Compositional and density stratification in oceanic lithosphere— causes and consequences

E. R. OXBURGH & E. M. PARMENTIER

SUMMARY

Partial melting processes at mid-ocean ridges generate oceanic lithosphere which is chemically and mineralogically zoned. Basaltic oceanic crust overlies a 20–30 km thick zone of *depleted* upper mantle. This zone has undergone partial melting and is largely free of the high density phase, garnet, has a higher MgO/FeO ratio, and in consequence has a lower density ($\Delta\rho$ = 0·06 gm/cm^3), than the undepleted mantle of the lower lithosphere. The lithosphere does not become gravitationally unstable upon the asthenosphere for 40–50 Ma, when increased density resulting from thermal contraction has offset the compositional buoyancy of the depleted zone and crust. During subduction the basaltic crust inverts to eclogite and the net compositional buoyancy of the lithosphere is eliminated. However, as the subducted lithosphere is heated it becomes less rigid and density differences both between different parts of the descending lithosphere and the surrounding mantle become important. The dense eclogite layer sinks through the underlying depleted zone at a rate determined by the temperature-dependent rheology. With further heating the depleted zone becomes less dense than the overlying undepleted mantle and will diapirically rise some 300–400 km behind the trench depending upon the angle and rate of subduction and the age of the subducted plate. Such diapirs are able to intitiate behind-arc spreading. In a continental setting the diapirs could both heat the lithosphere and produce exceptional elevation.

1. Introduction

IT IS NOW generally accepted that oceanic lithosphere forms by the cooling of divergent horizontal mantle flows away from ocean ridges. Because this boundary layer is cold it has two important properties: (1) it is relatively strong and is therefore able to act as a stress guide and to exhibit the large scale mechanical properties now associated with tectonic plates; (2) it is slightly more dense than the warmer mantle material beneath and thus rests unstably upon it and is liable to sink.

In the development of thermal models for plates a simplified petrology has generally been assumed: an oceanic lithosphere comprising a crust which was predominantly basaltic, variously hydrated, veneered with oceanic sediment, and underlain by chemically homogenous ultramafic upper mantle (e.g. Forsyth & Press 1971).

In this paper we take a different view and show that the production of basalt beneath a mid-ocean ridge gives rise to a chemical and mineralogical layering within the lithosphere, and to density variations which are of similar size but of opposite sign, to those which result from cooling and thermal contraction.

We examine the nature and magnitude of these density variations and consider their implications first for the initiation of subduction, and then for the subsequent behaviour of the subducted plate.

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2. Density variations related to composition

It has been shown elsewhere (Oxburgh & Turcotte 1969) that it is a geometrical consequence of seafloor spreading, that the mantle material which directly underlies oceanic crust has undergone partial melting and basalt loss beneath the mid-ocean ridge at which the structure was formed. There is therefore a three-fold chemical zonation of the igneous part of the oceanic lithosphere: basaltic crust which overlies depleted mantle, which in turn overlies undepleted mantle. The mineralogical zonation of the lithosphere depends on both its chemical zonation and its thermal structure. Near the ridge where there are significant lateral thermal gradients the mineralogical zonation may be complex but in the region considered here, away from the crest, isotherms and thus the mineralogical zones are nearly horizontal.

Several authors (O'Hara 1975, Oxburgh & Turcotte 1976) have pointed out that the ultramafic residues formed by mantle partial melting during the generation of basalt must be *less dense* than the ultramafic starting material. This results partly from the loss of the dense phase garnet during melting and partly from an increase in the MgO/FeO ratio of the residue. In order to evaluate this density difference we assume that the undepleted upper mantle has the composition of garnet lherzolite comprising the four phases olivine, orthopyroxene, clinopyroxene, garnet in the proportions 57:17:12:14 (Ringwood 1969, Wyllie 1971); this composition is certainly open to question, but we use it simply as a basis for calculation and our conclusions do not depend on it in detail. Table 1 shows the zero-pressure density of such a mantle and its dependence upon the densities of the constituent minerals; mineral compositions and densities have been assigned on the basis of analysed coexisting upper-mantle mineral assemblages (Boyd & Finger 1975).

