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THE FINE STRUCTURE OF THE BASALT-ECLOGITE TRANSITION[1]

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ABSTRACT

Experiments show that the basalt-eclogite transition consists of two sharp density changes: from basalt and pyroxene granulite ($\rho=3.0$ g/cm^3) to garnet granulite ($\rho=3.2-3.25$) and from garnet granulite to eclogite ($\rho=3.4-3.5$). An intermediate layer ($V_p=7.4$–7.8 km/s) occurs beneath some areas which have undergone recent vertical movement. This layer is interpreted as garnet granulite. The top part of the earth's upper mantle is believed locally to be eclogite. Here the Mohorovicic discontinuity may be either a chemical change from acidic or intermediate rocks to eclogite, or a phase change from garnet granulite to eclogite.

INTRODUCTION

Whether the Mohorovicic discontinuity within the earth is a phase change from basalt to eclogite or a chemical change from silica-rich rocks, granite and basalt, to silica-poor rocks, peridotite, has been a topic of speculation for a long time. Eclogite, which is about 15 percent denser than basalt, is a high pressure form of basalt. The basalt-eclogite transition has been extensively investigated (Yoder and Tilley, 1962; Ringwood and Green, 1966; Green and Ringwood, 1967a; Cohen, Ito and Kennedy, 1967), but the equilibrium phase relations have not been adequately determined.

The hypothesis that the Moho discontinuity is a phase change is attractive as it offers an explanation for a number of geological observations (Kennedy, 1959). However, several objections to the hypothesis have been raised. They are:

(1) The density of eclogite is about 3.5, which is higher than the density estimated by some for the subcrustal material.

(2) The basalt-eclogite transition is not sharp, so the boundary is spread over a considerable range of depth. The Moho discontinuity, however, is generally sharp.

(3) There seems to be no correlation between heat-flow variations on the earth surface and the depth of the Moho discontinuity.

The most recent review can be seen in a paper by Joyner (1967).

The first objection is based on the widely accepted belief that the density of sub-Moho material is about 3.3. However, this value is not unique. As a matter of fact, it is assumed from the hypothesis that the subcrustal material is peridotite whose density is about 3.3. Recently Press (1969) computed the density of the suboceanic upper mantle. He found it to be between 3.5 and 3.6 and proposed that the material is eclogite.

The second objection is based on the experimental observations that there is a wide pressure-temperature transition field between the basalt and eclogite stability fields. As pressure increases through the transitional field, 16–23 kbar at 1200° according to Ringwood and Green (1966), garnet gradually increases and plagioclase gradually decreases. However, the presence of a mineralogical transition zone does not necessarily imply that the elastic properties also change gradually through the transition. We have measured the density change through the transition and found that the transition consists of two fairly sharp density discontinuities. The discontinuities are so sharp that seismic wave reflections might well be detected from the phase boundaries.

The third objection seems to be the most serious and apparently precludes the possibility that the world-wide Moho discontinuity is an equilibrium phase change. It, nevertheless, remains a possibility that in places the Moho discontinuity is a chemical change and that in other places the Moho discontinuity is a phase change. If so, the question is that of distinguishing between the two. Here we report our experimental results on the fine structure of the basalt-eclogite transition and its contribution to the interpretation of the structure of the crust-mantle boundary.

EXPERIMENTS

Two basaltic compositions were investigated: (1) Olivine tholeiite, NM5, the same rock investigated by Cohen *et al.* (1967); the composition of this rock is close to the average ocean tholeiite. (2) High-alumina olivine tholeiite made of 98 percent NM5+2 percent Al_2O_3. Olivine tholeiite is the most common rock type in the ocean area and high-alumina olivine tholeiite is the second major type found in the mid-Atlantic ridge (Aumento, 1967). Glasses were prepared by fusing rock powders in iron crucibles in a steel bomb.

