). From the geophysical point of view, however, the facies classification of TURNER (1948) is sufficient for use in obtaining a velocity distribution

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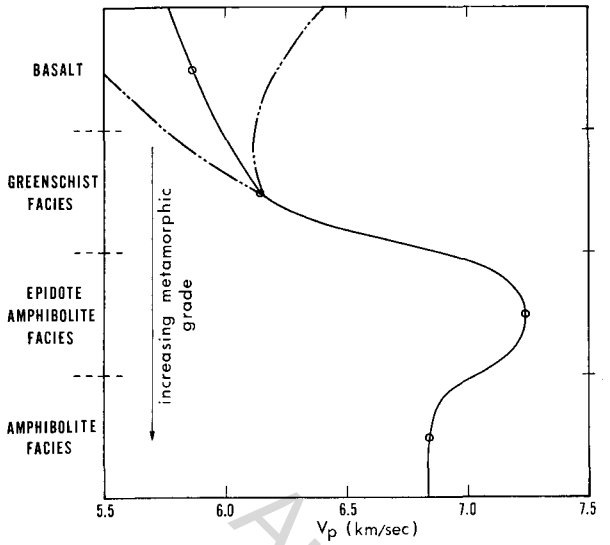

Fig.4. Average compressional wave velocities (circles) at 1.5 kbars for tholeiitic basalt and metamorphosed basic igneous rocks. Curved lines suggest approximate changes of velocity with progressive metamorphism.

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accompanying progressive metamorphism of basaltic rocks. This facies classification has been used in Fig.4; mineral assemblages of basic rocks stable within these facies are summarized in Table III.

TABLE III

MINERAL ASSEMBLAGES IN SOME METABASALTIC ROCKS (AFTER TURNER, 1948)

Facies	*mineralogy*
Greenschist	albite–actinolite–chlorite–epidote or albite–chlorite–epidote (calcite)
Epidote amphibolite	albite–hornblende–epidote (chlorite)
Amphibolite	plagioclase–hornblende (garnet) (epidote) (diopside)

The initial change in compressional wave velocity which results from metamorphism of basalt to rocks of the greenschist facies varies with the nature of the parent basalt (Fig.4). Metamorphism of basalt with abundant vesicules or glass will produce a rapid increase in compressional wave velocity. The formation of greenstone from relatively massive basalt will, on the other hand, result in a lowering of velocity. This is followed by a rapid increase of velocity to a maximum of about 7.3 km/sec for rocks of the epidote amphibolite facies. A lowering of velocity with further increase in metamorphic grade is primarily due to the formation of calcic plagioclase at the expense of epidote.

Thus, if the oceanic crust is composed of basalt which has undergone progressive metamorphism near ridge crests (where high thermal gradients prevail) to rocks of the amphibolite facies (CANN, 1968), we would expect to observe a seismic velocity profile similar to Fig.4. This velocity distribution cannot be reconciled with the seismic structure of the oceanic crust, since no regions of low velocity have been reported within layers 2 or 3. Furthermore, velocities as high as 7.3 km/sec are not common in oceanic crustal rocks.

This appears to leave two alternatives for the composition of layer 3: (*1*) rocks of this region are composed of metamorphosed basic rocks approaching a grade bordering the greenschist and epidote amphibolite facies; and (*2*) layer 3 is amphibolite. A lower oceanic crust of basaltic rocks containing mineral assemblages of greenschist or epidote amphibolite facies metamorphism would show a strong velocity gradient (Fig.4). An amphibolitic lower oceanic crust, on the other hand, is more probable because of the relatively simple mineral composition of amphibolite, which would produce relatively uniform velocities over the 4–5 km thickness of layer 3.

Amphibolites are composed mainly of hornblende and plagioclase. At pressures of 2 kbar aggregates of common hornblende have compressional wave velocities of about 6.8–7.0 km/sec and aggregates of plagioclase (andesine to labradorite) have velocites of 6.5–6.9 km/sec (CHRISTENSEN, 1965). These values are close to seismic velocities reported for layer 3. Furthermore, since hornblende and andesine–labradorite have similar velocities, the ratio of plagioclase to hornblende could vary through a fairly wide range in the lower oceanic crust and still produce the remarkably uniform seismic velocities observed for layer 3.

GENERATION OF THE OCEANIC CRUST

The discussion in previous sections suggests that the major features of oceanic crustal structure are most likely caused by two distinct petrologic assemblages, amphibolite and basalt. This conclusion, that the lower oceanic crust is composed of amphibolite overlain by a basalt layer 2, has important implications regarding the mechanism of crustal generation accompanying sea floor spreading.

The basalt of layer 2 most likely originates from sea floor eruptions along ridge crests and is carried away by horizontally spreading mantle. Rocks of the amphibolite facies are generally believed to have formed in a temperature range of about 550–750° C. Thus, it is necessary to generate layer 3 at depths along ridge crests where high thermal gradients prevail. The model of OXBURGH and TURCOTTE (1968) shows the probable region of formation of basalt magma. Magma solidfying above the zone of fusion is visualized as being metamorphosed to amphibolite and transported upward and away from ridge crests to form layer 3. This is supported by experimental work by YODER and TILLEY (1962) on the crystallization of a number of natural basalt compositions with $p_{H_2O} = p_{load}$. They

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concluded that, in the presence of water, all major basalts are "metamorphosed" to amphibolite even at relatively low temperatures. Furthermore, Yoder and Tilley found that a basaltic magma will crystallize directly as amphibolite in the presence of water. Recent experimental crystallization of basalt under hydrous conditions of $p_{H_2O} < p_{load}$ also demonstrates a large field of crystallization of amphibole (GREEN and RINGWOOD, 1968).

