

ties that satisfy the observed values in the upper mantle. As for dunite and peridotite, the ratio in volume of olivine to pyroxene of 8:2 to 9:1 may give the most probable value of velocity and density for the upper mantle.

2. *Poisson's ratio*. From the seismological observations, the Poisson's ratio of the upper mantle above the depth of 400 km should be about 0.275 and should not exceed 0.30 [see, Bullen, 1963, p. 233]. Under a pressure of about 10 kb the Poisson's ratio of eclogite is determined to be about 0.31. Only fresh dunite having more than 80% olivine has a Poisson's ratio of about 0.27, and the pressure dependence of this ratio is very small. Therefore, eclogite could not be the main constituent of the upper mantle.

Putting all this information together, we may suppose that the constituents of the upper mantle are olivine, pyroxene, and garnet and that the upper mantle material contains 80% (in mole, approximately the same as in volume) of olivine, 20% of pyroxene (mainly *enstatite*), and negligibly small amounts of garnet.

Even if the amount of the basaltic material, which will transform to 'eclogite' below a depth of about 30 km in the mantle, reaches one-fourth the composition of the whole upper mantle (the amount in the pyrolite model), it would not have any noticeable effect in the following calculation, because the elastic wave velocity of eclogite is nearly the same as that of peridotite and the differences in the Poisson's ratio and density between peridotite and eclogite are not so large.

Fe/Mg RATIO IN THE MANTLE

The density and seismic wave velocity of olivine and pyroxene are strongly controlled by the Fe/Mg ratio in the solid solution. The relation between velocity and the Fe/Mg ratio of minerals has not been established, however. On the other hand, compressibilities of most of the common rock-forming minerals have been determined in the wide range that covers the pressures of the lower mantle, and it is possible to estimate the density of the mixture of these minerals under the conditions in the mantle. Now we will try to determine the Fe/Mg ratio in the mantle materials from the viewpoint of density.

Fe/Mg ratio and density in the lower mantle.

The essential assumptions made in this density calculation are as follows: (1) The chemical composition does not change throughout the mantle. (2) Olivine and pyroxene in the upper mantle break down into oxides and there exists only a mixture of MgO, FeO, and stishovite or the most dense form of SiO₂. (Until recently it was widely believed that stishovite was the most dense form of SiO₂. *Ida et al.* [1967] have measured by X-ray experiment the compressibility of stishovite under high pressures and have discovered that the compressibility of stishovite shows an abnormal behavior similar to the behavior of MnO₂ and SnO₂, which crystallize in the same crystal structure as stishovite does (tetragonal). The most striking feature is that the *c* axis of these crystals actually increases with increasing pressure in the low-pressure region and has a maximum. The compressibility is very low at low pressure and then increases at pressures beyond the maximum in *c*. Another interesting fact is that in MnO₂ and SnO₂ an apparent first-order phase transition is observed in the higher-pressure range, though the high-pressure phase has not yet been investigated [*Clendenen and Drickemar*, 1966]. *Syono and Akimoto* [1968] investigated the polymorphic transition of PbO₂ (tetragonal at 1 atm) under high pressures and temperatures, and they found that PbO₂, which once crystallizes in an orthorhombic structure under high pressures, transforms to the cubic calcium fluoride CaF₂ structure in the higher-pressure region. They also found that the associated density increase through these transitions reaches about 8.2%. From these facts they suggested the possibility that the rutile-type of SiO₂, stishovite, transforms to the fluorite structure in the higher-pressure region.) (3) The Fe/Mg ratio is approximately the same in olivine and in pyroxene.

The densities of component minerals under high pressures were determined in the following ways:

1. *MgO and the most dense form of SiO₂*. From the many available data on the density at pressures, which have been deduced from the high-pressure X-ray measurements (MgO) and the shock compression experiments (MgO, SiO₂), we directly used the values for which the temperature effect had been considered; these are

the 4000°K isotherms obtained from the shock compression experiment of MgO and SiO₂ [Al'tshuler *et al.*, 1965]. Though the shock data of the matter in which the polymorphic transitions take place, such as SiO₂, are relatively less reliable, we had no alternative but to use this kind of results.