Experiments were done in a piston-cylinder high pressure apparatus with iron capsules used as the sample containers. All the runs were made on a *P-T* line $T(°C)=10.4\,P\,(\text{kbar})+1070$, which is 20–50 degrees below the solidus curve of NM5 rock. Runs were held at temperature and pressure for about 3 hours at 1200°C and for about $1\frac{1}{2}$ hours at 1300°C.

The transformation reactions were reversed. The reaction from the low-density mineral assemblage to the high-density ones were investigated by using glass to which was added about 10 percent finely ground garnet-granulite powder (garnet+clinopyroxene +plagioclase), which had been crystallized from the same glass under pressure. Seeding of garnet-bearing crystalline powder was found to be necessary as garnet nucleates with difficulty. In the previous experiments on NM5 basalt glass [Cohen *et al.*, 1967], the glass was not seeded with garnet-bearing powder, as a result the pressure at which equilibrium garnet first appears was set approximately 5 kbar too high in the earlier report.

The reverse reactions from a high-density mineral assemblage

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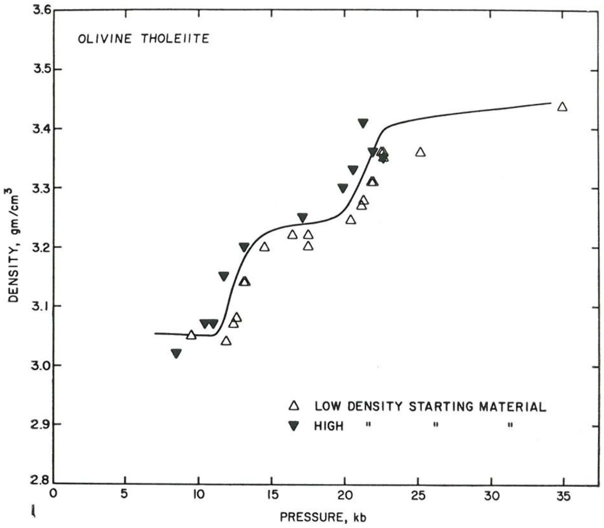

FIG. 1. Density-pressure diagram for an olivine tholeite composition.

to low-density ones were investigated by starting with either garnet-granulite powder or eclogite powder (garnet+clinopyroxine). These starting materials were prepared by crystallizing the glass at appropriate pressures and temperatures.

Temperatures were measured with platinum-platinum 10 percent rhodium thermocouples. Nominal pressures were calculated from the oil pressure and the ratio of cylinder area over ram area. Two kinds of runs, one on a compression cycle and one on a decompression cycle were made. Friction was assumed to be symmetrical.

A crystallized sample of about 30 mg was obtained from each run. The sample density was measured by weighing the sample in air and in toluene. A temperature correction for the density of toluene was made. The density of toluene was calibrated by measuring the apparent density of a quartz specimen.

A small amount of iron dissolved into the sample from the iron capsule and increased the density of the outer part of the sample. The density of a sample was first determined after the sample was taken from the capsule, cleaned in cold dilute HCl for about 10 minutes and then dried. After the first measurement, the sample was put in warm dilute HCl for 5–10 minutes. The leached surface of sample was sanded on carborundum paper and the sample cleaned, dried and weighed. The density of the second measurement generally decreased by 0.02–0.05 g/cm³ as a consequence of the loss of 1–3 mg of iron-enriched surface material. The procedure was repeated until the measurement became constant, usually after about 10 percent of the original sample was taken from the surface. The addition of iron from the capsule was limited to the surface of the sample. After the density was measured, the sample was crushed and the powder was examined both under the microscope and by X-ray diffraction. More detailed experimental results will be reported in a separate paper. This was not unlike a study on alkali basalts by Green [1967], although our results and conclusions differ.

At the lowest pressure, the "basalt" mineral assemblage, olivine+plagioclase+clinopyroxene, is stable. Near 10 kbar olivine disappears, by reaction with plagioclase to form pyroxene and the "pyroxene granulite" assemblage becomes stable. Garnet appears at slightly higher pressures and the "garnet granulite" assemblage, clinopyroxene +sodic plagioclase+garnet, is the stable assemblage at intermediate pressures. At higher pressures sodic plagioclase disappears and the "eclogite" assemblage, clinopyroxene +garnet, is stable. Here the albite component of the plagioclase reacts to form jadeitic pyroxene and the remaining aluminum in anorthite component of the feldspar is freed to form more garnet.