It is from material with the composition of this undepleted zone that both the oceanic crust and depleted zone must be formed. The igneous part of the oceanic crust is about 8 km thick (Sutton *et al.* 1971) and represents a degree of partial

TABLE 1: *Densities of depleted and undepleted model mantle and basalt of the oceanic crust. Note that density values at the precision given are significant only for the calculation of density differences.*

UNDEPLETED MANTLE ρ = 3.36 gm/cm^3

	%	ρ gm/cm^3
Olivine (Fo_{88})	57	3.32
Orthopyroxene (En_{88})	17	3.30
Clinopyroxene (Diopside)	12	3.25
Garnet (Pyrope 70, Almandine 12, Grossularite + Andradite 13; other components 5)	14	3.67
	100	

PARTIAL MELTING →

BASALT ρ = 3.0 gm/cm^3
(Inverts to Eclogite during subduction ρ = 3.60 gm/cm)

DEPLETED MANTLE ρ = 3.295 gm/cm^3
(Dunite/Harzburgite)

melting of between 20 and 30 per cent (Green 1971, Wyllie 1971); we shall assume a 25 per cent melt, and thus a 21 km thick depleted zone; this is taken to have a uniform degree of depletion and a sharp interface with the undepleted zone beneath. These assumptions considerably simplify the analysis and we shall show later that if, as seems more reasonable, there is a gradual transition between the two zones, or if the depleted zone is thicker with a lower degree of partial melting, our conclusion is not seriously modified.

We now consider the density of the depleted zone. Partial melting of undepleted zone material ($\rho = 3{\cdot}36$ gm/cm^3) to yield basalt should result in the loss of all garnet, most or all of the clinopyroxene and a little orthopyroxene and olivine (O'Hara 1968, 1970, Ringwood *et al.* 1964, Green 1971, Wyllie 1971). Garnet (3·67 gm/cm^3) is significantly denser than any other phase and its loss together with that of clinopyroxene would reduce the density of the residue to 3·315 gm/cm^3. This value is, however, further reduced by the fractionation of iron into the melt. Virtually all models for the chemical composition of the mantle have MgO/FeO values >5 while that for most basalts is <1. This is largely because during melting olivine and orthopyroxene react with the melt to enrich it in iron and to become more Mg-rich themselves. For our model mantle (MgO/FeO $\simeq 5{\cdot}7$) to yield a 25 per cent melt with an MgO/FeO ratio of 1, the MgO/FeO ratio of the residue must rise to 7·24. Insofar as the residue is likely to be almost entirely olivine and orthopyroxene this may be expressed as an increase in the Fo and En contents of these phases by 1·5 per cent; it is not important from the point of view of density whether the increase is of the same amount in both phases, because in the compositional range of interest (Fo_{89}, En_{89}) their densities are about the same, as are the dependencies of their densities upon Fe content (Fig. 1). This change in the MgO/FeO ratio reduces the density of the depleted mantle by a further 0·02 gm/cm^3, about half the size of the effect associated with the loss of garnet, but still significant.

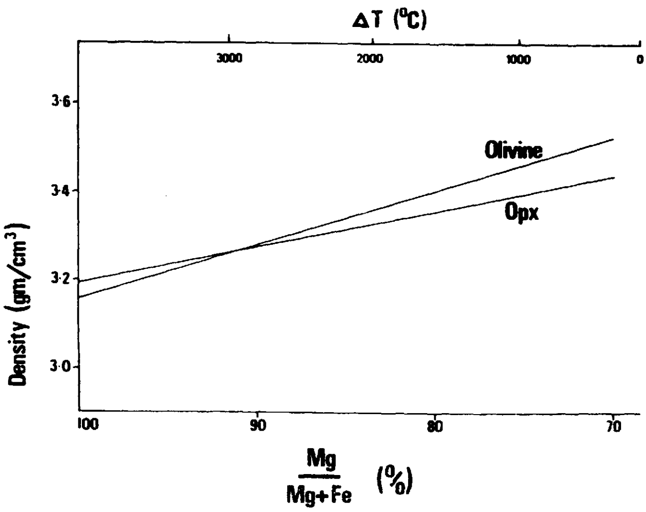

FIG. 1.