2. *FeO*. There exists only the static-pressure X-ray measurement, in which the maximum experimental pressure is 276 kb [Clen-den and Drickamer, 1966]. In this case a wide range of extrapolation is necessary, and the Murnaghan-Birch equation of state was used. The best fit is obtained between the experimental results and the equation of state when a value of 1.49 mb is chosen for K_0 (bulk modulus at 1 atm).

Though there is a possibility that a solid solution between MgO and FeO is formed, the density is calculated as a simple mixture of MgO, FeO, and SiO₂. The density distributions were calculated for the following four cases of the Fe/Mg ratio:

Case 1. Fe:Mg = 0:10.

Case 2. Fe:Mg = 1:9.

Case 3. Fe:Mg = 2:8.

Case 4. Fe:Mg = 3:7.

The calculated density distributions are shown in Table 1 and Figure 2, with the density distri-

bution of the Bullen model A-i [see Bullen, 1963] and the Birch models I and II [Birch, 1964].

In cases 1 and 4, the densities calculated are greatly different from the value given by Bullen and Birch for the lower mantle. If we adopt the density value of case 1 as the density of the lower mantle, yielded moment of inertia of core becomes

$$I_{\text{core}} = 0.449 M_{\text{core}} R_{\text{core}}^2$$

where I , M , and R are the moment of inertia, the mass, and the radius, respectively. This relation is of course impossible from stability considerations because the coefficient 0.449 exceeds the value 0.40 which applies to a sphere of constant density. In the case 4 the yielded moment of inertia of the core results in the relation

$$I_{\text{core}} = 0.354 M_{\text{core}} R_{\text{core}}^2$$

With this value we can estimate the ratio of the density at the core-mantle boundary to the density at the center of the core, if we suppose that density in the core increases linearly with the depth. From the simple consideration we obtain the relation

$$\begin{aligned} [(4/9) - 0.354] \rho_{\text{core boundary}} \\ = [(0.354/3) - 4/45] \rho_{\text{center}} \end{aligned}$$

TABLE 1. Density Distributions (grams per cubic centimeter)

| Depth, km | This Paper | | | | Bullen Model A-i | Birch Model | |
|--------------|---|--------|--------|--------|---------------------|-------------|------|
| | Case 1 | Case 2 | Case 3 | Case 4 | | I | II |
| | <i>Upper Mantle (olivine + pyroxene)</i> | | | | | | |
| 0 | | 3.34 | 3.46 | | | | |
| 33 | | | | | 3.32 | 3.43 | 3.32 |
| 100 | | 3.42 | 3.53 | | 3.38 | 3.43 | 3.32 |
| 200 | | 3.49 | 3.62 | | 3.47 | 3.45 | 3.35 |
| 300 | | 3.56 | 3.69 | | 3.55 | 3.53 | 3.44 |
| 400 | | 3.63 | 3.77 | | 3.63 | 3.68 | 3.62 |
| | <i>Lower Mantle (MgO + FeO + SiO₂)</i> | | | | | | |
| 0 | 3.90 | 4.04 | 4.19 | 4.33 | | | |
| 1000 | 4.34 | 4.52 | 4.69 | 4.87 | 4.65 | 4.52 | 4.58 |
| 1400 | 4.54 | 4.73 | 4.92 | 5.10 | 4.88 | 4.74 | 4.81 |
| 1800 | 4.75 | 4.94 | 5.14 | 5.32 | 5.10 | 4.95 | 5.02 |
| 2200 | 4.96 | 5.16 | 5.36 | 5.55 | 5.31 | 5.16 | 5.22 |
| 2600 | 5.13 | 5.34 | 5.55 | 5.75 | 5.51 | 5.36 | 5.42 |
| 2800 | 5.22 | 5.43 | 5.64 | 5.85 | | 5.46 | 5.52 |
| 2898 | | | | | 5.66 | | |