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INTERACTION OF MANTLE DREGS WITH CONVECTION: LATERAL HETEROGENEITY AT THE CORE-MANTLE BOUNDARY

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Abstract. Preliminary numerical models indicate that chemically denser material (dregs) at the base of the mantle would have substantial lateral variations in thickness induced by convection of the overlying mantle, and might well form discontinuous aggregations below mantle upwellings. A model with a density contrast of about 2 per cent and an initial uniform thickness of the denser layer of 100 km yields a discontinuous distribution with maximum thickness 230 km and bottom topography of several kilometers amplitude, in reasonable accord with recent seismological observations of vertical and lateral structure. Heat flux out of the core is probably strongly modulated laterally by mantle convection, while mantle dregs will complicate and possibly amplify this effect. Such modulation may be relevant to long-term (10^7 - 10^8 year) variations in the magnetic field.

Introduction

Both seismologists and geochemists have appealed to the existence of a chemically distinct, denser layer at the base of the mantle. To the long established observations of lower seismic velocity gradients in the lowermost 200 km have been added specific suggestions of increases in compressional wave [Wright et al., 1985] and shear wave [Lay and Helmberger, 1983; Lay, 1986] velocity of 2.5-3 per cent 200-300 km above the core-mantle boundary and several observations of lateral heterogeneity in the vicinity of the core-mantle boundary [Morelli and Dziewonski, 1986; Creager and Jordan, 1986; Lavely and Forsyth, 1986]. Petrologists have suggested that subducted lithosphere, or the crustal component thereof, might sink to the base of the mantle because of intrinsically greater density than the surrounding mantle [Dickinson and Luth, 1971; Ringwood, 1975]. Hofmann and White [1982] appealed to this mechanism to provide long-term storage of subducted oceanic crust to explain isotopic apparent ages of several billion years in mantle-derived rocks [Chase, 1981].

Mantle convection can be expected to have significant effects on such a layer, and may even prevent it from forming or surviving. Even in the absence of such a layer, mantle convection will cause strong lateral variations in the heat flux from the core [Davies, 1986]: the addition of a chemical layer will amplify and complicate this modulation, with possible implications for the magnetic field. Neither of these effects has been quantitatively explored. Gurnis [1986a] simulated the interaction of dense subducted material and mantle convection using "heavy" tracers in a numerical model and found that segregation of dense material to the bottom was quite inefficient for likely mantle parameters. Only in the event of complete separation of the oceanic crust from the more buoyant, depleted lithosphere component might some segregation occur. Nevertheless, some segregation may still occur, and Gurnis' models did not address the survival of a pre-existing layer, such as might date from the formation of the earth.

In this paper, we report some initial results of numerical models of the interaction of convection with a pre-existing, thin layer of denser fluid. Although only a limited range of steady models has yet been explored, the results indicate that the interaction is likely to be strong in the mantle.

Model Description

We have considered two-dimensional convection in a 4x1 box, as illustrated in Figure 1. (A smaller width box might be more appropriate, but our main intention here is just to demonstrate the phenomena.) This type of model, and the relevant equations and methods of solution, have been described in detail by Davies [1986], Gurnis and Davies [1986] and Gurnis [1986a]. A single long cell is maintained in the box by imposing a uniform horizontal velocity on the top boundary and a higher viscosity "lid" in the top 15% of the layer (with maximum viscosity 15 times that of the fluid interior). This is intended to simulate the role of the lithosphere in the mantle. Its primary effect is to control the structure of the flow by determining where the main vertical flows occur and by suppressing small-scale instabilities in the top boundary layer: both of these effects are to be expected in the mantle. The boundary velocity is chosen to be comparable to the free flow velocities. The bottom and sides of the box are free-slip and the sides are insulating. The main differences from previous models are in the mode of heating and the method of including dense fluid.

Davies [1986] noted that the most appropriate heating mode for models of the deep mantle was to combine internal heating with an isothermal base, a combination which did not seem to have been previously considered. The core viscosity is presumably so low and its mixing so relatively rapid as to keep lateral temperature variations in the core negligibly small compared to those in the mantle [Rochester, 1986], so the base of the mantle will be isothermal to a good approximation. While most of the heat emerging at the earth's surface is probably generated in the mantle or due to mantle cooling, something less than 25 per cent may be flowing from the core [e.g., Stacey, 1977, Davies, 1980]. Therefore we have included uniform internal heating in the models, and adjusted the value until about 85 per cent of the steady-state surface heat flux was being generated internally, the remainder coming from the lower boundary.

