THE UPPER MANTLE TRANSITION REGION: ECLOGITE?

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Abstract. The upper mantle transition region is usually considered to be peridotite which undergoes a series of phase changes involving spinel and post-spinel assemblages. There are difficulties associated with attempts to explain the 220, 400 and 670 km discontinuities in terms of phase changes in a peridotitic mantle. Moreover, in a differentiated earth there should be large quantities of eclogite in the upper mantle. Eclogite is denser than Al_2O_3-poor mantle to depths of 670 km, but it stays in the garnet stability field to pressures in excess of those required to transform depleted mantle to denser phases such as ilmenite and perovskite. Eclogite, therefore, remains above 670 km. The seismic properties of the transition region are more consistent with eclogite than peridotite. Most of the mantle's inventory of incompatible trace elements may be in this layer, which is a potential source region for some basalt magmas. The radioactivity in this layer is the main source of mantle heat flow, 0.7 $\mu cal/cm^2$ sec, and drives upper mantle convection.

Introduction

There is abundant evidence that peridotite is important in the uppermost mantle. It is not clear that this situation can be generalized to the whole upper mantle. Depleted peridotite is less dense than any plausible parental material and it should concentrate immediately below the crust. There are several arguments that suggest that eclogite should be abundant somewhere in the upper mantle.

The earth is probably at least as well differentiated as the moon in which case it should have a basaltic crust more than 200 km in thickness. Estimates of the composition of the mantle [Ganapathy and Anders, 1974; Mason, 1966; Ringwood, 1975; Carter, 1970] are capable of yielding a basalt layer of the order of 400 km in thickness.

Whether the upper mantle is still enriched in a basaltic component depends on the relative densities of eclogite and Al_2O_3-poor mantle assemblages. This is controlled by the stability fields of garnet and such deep mantle phases as ilmenite and perovskite.

Chemical Stratification of the Mantle

There is now abundant evidence that even small bodies such as the moon can differentiate and produce basaltic magmas. It is likely that the earth was producing basaltic magmas while it was accreting. This would produce a chemically layered body with Al_2O_3, CaO, K_2O, Na_2O, U, Th, Ba, Sr, etc. concentrated toward the outside. If estimates of mantle composition are anywhere near correct the earth could have produced the equivalent of more than 400 km of basalt. The current crust represents only a small fraction of this inventory. Subduction continuously returns basalt to the mantle but its ultimate fate has not been adequately addressed. Basalt inverts to eclogite at shallow depths and apparently sinks rapidly through the upper mantle. Basic questions are how deep does it sink and does it get reassimilated into the mantle?

The densities of the phases of basalt and Al_2O_3-poor silicates (primitive residual lower mantle) are shown in Figure 1 (from data of Liu [1977], Akaogi and Akimoto [1977] and Ito and Matsui [1977]). The assemblages in the two systems are quite different. The Al_2O_3-rich assemblage provides the lightest (basalt) and heaviest (eclogite, garnetite) material at upper mantle (<670 km) pressures. Eclogite can sink to 600-700 km before it encounters assemblages in Al_2O_3-poor mantle which are denser. In the process of sinking it displaces residual lower mantle material upwards. The stratification of the mantle, whether achieved by early upper mantle-lower mantle separation or by subsequent subduction will therefore be basalt, peridotite, eclogite, garnetite overlying depleted, Al_2O_3-poor, lower mantle. This can give an average mantle composition identical to current estimates but the stratification is quite different.

The 220 km level is an important seismic [Lehmann, 1967; Hales et al., 1976] and tectonic discontinuity in the upper mantle [Anderson, 1979]. It is a good reflector of seismic energy which suggests that it is a compositional boundary. There is another sharp seismic discontinuity near 670 km [Anderson, 1967; Burdick and Helmberger, 1979; King and Calcagnile, 1976]. This appears to be a chemical as well as a phase boundary [Anderson, 1977; Gaffney and Anderson, 1973; Burdick and Anderson, 1975; Butler and Anderson, 1978]. We will refer to the region between 220 and 670 km as the transition region.

Figure 2 is a bulk modulus-density plot of various minerals and mineral assemblages. The triangles are extrapolated zero-pressure values for different depth intervals of the mantle. Region I (200+ km) is consistent with either peridotite (olivine + opx + garnet) or eclogite (garnet + cpx). In a peridotite mantle region II would be β-spinel + pyroxene ± garnet s.s. ± stishovite while in an eclogite mantle it would be garnet s.s. Both possibilities are consistent with the data. Below 500 km a peridotite mantle transforms to γ-spinel + garnet s.s. ± stishovite while an eclogite mantle remains garnet s.s. The bulk modulus is therefore consistent with either peridotite or eclogite between ∿200 and 670 km. Although this is a weak conclusion, it is significant that eclogite cannot be ruled out.