The density of the sample was plotted as a function of corrected pressure along the near-solidus line, $T = 10.4\ P$ (kbar)$+1070$°C, in Figures 1 and 2. The density change in the basalt-pyroxene granulite transformation is very small because of the small volume change associated with the reaction, olivine+plagioclase=aluminous pyroxene, and the small amount of olivine in the norm of our basalt NM5. The density of the basalt is about 3.00 and the density of the pyroxene granulite (50% clinopyroxene +50% plagioclase) is about 3.05. The density sharply increases over the pressure interval between 12 and 14 kbar where garnet appears. The amount of garnet increases and the amount of plagioclase decreases abruptly. The density of the garnet granulite, approximately made of equal parts of plagioclase, clinopyroxene and garnet, is between 3.20 and 3.25 and increases very slowly with increasing pressure. The density sharply increases again at pressures between 20 and 22 kbar where the amount of plagioclase decreases and that of garnet increases abruptly again. At 24 kbar, where plagioclase has disappeared, the density of eclogite (60% garnet and 40% clinopyroxene) is close to 3.4. It gradually increases to 3.45 at 35 kbar.

Thus olivine tholeiite undergoes two relatively sharp transformations: (1) The transition from pyroxene granulite to garnet granulite with a density change of about 0.17 at a pressure range between 12 and 14 kbar at about 1200°C. (2) The transition from garnet granulite to eclogite with a density change of about 0.15 at a pressure interval between 20 and 22 kbar at about 1300°C.

The high-alumina olivine tholeiite also undergoes two

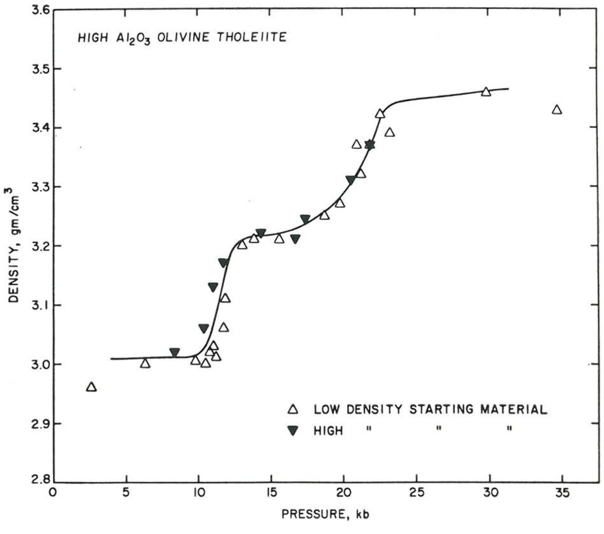

FIG. 2. Density-pressure diagram for a high-alumina olivine tholeiite composition.

sharp transformations with density changes of 0.20 and 0.18, respectively. The density jump is slightly larger in the high-alumina olivine tholeiite than in the olivne tholeiite owing to the added amount of garnet. In the two rocks, the first transition from pyroxene granulite to garnet granulite is quite sharp but the second transition from garnet granulite to eclogite seems to be less sharp.

Measured density changes are reasonably consistent with observed mineralogical changes, although the measured densities of garnet granulite and eclogite are slightly lower than the calculated density. Assuming that the densities of plagioclase, clinopyroxene, and garnet are 2.7, 3.3, and 3.8, respectively, the densities of pyroxene granulite, garnet granulite, and eclogite from NM5 rock should be 3.0, 3.3, and 3.6, respectively. The low measured densities for our garnet granulite and eclogite may result from internal cracks or it may be that the glass did not crystallize completely and some low-density intergranular material remained. Thus the absolute values of measured density are somewhat low, but the relative density changes remain valid.