Thus, layer 3 may initially originate from direct crystallization of basaltic magma in the presence of water vapor and therefore it is not necessary to require metamorphism of basalt to produce an amphibolitic lower oceanic crust. Amphibolite formed in this manner should probably be called hornblende gabbro to indicate that it was precipitated directly from basaltic liquid (YODER and TILLEY, 1962). However, both the metamorphic amphibolite and its igneous counterpart will have similar mineralogy and seismic velocity. HUGHES and MAURETTE (1957) have reported a compressional wave velocity of 6.74 km/sec at 1 kbar for a sample of hornblende gabbro. This is almost identical to the average velocity of 6.7 km/sec of layer 3.

CANN and VINE (1965) have recovered three specimens of hornblende gabbro along the crest of the Carlsberg Ridge which may very well be samples of layer 3. Petrographic studies of thin sections from one specimen showed irregular banding, crushing the grains, and some alteration, which clearly indicate a complex history. Cann and Vine concluded that the assemblage hornblende–labradorite in these rocks could be a primary igneous assemblage or may have formed during relatively high-grade amphibolite facies metamorphism.

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DISPOSAL OF THE OCEANIC CRUST

One of the advantages of Hess' serpentinite crust is that it is disposable. HESS (1965) visualized that serpentinite crust is carried into the mantle along downward moving limbs of convection cells in the vicinity of oceanic trenches. The reaction peridotite + water → serpentine + heat is thus reversed producing mantle peridotite and water.

As most of the major trenches in the world border the Pacific, it is difficult to envisage how serpentinite oceanic crust from the Atlantic could extend under the continents to the Pacific border. Velocities in the lower continental crust are similar to the oceanic layer 3 (e.g., GUTENBERG, 1959) and therefore could correspond to partially serpentinized peridotite. However, temperatures in some portions of the lower continental crust are most likely above 500° C (BIRCH, 1955; CLARK, 1961, 1962); thus serpentine would not be stable in these regions. Amphibolite and hornblende gabbro, on the other hand, are most likely stable under the p–t conditions believed to exist within the lower crust. Several authors have, in fact, proposed that portions of the lower continental crust are composed of amphibolite (e.g., CHRISTENSEN, 1965; RINGWOOD and GREEN, 1966; JAMES et al., 1968).

We should expect to find evidence of partial melting of amphibolite along the Pacific margins where crustal rocks have been carried into the mantle by downward convection. The typical igneous rocks around the periphery of the Pacific Basin belong to what is commonly referred to as the calc-alkaline series. This suite of rocks is dominantly composed of high-alumina basalt, andesite, and dacite together with their plutonic equivalents. Andesite is the most abundant and its location, encircling most of the Pacific Basin, has been the subject of much speculation. Clearly, the origin of andesite and associated rocks of the calc-alkaline series is closely related to processes restricted to orogenic belts and island arcs.

Theories of origin of this suite have recently been reviewed by HAMILTON (1964), TAYLOR and WHITE (1965), and GREEN and RINGWOOD (1966). Briefly, explanations advanced for the origin of the calc-alkaline series include: (*1*) fractional crystallization of basaltic magma; (*2*) assimilation of crustal material in basaltic magma; (*3*) fractional melting of the upper mantle; and (*4*) partial melting of crustal rocks carried into the mantle by faulting or convection. Experimental data bearing on this problem recently have been reported by GREEN and RINGWOOD (1968). In this investigation the sequence of crystallization and compositions of the crystalline phases were determined for basalt, basaltic andesite, and andesite under pressures of 9–10 kbars and hydrous conditions. Their results demonstrated that the calc-alkaline series may originate from fractional crystallization of hydrous basalt at 30–40 km depth ($p_{H_2O} < p_{load}$), or by partial melting of amphibolite under similar p–t conditions. Using these results, Green and Ringwood suggested that the calc-alkaline suite may originate from a two-stage process. First, piles of saturated basalt develop near continental margins and are subsequently metamorphosed to amphibolite in lower regions. This is followed by partial melting of the amphibolite, which produces calc-alkaline magma and a residium which transforms to eclogite and sinks into the mantle.

With the amphibolite or hornblende gabbro crust proposed in this paper, the first stage in this process is unnecessary. The source of amphibolite which forms andesites and related rocks is the oceanic crust. Downward moving convection currents along the periphery of the Pacific are envisaged as dragging the oceanic crust downward. Disposal of the crust is accomplished by formation of the calc-alkaline suite of rocks and eclogite. It is interesting to speculate that the formation of the andesites and their related rocks has played a major factor in continental growth by accretion, similar to the process suggested by WILSON (1959).

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