The density of olivine and orthopyroxene. The lower scale shows the variation of density with composition for olivine and orthopyroxene at room temperature. The upper scale shows how the density of olivine changes for a given change in temperature. It is evident that a change of 8 per cent in composition produces a density change equivalent to a temperature difference of 1000°C.

Experimental studies on appropriate bulk compositions have shown that one mineral assemblage (olivine-orthopyroxene) is stable throughout the depth of the assumed depleted zone (8 to 29 km). If the high temperature region of the ridge crest is ignored, two mineral assemblages are expected in the undepleted zone (MacGregor 1964): (1) spinel-aluminous clinopyroxene-aluminous orthopyroxene-olivine and (2) garnet-clinopyroxene-orthopyroxene-olivine. The second assemblage underlies the first and is about 0·015 gm/cm^3 more dense.

Our proposed lithospheric structure is summarized in Table 2. It may be compared with that of Ringwood (1969); the two are generally similar except in two respects—we have adopted sharp boundaries between our layers for the reasons given above, and we have taken into account the depletion of the uppermost mantle. Such depletion is incompatible with the presence of significant amounts of water (which should have been strongly fractionated into any melt) and in consequence we do not include amphibole in our depleted zone assemblage.

It is evident that the density *differences* between the layers of the oceanic lithosphere shown in Table 2 are not sensitive to the specific model on which we have based our calculation, although the *absolute* values of density would be. Almost any ultramafic mantle model must contain garnet which is mostly or entirely lost in basalt production; equally there is no alternative to an Mg-enriched mantle residue. If the depleted zone extended deeper, reflecting a lower percentage mantle melt and/or a broad gradation into the undepleted zone, the same net density deficiency would be present in the uppermost mantle; it would simply be expressed as a slightly lower density difference affecting a greater depth interval. We have made no mention of the thin sedimentary capping to the lithosphere. These sediments probably have no general or effective mechanical coupling to the 'igneous lithosphere' and are omitted from our model.

TABLE 2: *Model oceanic lithosphere*

Layer	Layer thickness	Rock type	Density*	
CRUST**	8 km	BASALT	~3.0	
				M
DEPLETED ZONE	21 km	DUNITE/HARZBURGITE	~3.295	
UNDEPLETED ZONE	20 km	SPINEL LHERZOLITE	3.345	
		GARNET LHERZOLITE	3.36	

*These densities are model densities based on the garnet lherzolite value of 3.36 gm/cm^3 ; the second and third decimal places are significant only for the derivation of density differences.

**The sedimentary component of the crust is omitted.

3. The buoyancy of a compositionally stratified lithosphere

It is evident from the preceding section that the crust and depleted zone should be less dense than the undepleted mantle upon which they rest. We now superimpose the effects of cooling and lithosphere formation upon this compositional structure and show that *young* lithosphere is buoyant until significant cooling has occurred. Lithosphere is a thermal boundary layer which thickens with age or distance from ridge crest by conductive cooling (Turcotte & Oxburgh 1972). A measure of the buoyancy of the lithosphere is given by the density defect thickness, δ, which can be represented in the form of an integral over the density ρ, which varies with depth within the lithosphere:

$$\delta = \int \left(\frac{\rho_m - \rho}{\rho_m} \right) dz \tag{1}$$

where ρ_m is the density of undepleted mantle rock at the temperature of the mantle beneath the plate and z is the depth coordinate. The integration is taken over the thickness of the lithosphere. A positive value of δ implies that the lithosphere has an average density less than the underlying mantle and that the lithosphere is therefore buoyant.