The compressional wave velocity of eclogite has been measured under pressure by Birch (1960) and Kanamori and Mizutani (1965). Birch's eclogites contain relatively small amounts of garnet (12–26%) and the densities are low (3.33–3.44). They also contain significant amount of alteration products, consequently he obtained rather low velocities, between 7.62 and 7.94 km/s, at 6 kbar. These cannot be considered to be the representative values of the compressional wave velocity in eclogite. Birch kindly informed us of recent measurements of velocity on fresh eclogites of Norway. They have densities of 3.543 and 3.559, and compressional wave velocities of 8.21 and 8.35, respectively, at 10 kbar. Kanamori and Mizutani's eclogite consisted of 53 percent garnet, 42 percent clinopyroxene and 5 percent hornblende, which is close in mineral proportion to that of our eclogite formed from the olivine tholeiite. They obtained a compressional velocity between 8.30 and 8.54 km/s at 6 kbar. Although the amount of alteration products in their eclogite is small, the garnet crystals are rimmed by alteration products. This alteration rim could reduce the ultrasonic wave velocity significantly.

High compressional wave velocities in eclogite can be expected from the measured high velocities in garnet and jadeite pyroxene, the main constituents of eclogite. The compressional wave velocity in garnet ranges from 8.5 for Mn-Fe garnet to 8.6 for a Mg-Fe garnet to 8.7 for a Ca garnet (Verma, 1964; Ryzova *et al.*, 1966). The Mg-Fe garnet which is the dominant garnet of eclogite is expected to have a compressional wave velocity 8.6 km/sec at 1 atm. The compressional wave velocity of pure jadeite from Burma is 8.75 km/sec at 6 kbar (Birch, 1960). Other pyroxenes have lower wave velocities: 7.22 for augite at 1 atm. (Alexandrov *et al.*, 1964) and 7.8 for bronzite at 6 kbar (Birch, 1960). The pyroxene in our eclogite is estimated to contain 60 percent jadeite pyroxene. Therefore, its compressional wave velocity may be estimated to be approximately 8.2 km/sec. From these values, the compressional wave velocity of our eclogite synthesized from olivine tholeiite, containing 60 percent garnet+40 percent clinopyroxene, is estimated to be circa 8.4 km/sec and that of the eclogite from the high-alumina olivine tholeiite, is 8.5 km/sec.

The compressional wave velocity in garnet granulite has not been reported, but can be estimated from the wave velocity-density correlation. If eclogite with a density of 3.5 has a wave velocity of 8.5, garnet granulite with a density of 3.25 should have a velocity of around 7.8. Similarly, pyroxene granulite with a density of 3.0 should have a velocity around 7.0 km/sec. The effect of temperature in the mantle tends to lower the wave velocities of these rocks at the crust-mantle boundary, but, except in regions of high heat flow, this effect is largely or completely offset by the effect of pressure.

The estimated wave velocity for each rock depends strongly on the bulk chemical composition. High Al_2O_3 probably increases the wave velocity in garnet granulite and eclogite because such rocks contain a large amount of garnet. A low alkali garnet granulite would have a relatively high wave velocity because only a minor amount of feldspar would remain as plagioclase. If Na_2O is low and Al_2O_3 is high in an original basalt, the pyroxene granulite-garnet granulite discontinuity would be sharp but the garnet granulite-ecologite discontinuity might be obscure.