Estimates of the seismic velocities of various regions of the mantle extrapolated to surface conditions are given in Table 1 along with measurements on various rocks and minerals. The compressional velocity cannot distinguish between pyrolite and a garnet-rich eclogite. The shear

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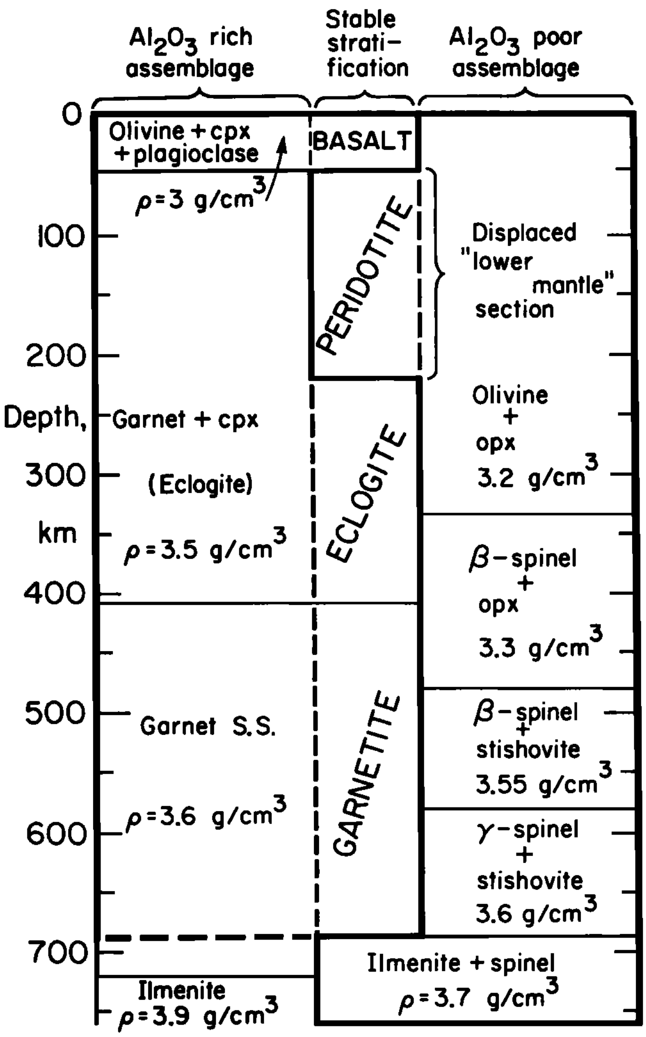

Figure 1. Zero pressure densities of basalt-eclogite (left) and peridotite (right) taken as a 1:1 mix of $MgSiO_3 \cdot Mg_2SiO_4$. The center column gives the gravitationally stable configuration. The broad stability field of garnet prevents eclogite from sinking below about 670 km.

velocity is much lower than olivine-rich assemblages but is well matched by eclogite. The V_P/V_S ratio is about 1.77, which is higher than olivine-rich assemblages. Garnet is the only major mineral that could increase this ratio. Some eclogites have properties similar to those required. These eclogites have densities of about 3.48 g/cm^3 compared to 3.38 g/cm^3 for garnet pyrolite. Eclogite would, therefore, displace garnet peridotite, at least down to 400 km. Figure 1 shows, in fact, that eclogite will sink to 670 km. It appears that eclogite is not only consistent with the seismic data but is superior when V_S and V_P/V_S are considered. From density considerations one can argue that eclogite, formed by subduction or by crystal fractionation, should settle into the transition region.

The velocity, density and V_p/V_S change across the 400 km discontinuity [Hart et al., 1977] are much smaller than expected for the olivine-spinel transition [Liebermann, 1973, 1975]. This would be the bottom of the cpx + ga to ga s.s. transition in the present model.

Trace Elements

We assume that the transition region has the composition of oceanic basalt [Engel et al., 1965] and have added this contribution to Gast's [1972] estimates of crustal abundances (Table 2). A recent estimate of whole earth abundances is also given. The crust and the eclogite layer account for nearly the total estimated incompatible trace element inventory of the earth. Incompatible trace elements in residual mantle (10% of basalt) yields a total mantle inventory 50% greater than computed above. The use of basaltic abundances rather than the average abundances in oceanic crust, may over-estimate the trace element content of the transition region.

The heat production of the crustal and eclogite layers corresponds to a heat flow of 1.2 $\mu cal/cm^2$ sec of which about 70% comes from between 220 and 670 km. The remainder of the mantle adds 0.5 $\mu cal/cm^2$ sec bringing the total heat flow in steady state to 1.7 $\mu cal/cm^2$ sec.

Upper mantle convection in this system is composed of two superposed interacting layers. The lower layer (eclogite) is internally heated and therefore is characterized by narrow descending plumes [McKenzie et al., 1974]. The upper layer (peridotite) is driven primarily by heating from below and is characterized by narrow ascending plumes. Subduction zones will appear to be continuous since cold material delivered from the surface to 220 km will control the locations of the cold descending plumes in the underlying layer. The thick conduction layer associated with shields may also affect convection in the eclogite layer.