The density variation can be considered as the superposition of compositional and thermal variations. Similarly, δ can be expressed as the sum of thermal and compositional contributions. For the compositional density variation, the contribution to δ is evaluated as the product of $(\rho_m - \rho)/\rho_m$ and layer thickness summed over all of the compositional layers, in this case the basaltic crust and the depleted layer. The necessary data are given in Table 2.

The thermal density variation can be expressed as:

$$\left(\frac{\rho_m - \rho}{\rho_m} \right)_{\text{thermal}} = \alpha\,(T - T_m) \tag{2}$$

where T_m is the mantle temperature beneath the plate and α is a mean coefficient of thermal expansion for the upper mantle. Therefore, knowing the temperature distribution in the thermal boundary layer, the thermal contribution to δ can be determined.

Unfortunately, the temperature distribution within the lithosphere is not well known. A variety of temperature distributions have been proposed. The simplest and physically most reasonable model is a thermal boundary layer which develops in a mantle of uniform temperature, T_m, moving away from the mid-ocean ridge with a uniform horizontal velocity, U. This model predicts a temperature distribution of the form:

$$T = T_m \operatorname{erf} \left(z \sqrt{\frac{U}{4\kappa x}} \right) \tag{3}$$

where κ is the thermal diffusivity of mantle rock and x is a coordinate measured along the plate away from the accreting plate boundary. However, this temperature distribution predicts heat flows in oceanic plates older than 80 Ma that fall below observed values (Sclater *et al.* 1975, Richardson 1975).

It is preferable to evaluate the thermal contribution to δ directly from observations. This can be done in several ways, most directly from heat-flow measurements. Since cooling of the lithosphere is by heat conduction along a vertical temperature gradient, it is simple to show by relating the heat loss to the temperature distribution within the thermal boundary layer, that:

$$\delta_{\text{thermal}} = -\frac{\alpha}{\rho C p U}\int_0^x q dx \tag{4}$$

where Cp is the specific heat and q is the measured heat flow on the ocean floor. Heat-flow is, however, difficult to measure and may be influenced by a variety of factors which are not yet well understood (cf. Williams *et al.* 1974).

Alternatively, with a few additional assumptions, the thermal contribution to δ can be evaluated empirically from the topography of the ocean floor. The topography has been shown to be a nearly universal function of age and so has been attributed to cooling and contraction of the oceanic lithosphere (Sclater *et al.* 1971). This is based on the assumptions that the topography is due primarily to density variations within the lithosphere which are isostatically compensated, and that pressure gradients and viscous stresses due to flow in the more fluid mantle beneath the lithosphere are small. These assumptions are supported by the observation that topography expressed as a function of age of lithosphere is nearly independent of spreading rate and that the free-air gravity anomalies associated with spreading ridges are small. Therefore the total mass within any water-crust-mantle column must be independent of distance from the ridge. This is expressed mathematically as:

$$\frac{d}{dx}\left(\delta + \frac{\rho_m - \rho_w}{\rho_m} d\right) = 0 \tag{5}$$

where d is the water depth and ρ_w is the density of water. We note that the compositional contribution to δ is built into the lithosphere near the ridge axis, within a few hundred kilometres, and so is not dependent on x. Then, taking $\delta_{\text{thermal}} \simeq 0$ on the ridge axis where the water depth is d_o, equation (5) gives:

$$\delta_{\text{thermal}} = -\left(\frac{\rho_m - \rho_w}{\rho_m}\right)(d - d_o) \tag{6}$$

It is difficult to assign a precise value to d_o because the topography of ridges is often most complicated near the ridge axis, particularly for slowly spreading ridges; also the width of the zone over which the compositional structure is established cannot be determined from observations. However, this introduces an uncertainty in δ_{thermal} probably no larger than a few tens of metres.