THE CRUST-MANTLE STRUCTURE

Double discontinuities at the crust-mantle boundary are not uncommon. In many regions, a layer which has an intermediate wave velocity between typical crustal material and mantle material has been found. Double discontinuities at the crust-mantle boundary are listed in Table 1. The intermediate layer is absent under the stable shields and deep ocean basins. It most prominently occurs under young tectonic zones which have undergone recent vertical movement. Although the double discontinuity has not been detected by seismic-refraction method under island arcs, the intermediate-velocity material ($Vp=7.4$–7.8 km/sec) is typical beneath the island arc crust such as Japan (Research Group for Explosion Seismology, 1966). Kanamori (1963), studying the crust-mantle structure in Japan on the basis of gravity and surface-wave data, proposed the presence of the intermediate layer with a P-wave velocity of 7.5 km/sec with the average thickness of 40 km overlying the normal mantle layer with $Vp=8.1$ km/sec. Also under elevated regions in the oceans such as midocean ridge (Ewing and Ewing, 1959; Le Pichon *et al.*, 1965), ocean rises (Ewing *et al.*, 1950; Raitt, 1956), and oceanic islands (Furumato *et al.*, 1968), the intermediate-velocity material is common beneath the typical crust. Though the lower boundary may be gradual and cannot be detected seismically, the thickness of the intermediate layer has been estimated from gravity data to be about 30 km under the mid-Atlantic ridge (Talwani *et al.*, 1965).

TABLE 1. THE DOUBLE DISCONTINUITY AT THE CRUST-MANTLE BOUNDARY

Location	Bottom crust		Intermediate layer	Top mantle		Reference
	V_p (km/sec)	Depth (km)	V_p	Depth	V_p	
North America						
Basin and Range	6.11	24	7.66	50	8.11	Press [1960]
	6.33	25	7.59	72	7.97	Berg *et al.* [1960]
Central Mexican Plateau	6.01	33	7.63	43	8.38	Meyer *et al.* [1961b]
Montana	6.97	34, 40	7.58	57, 50	8.07	Meyer *et al.* [1961a] Glover and Alexander [1969]
Mississippi Embayment	6.5	30	7.4	45	8.1	McCamy and Meyer [1966]
Newfoundland	6.70	23	7.52	41.5	8.69	Ewing *et al.* [1966]
Gulf of St. Lawrence	6.35	30	7.35	45	8.50	*ibid*
Asia						
Central Asia Depression	6.5	30–35	7.4	45–55	8.3	Godin *et al.* [1960]
	7.0	30–32	7.6	50	8.4	Beloussov *et al.* [1962]
	6.3	35	7.4	50	8.2	Volvovskiy and Volvovskiy [1962]
	6.6–6.7	27	7.5	43	8.5	Fomenko [1962]
Siberian Lowland	6.8–7.2	33	7.4–7.7	43	8.3	Pashutina *et al.* [1962]
Dnieper Graben	—	25	7.4	39	8.2	Sollogub [1969]
Europe						
Rhinegraben	6.5	21	7.9	50	8.1	Mueller *et al.* [1967]
	6.7	28	7.2–7.4	36	8.2	Meissner and Berckhemer [1967]
Central Germany	6.2–7.6	23	7.6	33	8.16	Fuchs and Landisman [1966]
S. W. Germany	6.6	28	7.2	34	8.2	Landisman and Mueller [1966]
Central Alps	6.9	48	7.4	65	8.2	Behnke [1967]
West Alps	6.0	10	7.4	50	8.15	Fuchs *et al.* [1963]
Western Mediterranean	6.6	11, 31	7.7	29, 58	8.17	Berry and Knopoff [1967] Payo [1969]
Oceans						
Shatsky Rise	6.85	13	7.3	17	8.4	Den *et al.* [1969]
	6.96	17	7.5	24	8.2	*ibid*
Darwin Rise	6.8	10.5	7.6	13.5	8.5	Maynard *et al.* [1969]

We propose that this intermediate-velocity material with a compressional velocity between 7.4 and 7.8 is garnet granulite with varying amounts of feldspar. It is interesting to note that, in general, intermediate layers are associated with areas which have undergone recent vertical tectonic movement, subsidence to form sedimentary basins or uplift to form rifts and rises. When the areas are rising, they are also associated with high heat flow such as in the Basin and Range, Rhinegraben and midocean ridges. This is possibly pertinent to our hypothesis, since the volume change accompanying the basalt-eclogite transition offers a reasonable mechanism for the sequence of vertical movements of the crust (Kennedy, 1959; Joyner, 1967).

It is also interesting to note that the lower boundary of the intermediate layer in many areas cannot be determined from a travel-time curve, but is estimated from the mean crustal-wave velocity, surface-wave dispersion and gravitational data. The lack of sharpness of the garnet granulite-eclogite transition as shown in our figures, may offer a explanation.