Rather than use the "heavy tracer" method of simulating dense fluid [Gurnis, 1986a], we found it was necessary here to use tracers to delineate the interface between the fluids, as done by Christensen and Yuen [1984]. (The heavy tracers undergo

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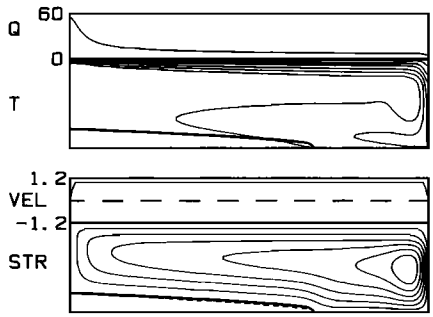

Fig. 1. Thermal structure (top) and flow structure (bottom) of a convection model. Parameters are given in Table 2. A denser fluid, with density parameter B = 0.25, was introduced initially with a uniform thickness of 0.1D at the bottom of the box. Q denotes top surface heat flux (dimensionless units), T denotes temperature (isotherms), VEL denotes the imposed top surface horizontal velocity (dimensionless) and STR denotes the stream function (streamlines). The heavy curve shows the interface between the fluids and the dashed curve is the zero streamline. The denser fluid covers only part of the bottom surface, and terminates in a steep "nose".

a spurious sedimentation, due to approximating the continuous density anomaly by points, which was not negligible in the present context.)

The standard definition of Rayleigh number, R_T, in terms of the temperature difference, ΔT, across the layer, is used here. See Davies [1986] for more details on this and the following quantities. The proportion of heat generated internally is μ_T. The velocity, V_P, of the top boundary (or plate) is represented by the dimensionless Peclet number, Pe. The density difference $\Delta\rho$ between the bottom layer and the rest of the fluid is included as its ratio with a characteristic thermal density anomaly:

$$A = \Delta\rho/\rho_0\alpha\Delta T \qquad (1)$$

where ρ_0 is the normal fluid density. Values of the relevant quantities for both the mantle and the models are given in Tables 1 and 2.

The Rayleigh number for the mantle (or the lower mantle) is too high to model with a reasonable number of grid points in our finite difference calculations. The models were therefore run with $R_T = 5\times10^5$ and it was also necessary to use a smaller value of Pe (300) in order that the contributions

TABLE 1. Primary Mantle Parameters

Symbol	Quantity	Value
D	depth of mantle	3000 km
g	acceleration due to gravity	10 m s^{-2}
ρ_0	normal fluid density	4×10^3 kg m^{-3}
$\Delta\rho_b$	density jump at CMB	4.4×10^3 kg m^{-3}
α	thermal expansion coefficient	2×10^{-5} K^{-1}
κ	thermal diffusivity	10^{-6} m^2s^{-1}
K	thermal conductivity	3 $Wm^{-1}K^{-1}$
η	dynamic viscosity	10^{21} Pa s

TABLE 2. Mantle and Model Parameters

Parameter		Model value	Mantle value
R_T	Rayleigh number	5×10^5	3.2×10^7
ΔT	temperature	23.7 K	1515 K
μ_T	internal heating	9.5	-
Pe	Peclet number	300	4800
V_p	plate velocity	3.2 mm yr^{-1}	50 mm yr^{-1}
B = 0.25			
A	density anomaly ratio	1.0	0.25
$\Delta\rho$	chemical density difference	1.9 kg m^{-3}	30.3 kg m^{-3}
B = 1.0			
A		4.0	1.0
$\Delta\rho$		7.6 kg m^{-3}	121 kg m^{-3}

to the flow from the thermal buoyancy and the moving upper boundary remained in proportion [Davies, 1986].

An analogous argument applies to the value of the density ratio A. We are interested in finding the amplitude of topography on the internal fluid interface, which will be determined by a balance of stresses arising from the driving thermal buoyancy and from the sloping interface between fluids of different density. The thermal buoyancy stresses are proportional to $R_T^{2/3}$ [Davies, 1986]. Since we are assuming that the interface slope is unchanged when R_T is changed from the model value to the mantle value, its influence is proportional simply to $\Delta\rho$. Thus we must assume that $\Delta\rho$ increases only in proportion to $R_T^{2/3}$, rather than R_T, so that, from (1), A is proportional to $R_T^{-1/3}$, rather than being constant. Thus a