The net density defect thickness of the lithosphere, δ, the sum of thermal and compositional contributions, is plotted as a function of age of lithosphere, τ, in

Fig. 2. The compositional contribution is calculated from the data of Table 2, and the thermal contribution is calculated in several different ways for the purpose of comparison. Noting that for a constant spreading rate, $\tau = x/U$, the values shown by the triangular symbols are determined by calculating δ_{thermal} from equation (6) using the empirical depth/age relationship of Sclater *et al.* (1971). The solid curves are calculated using a value for δ_{thermal} determined by substituting the temperature distribution given by equation (3) into equation (2), integrating equation (1) and using this result to give:

$$\delta_{\text{thermal}} = \frac{-2}{\sqrt{\pi}}\, \alpha\, T_m \sqrt{\kappa\tau} \tag{7}$$

Curves are plotted in Fig. 2 for $\alpha = 3 \times 10^{-5}/°\text{C}$, $\kappa = 10^{-2}\ \text{cm}^2/\text{sec}$, and $T_m = 1000°\text{C}$ and $1200°$.

The slope $d\delta/d\tau$ which can be determined from equation (4) is proportional to the heat flow q. The variation of δ shown for all the curves in Fig. 2 is consistent with heat-flow values slightly greater than 1 HFU for ages of 70 to 80 Ma. Measured heat flow in this age range is about 1·2 HFU.

As can be seen from Fig. 2, lithosphere remains buoyant on the underlying mantle (i.e. $\delta > 0$) until it is more than 40 Ma old. For older lithosphere, the thermal contribution to δ is larger than the compositional contribution making δ negative. Lithosphere is then denser than the underlying mantle and subduction should occur if this unstable state is sufficiently disturbed.

We emphasize that there is no *net* compositional difference between the lithosphere as a whole, and undepleted mantle; the material extracted from the depleted zone overlies it with low-pressure mineralogy of the oceanic crust. If the oceanic crust is converted to eclogitic mineralogy the mean density of the lithosphere becomes very close to that of undepleted mantle and the compositional buoyancy is eliminated. If this happens lithosphere is liable to subduction even if it is less than 40 Ma old.

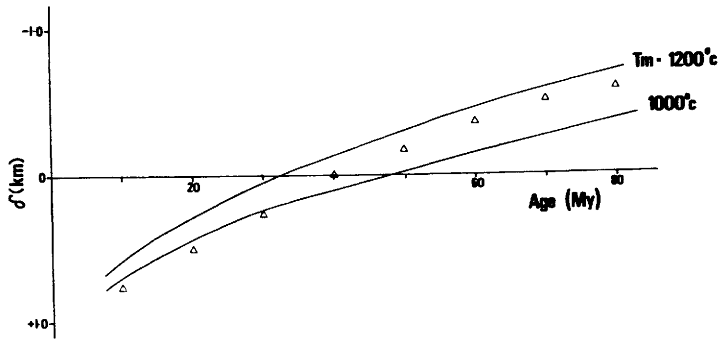

FIG. 2. The density defect thickness of the lithosphere as a function of age. Mean density of the lithosphere is less than that of underlying mantle for $\delta > 0$ and greater for $\delta < 0$. See text for further discussion.

4. Compositional density variations and the behaviour of subducted plates

In a subducting plate, the basaltic crust will transform to eclogite at a depth related to temperature but less than about 90 km. There should no longer be any compositional density deficiency associated with the lithosphere as a whole, and the thermal contribution to δ will provide a body force which may be dominant in driving the motion of plates (Forsyth & Uyeda 1975). The compositional density variation within the plate can still be important, however, since as a plate is heated during subduction, its hotter portions will no longer behave rigidly. To understand the possible consequences of this, we first consider density distributions within a descending lithospheric slab.

Density distributions within a descending lithospheric slab can be determined from the variation of temperature and composition within the slab; various models have been proposed (cf. McKenzie 1970, Oxburgh & Turcotte 1970, Minear & Toksöz 1970, Turcotte & Schubert 1973). We will follow that proposed by Turcotte & Schubert (1973), which assumes that subduction is a steady process. The temperature distribution within the slab just prior to subduction is that in the lithospheric thermal boundary layer given by equation (3). Heat is supplied to the slab by shear heating on the slip zone between the slab and the overlying mantle. Shear heating continues to the depth at which the top of the slab reaches the basalt solidus temperature. At greater depth, the temperature at the top of the slab is fixed at the basalt solidus temperature.