Drake and Nafe (1968), correlating the seismic refraction data with geologic structure, suggested that the material in the velocity range 7.2–7.7 km/sec may be of a transient nature and may appear and then disappear during the orogenic history of a given region. They pointed out that the intermediate-velocity material under midocean ridges disappears when the ridges spread horizontally. If so, this aspect of the intermediate-velocity material precludes the possibility that the material is a mixture of crust and mantle or that it is a rock with intermediate chemical composition like picrite.

The upper boundary of a subcontinental intermediate layer is more likely to be a chemical change from acidic or intermediate rocks to basic garnet granulite rather than to be the basalt-garnet granulite phase change, because the velocity of the overlying layer is generally less than that of basalt; the lower boundary may well be the garnet granulite-eclogite phase change. Thus the subcrustal material in many places may be eclogite.

In areas where the intermediate layer is absent, typically in stable shield areas, the Moho discontinuity may be a chemical change from acidic or intermediate rocks to eclogite. This hypothesis differs from existing chemical-change hypotheses, basalt-peridotite or granite-peridotite, in assuming eclogite to be the subcrustal material. The stability field of eclogite at low temperatures is not known, but it appears probable that eclogite may be stable at the Moho discontinuity beneath the shield areas where the heat flow is low and the temperature at the Moho at 35 km is around 400°C, and that garnet granulite is stable at the bottom of the crust beneath young continents where the heat flow is high and the temperature at the bottom of the crust, circa 50 km, is above 700°C.

In suggesting that the subcrustal material is eclogite, we do not mean that the upper mantle is of eclogite, but that eclogite is limited to the uppermost part of the mantle; below it is peridotite.

The upper boundary and the lower boundary of an intermediate zone where present in oceanic areas may well be phase changes. An intermediate zone is typical under midocean ridges. Le Pichon (1969) noted that the oceanic crust thickens from ridge crest to basins and suggested that this could be explained by a progressive hydration of the topmost part of the mantle. The thickening of the crust is owing to a thickening of layer 3 ($V_p=6.5$–7.0 km/sec) in spite of a thinning of layer 2 ($V_p=4$–6 km/sec). Cann (1968) however, suggested that layer 3 was developed from basaltic crustal material. The apparently contradicting suggestions of Le Pichon and Cann are brought in harmony if the oceanic crust is assumed to be a hydrous metamorphic phase of basalt, such as green-schist or glaucophane-lawsonite schist. Consequently, we suggest that layer 3 consists of schist, the intermediate layer is of garnet granulite, and the topmost part of the normal mantle is eclogite. As the ocean floor spreads away from ridge-crest and temperature decreases, the intermediate layer of garnet granulite disappears forming a crust of schist by progressive hydration near the surface and forming the normal mantle layer of eclogite by isochemical transition at greater depths. The phase boundaries may well be controlled by reaction rate and may not be equilibrium boundaries. Alternatively locally the sub-Moho oceanic material may be peridotite, though this is not consistent with Press' recent density models.

The disappearance of garnet-granulite phase at low temperatures may be expected from experimental data on synthetic related systems. The dP/dT of the reaction albite=jadeite+quartz is large (20 b/°C, Birch and LeComte, 1960) while that of the reaction forsterite+anorthite=aluminous pyroxenes+spinel is low (3.5 b/°C, Kushiro and Yoder, 1966). Therefore, the upper boundary and lower boundary of the granulite field may converge at a low temperature (Ito and Kennedy, 1968, Fig. 8).

Because the basalt-garnet granulite-eclogite transition is dependent strongly on bulk chemical composition of rocks and the composition of basaltic rocks in the continents and the oceans may be substantially different, the same phase boundary cannot be applied to the garnet granulite-eclogite transition boundary beneath the continents and beneath the oceans. Thus Bullard and Griggs' criticism (1961) that reasonable geotherms for the continents and the oceans cannot be constructed so that they intersect the same transition curve is not fatal to the basalt-eclogite hypothesis (Joyner, 1967).