Residual Mantle

In the perched eclogite model the primitive upper and lower mantles are identical and represent the residuum after ∿20% partial melting. The present

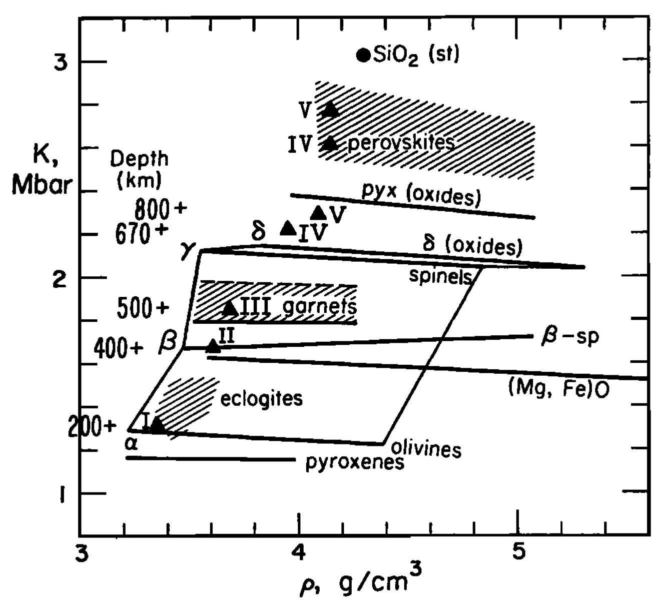

Figure 2. Bulk modulus vs. density for various minerals and various regions of the mantle [modified from Anderson, 1976, 1977]. Upper points for the lower mantle from Butler and Anderson [1978].

TABLE 1

Properties of Mantle Between 300 and 400 km Depth, Extrapolated to Surface Conditions

V_P	V_S	V_P/V_S	
			MANTLE
8.29	4.67	1.77	Anderson and Hart [1976]
8.19	4.69	1.75	Helmberger and Engen [1974]
8.36	4.73	1.77	Hart et al [1977]
			MINERALS (1)
8.48	4.93	1.72	olivine
7.84	4.74	1.65	orthopyroxene
8.06	4.77	1.69	diopside
9.16	5.20	1.76	garnet
9.00	5.00	1.80	garnet
			MINERAL ASSEMBLAGES (1)
8.37	4.87	1.72	pyrolite
8.30	4.86	1.71	peridotite
8.22	4.63	1.78	eclogite (23% ga)
8.31	4.66	1.78	eclogite (37% ga)
8.43	4.69	1.80	eclogite (51% ga)

(1) Green and Liebermann [1976]
Simmons and Wang [1971]
Manghnani et al [1974], Anderson [1977]

upper mantle may have been further processed and be slightly more depleted in SiO_2, Al_2O_3, CaO, etc. than the present lower mantle. To calculate the composition of residual mantle, we subtract the eclogite layer from primitive mantle [Ganapathy and Anders, 1974]. The residual mantle, which we assume occurs above 220 km and below 670 km, is remarkably similar to other estimates of upper mantle composition (Table 3).

Discussion

If the earth began to differentiate while it was still a small body the surface layer would be basaltic. As the upper mantle cools it enters the eclogite stability field and the surface layer becomes gravitationally unstable. This material is denser than "normal" mantle, residual upper mantle or primitive depleted lower mantle at moderate pressure. It therefore sinks into the lower mantle to a depth controlled by phase changes in eclogite and the lower mantle. Ilmenite and perovskite phase changes occur at lower pressure in Al_2O_3-poor material than in eclogite or primitive upper mantle. Eclogite can therefore sink to 670 km but no further. This becomes the boundary between the upper and lower mantle. Eclogite displaces primitive lower mantle upwards. The displaced portion of the lower mantle rides on top of the eclogite layer, which is presently at $\sim$220 km, and provides a source region quite distinct from the underlying eclogite.

Convection in the eclogite layer results in thermal boundary layers on both sides of the chemical discontinuity at 220 km. This high thermal gradient may be the cause of "kinked geotherms" found in some kimberlite inclusions [Boyd, 1973].

TABLE 2

Abundances of Some Trace Elements in Crust Plus Transition Region (220-670 km) and in the Whole Mantle (Expressed in ppm in Whole Mantle)

Element	(1)	(2)
Ba	2.5	8.5
U	0.04	0.03
Sr	28.2	30
Rb	0.55	0.97
K	305	283

(1) Crust (Model C, Gast, 1972) plus transition region (taken as oceanic tholeiite, Engle et al, 1965)

(2) Mantle [Ganapathy and Anders, 1974]

TABLE 3

Upper Mantle Composition

	(1)	(2)	(3)
SiO_2	46.8	44.5	46.1
Al_2O_3	2.3	2.6	4.3
FeO	7.8	8.6	8.2
MgO	40.9	41.7	37.6
CaO	2.0	2.3	3.1

(1) This study. Residual mantle after removal of eclogite layer (220-670 km) taken to have composition of oceanic tholeiite [Kay et al., 1970] from primitive mantle of Ganapathy and Anders [1974].

(2) Upper mantle [White, 1967].

(3) Pyrolite [Ringwood, 1975].

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