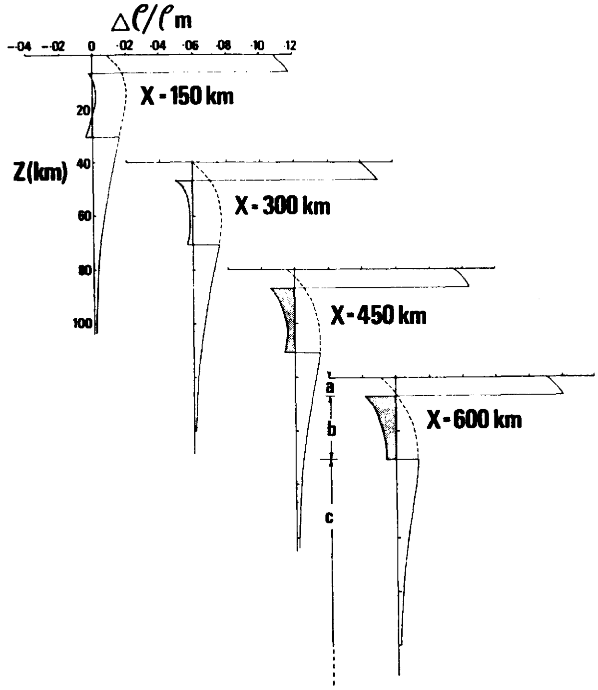

FIG. 3.

The buoyancy of the lithosphere at four distances, x, along its length from the surface; z is distance into the slab normal to its upper surface. The variation in density contrast between different parts of the slab and the surrounding mantle is given by $\Delta\rho/\rho_m$: regions with negative values are buoyant (stipple). The former oceanic crust is zone a; zone b is the depleted zone and c is the undepleted part of the lithosphere. See text for further discussion.

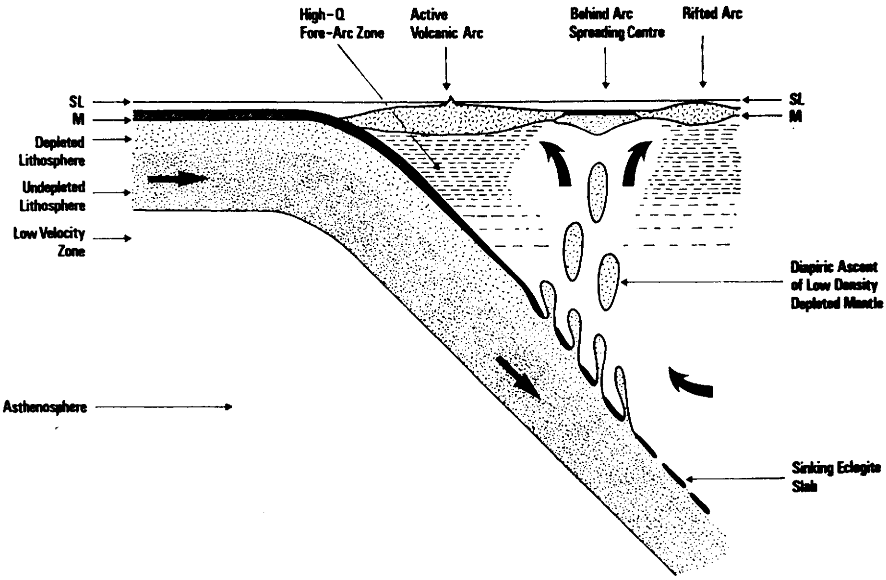

FIG. 4. Compositionally stratified descending lithosphere. Eclogite layer (black) sinks through less dense depleted mantle (light stipple) as the top of lithosphere is warmed. Depleted material continues to rise through the overlying mantle providing a mechanism for behind-arc spreading. See text for further discussion.

For the purposes of this discussion we ignore the olivine-spinel phase transition. Experimental evidence indicates that the composition-related density differences upon which our argument depends, persist through the phase change (Ringwood 1970).