The presence of an eclogite layer at the top of the peridotite mantle is gravitationally unstable as has been pointed out by Ringwood and Green (1966). This will activate the dynamic movement of the upper mantle. Press (1968, 1969) suggested that the basalts, formed by fractional melting of the peridotite mantle, rise under the oceanic ridges, spread bilaterally, cool and transform to eclogite. His idea is consistent with recent high-pressure experiments on melting of basalts and peridotite (Green and Ringwood, 1967b; Ito and Kennedy, 1967, 1968). Garnet and clinopyroxene, which are the densest fraction in peridotite, melt first to form a light liquid, at high pressures.

In this model, mantle convection is aided because the densest fraction of the mantle melts to form the lightest liquid in the rising cycle. The light liquid activates the rising current. Then the liquid freezes to the densest rock, eclogite, at the surface, the dense eclogite helps the effectiveness of the sinking cycle. A hypothetical sketch of the crust-mantle structure is given as Figure 3.

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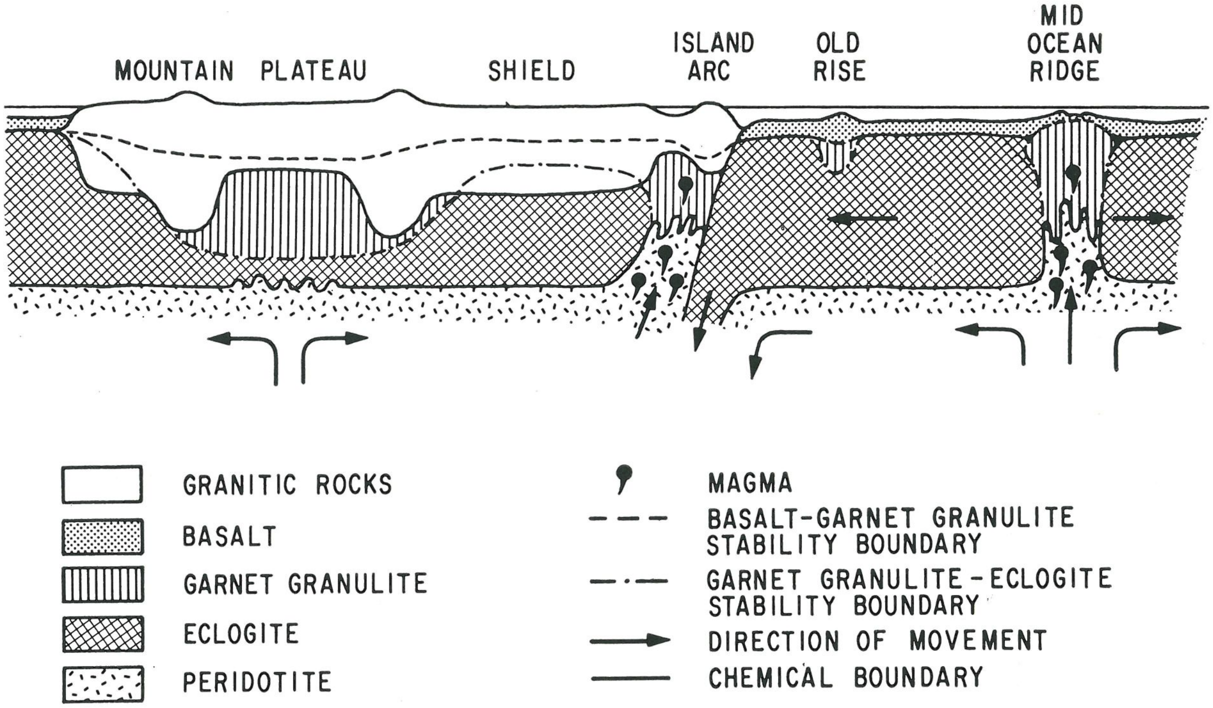

Fig. 3. Hypothetical crust-mantle structure.

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