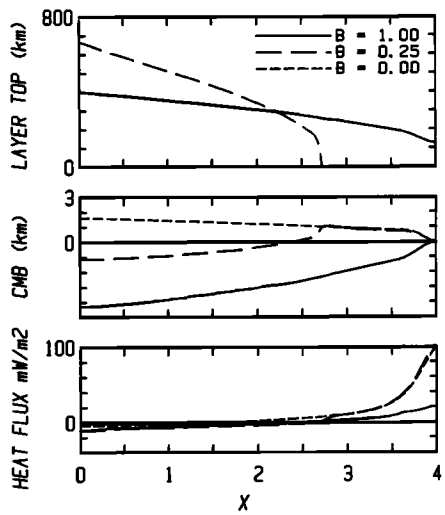

Fig. 2. The shape of the top of the denser fluid, the core topography (CMB) and the bottom heat flux for two cases started with a dense layer of uniform thickness 0.1 (300 km). The case with density parameter B = 0.25 is that shown in Fig. 1. The bottom topography and heat flux are shown also for the case with no dense layer (B = 0) for comparison. Note the substantial reduction in heat flux when the layer covers the entire base.

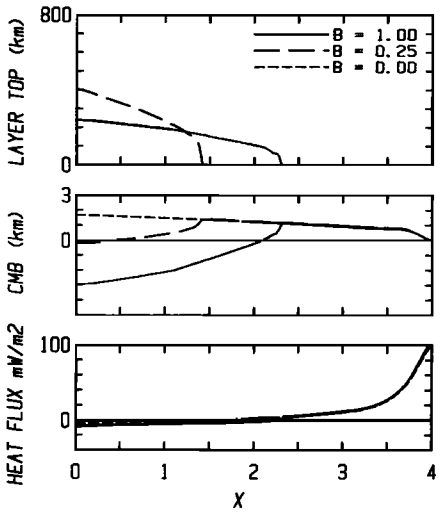

Fig. 3. As in Figure 2 for cases started with a uniform layer thickness of 0.033 (100 km). The smaller volume of dense fluid is unable to cover the bottom in either case here.

model value of A = 4 will scale to a mantle value of A = 1 (see Table 2). This scaling has been confirmed by Gurnis [1986b].

It is useful to define a density anomaly parameter, B, which is equal to A for the mantle, but which explicitly corrects for this dependence on R_T:

$$B = A(R_{T_{model}}/R_{T_{mantle}})^{1/3} \qquad (2)$$

so that similar values of B should correspond to similar interface slopes.

Deflections of the bottom boundary are caused by both thermal and chemical density variations. They are calculated here from the total normal stress on the boundary, as described by Davies [1986], the only modification being to take account of the change in core-mantle density difference when anomalous fluid is adjacent to the boundary.

Results

A series of four models is presented with two different values of B and two initial thicknesses of the anomalous layer. The thermal and flow structures for a typical case are shown in Figure 1. Parameters of this and the other models are given in Table 2. The internal interface between the fluids is shown by the heavy curve. The denser fluid has been swept to one side of the box in this case, with part of the lower boundary free of any anomalous fluid. The main convective circulation is deflected around the interface; the zero streamline (dashed) would coincide exactly with the interface in the exact steady state.

The model shown in Figure 1 is not an exact steady state. All of the models were run for 4-8 transit times, i.e., for 4-8 times the time required to travel a distance D at velocity V_P. At later times there will continue to be some change because the heat flux through the layer has been suppressed and the system will have to adjust by developing a thermal boundary layer at the interface. This could have significant effects, and might even destabilise the layer. This longer-term evolution has not yet been explored.

In Figure 2 we show the layer thickness, bottom topography and bottom heat flux for the two cases starting with a thick layer. Results have been scaled to the mantle Rayleigh number in the manner described above. The initial layer thickness (0.1D) may be unrealistically large, but these cases illustrate the phenomena. Whereas with B = 0.25 (the case shown in Figure 1) the bottom layer is discontinuous, with B = 1 it is continuous.

The effect of the denser layer on the bottom topography is the opposite of the purely thermal effects (B = 0, short-dashed) and in all cases shown here is larger in magnitude, although smaller for the smaller value of B. Thus the correlation between bottom topography and internal thermal structure is the reverse of the purely thermal case. This provides a possible explanation for the results of Morelli and Dziewonski [1986], who find the core-mantle boundary to be depressed under mantle regions with low seismic velocities, which are inferred to be warmer.