The important variables which affect the temperature distribution are the subduction rate, the age of the plate being subducted, and the shear stress on the slip zone. We will consider as a typical set of conditions a subduction rate of 80 km/Ma, a plate age of 40 Ma, and a shear stress of 1 kb. Density distributions within the slab at various depths are shown in Fig. 3. In this figure, x is a coordinate measured along the dip of the slab, and z is a coordinate normal to the x direction.

The dashed curve in each case shows the thermal density variation in the cold, descending lithosphere. It decreases in successive profiles down the slab as the latter approaches the temperature of its surroundings. The density contrast between any part of the slab and its surroundings is the sum of the thermal and compositional components and is shown as the solid line. In the undepleted zone of the lower lithosphere dashed and solid lines are coincident. The eclogite zone representing the oceanic crust has such a dense mineral assemblage that it remains much denser than its surroundings at any depth—in contrast, the underlying depleted zone becomes buoyant relatively quickly. This unstable density stratification ($\Delta\rho \simeq 0{\cdot}4$ gm/cm^3) is likely to become important only when the upper part of the slab is sufficiently warm for it to behave in a ductile fashion. The eclogite layer should then break up and sink through the depleted zone (Fig. 4). Ringwood & Green (1966) have drawn attention to such consequences of the high density of eclogite. At the same time or possibly a little later, the depleted zone should separate from the colder and intrinsically denser part of the descending slab, and ascend either as a series of separate diapiric masses or as some kind of semi-continuous flow. Until that time there should be no *net* compositional density contrast between the slab and the mantle through which it sinks. Note that the effect of flow caused by the sinking of eclogite through the depleted layer is not taken into account in calculating the thermal structure of the slab. This flow would enhance the transfer of heat into the top of the slab causing more rapid heating than is predicted on the basis of heat conduction alone.

The depth at which such diapirs would begin to separate and their rate of ascent through the overlying mantle are difficult to predict since they depend on factors such as the geometry of the flow, the dependence of viscosity on temperature, and the temperature difference between the depleted material and that of the mantle through which it rises. It is clear, however, that the depleted material with a compositional density difference of 0·06 gm/cm^3 with respect to undepleted mantle at the same temperature must rise through the overlying mantle unless this region of the mantle is itself depleted or at least 600°C hotter than the rising material.

Such diapirs which have a contrast in composition rather than a contrast in temperature with the surrounding mantle could give rise to the ascending mantle flows which have been favoured by various authors (Karig 1971, Oxburgh & Turcotte 1971) to explain behind-arc spreading. These flows have previously been thought to be thermally driven.

If depleted mantle wells up to form the asthenosphere in behind-arc basins, it should give rise to a crustal structure and surface elevation pattern different from that in normal ocean basins. Assuming full isostatic compensation and crusts of similar thickness, a behind-arc basin should have the M-discontinuity somewhat shallower than under an oceanic area of the same age—approximately 180 m shallower for every 10 km of depleted, low density mantle underlying it and not present under the ocean. It also follows that the igneous layer of the behind-arc crust which is derived by the partial melting of the upwelling mantle must either be a further melt from mantle already once depleted, or a first melt from limited amounts of undepleted mantle carried up between or within the diapirs. In either case the volume of melt is likely to be less than at a mid-ocean ridge and in the second case it might vary in both space and time.

Observations on the crustal structure of behind-arc basins are at present inadequate to test these proposals. In general the basins are known to be close to isostatic equilibrium (Sclater *et al.* 1976) and to contain variable but relatively large volumes of sediment (Menard 1967). The basins range in depth from rather shallow, where the sediment cover is thick, to 6000 m in a few places. A variety of seismic refraction studies have been interpreted as indicating a thin and variable igneous layer to the crust (Sclater *et al.* 1976). In the absence of reliable reflection studies, further speculation is unwarranted at present.

The model proposed does not necessarily lead to the development of a single, localized spreading centre behind an arc; it is possible to visualize situations in which diapirs rise over a broader zone and give rise to a number of transient, separate, spreading zones. Equally there would probably be a tendency for the main zone of diapirism to migrate with time, as the amount of depleted mantle in any particular region overlying the slip zone becomes unusually high.

If subduction occurs against a continental margin, diapirs of depleted mantle are likely to rise against and possibly penetrate the base of the continental lithosphere. The effect of this activity should be to elevate the surface topography both as a result of heating and thermal expansion of the lithosphere, and of the addition of low-density depleted mantle beneath. The elevation of the Colorado Plateau and the Altiplano in the Andes may have occurred in this way.

The mineralogical zonation within the descending lithosphere has an important bearing on the 'eclogite engine' proposed by Ringwood & Green (1966) to drive the motions of plate tectonics. Drawing attention to the high density of eclogite, they argued that it should sink in 'average mantle' and that once the subducted oceanic crust had inverted it should drag the rest of the plate down with it. Because the concentration of dense garnet and pyroxene which occurs in the former oceanic crust, is approximately balanced an equivalent depletion in the underlying depleted zone of the slab, a descending slab with a eclogite capping has no net mineralogical density contrast with surrounding 'average mantle'. The eclogite engine cannot therefore function, at any rate, not in the way originally proposed.

5. Conclusions

The oceanic lithosphere is gravitationally stable on the asthenosphere until it is 40–50 Ma old. At that time the cooling and thermal contraction are sufficient to offset the buoyancy of the depleted upper mantle and the oceanic crust.

During subduction the oceanic crust inverts to eclogite and thereby eliminates the net mineralogical density difference between the descending lithosphere and the surrounding mantle.

As the top of the slab is heated during subduction it becomes ductile and the eclogite of the former oceanic crust should sink into low density depleted zone which it overlies.

The depleted zone is also less dense than the surrounding 'average mantle' and when it is sufficiently ductile should rise diapirically several hundred kilometres behind the trench. It may be the cause of behind-arc spreading.

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E. R. OXBURGH & E. M. PARMENTIER, Department of Geology and Mineralogy, Parks Road, Oxford, OX1 3PR.

Discussion

Dr B. J. Walton said that Dr Oxburgh appeared to imply that convection currents were likely to be responsible for plate movements. This did not seem to be necessary if, for example, a lithospheric dome developed by radioactive heating at the base of a large initially stationary area of continental crust, with the result that the lithosphere slid off the dome under its own weight, and started to descend into the lighter asthenosphere at subduction zones. In this case the flow through the asthenosphere would be from the trench to the oceanic ridge, opposite to that generally envisaged for convection currents.

Dr R. B. McConnell referred to the outflow of heat at the mid-ocean ridges referred to by Dr Oxburgh and asked whether he thought that the present plan of the world-wide rift system was due to an original pattern of convection cells in the Earth's mantle, as had been formerly suggested, or whether it was not more likely that the heat produced in the mantle by radioactive decomposition found its outlet by making use of segments of a pattern of deep dislocations imposed on the Earth in the early Precambrian? It had been pointed out by several authors that some of the ancestral pre-drift fractures, which were to develop into spreading oceans, can be seen to continue into the continents as deep faults which had been shown in some instances to be resurgent Precambrian lineaments, some segments of which might have served as outlets for convection heat and developed into spreading oceans. Other Precambrian lineaments, or segments of lineaments, remained intracontinental and were inhibited from widening by pressure from actively spreading mid-ocean ridges.

Dr Oxburgh replied to Dr Walton that he did indeed believe that convection was responsible for plate movements and that the motion of plates of finite thickness made mantle convection inevitable. This did not, however, imply any regular system of currents. The processes described by Dr Walton were necessarily convective and were intended by the speaker to be included in his use of that term. In answer to Dr McConnell, the speaker said that the origin of fractures which were to develop into spreading centres was an unresolved problem. It seemed likely that any major lithospheric weaknesses might well control the pattern of fracture initiated by mantle convection or any other major tectonic process.