Even in the absence of a denser layer, the heat flux from the bottom bottom boundary is strongly modulated by convection (Figure 2, bottom, short-dashed). The flux actually is negative over part of the boundary, a possibility which was noted by Davies [1986]; whether or not this occurs will depend on details of the boundary and the proportion of internal heating. Regardless of this possibility, the heat flux out of the core will be strongly concentrated where cooler descending mantle approaches the core.

The short-term effect of a dense layer on the bottom heat flux is to reduce it. The effect is small except where the heat flux was previously large (Figure 2, B = 1). Although in the longer term this effect may change, these results demonstrate that relatively large and rapid redistributions of core heat flux may be produced by changes in mantle flow structure.

Figure 3 shows results for layers with an initial

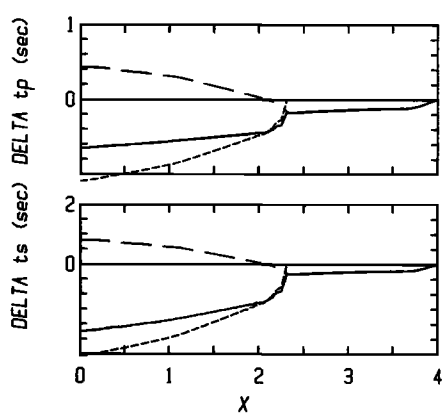

Fig. 4. Estimated two-way vertical travel time anomalies of P and S waves for the case with B = 1, Figure 3. The advance due to the faster layer (short dashed) outweighs the delay due to depression of the bottom surface (long dashed). The solid curves show the total anomaly. This Figure is for guidance only, and would be most relevant for core-reflection phases. Velocity anomalies in the layer were estimated by assuming $(d \ln V_P/d \ln \rho) = 1$ and $(d \ln V_S/d \ln \rho) = 1.5$. Empirical estimates of these parameters are 1 - 1.3 for V_P and 1 - 2 for V_S [Davies, 1976].

thickness of 0.033D (100 km), which is closer to the usual seismological estimates. Results are quite analogous to those in Figure 2, except that now neither layer is continuous. The total amplitudes of layer thickness and topography are smaller, an effect due mainly to the smaller volume of anomalous fluid: the slopes of the interfaces are similar to those in Figure 2, consistent with the expectation that the perturbing convective stresses are similar.

A guide to seismological effects is given in Figure 4, which shows estimates of the two-way vertical travel time anomalies of P and S waves for the thin layer with B = 1 (Figure 3). In both cases, the faster waves in the anomalous layer more than compensate for the delay due to the depression of the bottom boundary. Thus the net effect of a denser layer might appear to steeply incident waves as an elevation of the core-mantle boundary.

Discussion

Although preliminary, these results very clearly imply that any chemically distinct denser material at the base of the mantle would not be distributed in a layer of uniform thickness, and might well occur as discontinuous segregations concentrated under mantle upwellings. This accords well with several recent seismological studies indicating significant lateral heterogeneity near the core-mantle boundary [Morelli and Dziewonski, 1986; Lavely and Forsyth, 1986; Creager and Jordan, 1986]. If we take as constraints the thickness of about 280 km observed by Lay and Helmberger [1983] and the 10 km peak-to-peak amplitude of core topography estimated by Morelli and Dziewonski [1986], then the case with initial (or average) layer thickness of 100 km and B = 1 (Figure 3) seems to be the most appropriate. This yields a density anomaly of about 0.12 g/cm^3 or 2 per cent, which corresponds reasonably with the long-period shear velocity jump of 2.75 per cent at 280 km found by Lay and Helmberger [1983] and the short-period compressional velocity jump of 2.5 - 3 per cent at 200 km suggested by Wright et al. [1985].

The model results shown correspond to times of 150-250 million years since the introduction of the anomalous fluid layer into the model. While there will be some significant changes in the thermal regime on longer time scales, the steady state may not be particularly relevant, since the mantle flow structure probably changes on this time scale due to plate evolution at the surface. Nevertheless, it will be important to extend the models to longer times.

The results on heat flux are of particular interest for their possible relevance to long-term changes in the magnetic field, since they indicate large lateral variations in the thermal boundary condition on the core which would presumably have a substantial effect on core flow structure. Futhermore, it is possible that this boundary condition could change quite rapidly (in a few times 10^7 years) in response to a new influx of cooler mantle material.

Although the term "dregs" is usually applied to solid sediment, it seems also to describe accurately and concisely the liquid "sediment" considered here. To be consistent, of course, we should also regard continental crust as scum.

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