The influence of phase transitions and chemical heterogeneity on mantle convection

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The influence of phase transitions and chemical heterogeneity on mantle convection[1])

By U. CHRISTENSEN[2])

Extended abstract

Seismic discontinuities in the earth's mantle can be caused either by isochemical phase transitions, or by a change in composition, or perhaps by a combination of both. The major discontinuity at 400 km depth is almost certainly due to the transformation of olivine into the β-spinel crystal structure. The 670-km-discontinuity is probably caused by the change of the spinel structure into a combination of magnesium- perovskite (Mg, Fe)SiO_3 and magnesiowüstite (Mg, Fe)O. However, it is not clear whether or not an additional change of composition is present. The sharpness of the 670-km-discontinuity, which is deduced from the reflection of seismic body waves, would support the hypothesis of a superimposed change in chemical composition. Finally, the so-called D″-layer, i.e. the lowermost 200 km of the mantle above the core, has aroused interest in recent years. It is characterized by low gradients of the seismic velocities, considerable heterogeneity, and a seismic discontinuity 200–300 km above the core-mantle boundary has been identified in some places (LAY 1989, WEBER & DAVIS 1990). Because no phase transitions have been found in high-pressure experiments for this depth range, this would indicate chemical heterogeneity or perhaps very strong thermal gradients.

The major forms of vertical convective flow in the mantle are subducting lithospheric slabs and mantle plumes, i.e. narrow columns of hot rising flow. However, how deep do slabs descend into the mantle and from what depth do the plumes come? This question is discussed now for two decades, but a generally accepted answer is not reached. According to one end-member model (e.g. RICHTER & MCKENZIE 1981, ANDERSON 1987) slabs penetrate no deeper than to about 700 km depth, and plumes would come from a thermal boundary layer between upper and lower mantle which divides the mantle into two separately convecting layers with very limited mass exchange between them. The other end-member model is that of whole mantle convection (e.g. PELTIER & JARVIS 1982, DAVIES 1988) without a significant barrier to vertical flow at any depth. The nature and the physical parameters of the seismic discontinuities in the mantle are the controlling factors on the amount of mass exchange across

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[2]) Max-Planck-Institut für Chemie, Saarstrasse 23, D–6500 Mainz, Germany.

them. Numerical model calculations of mantle convection are a useful tool to understand and quantify the influence of the various parameters on the form of convection.

In the case of a major compositional change across the 670-km-discontinuity, the chemical density difference in relation to the density increase by thermal contraction in cold descending slabs is the important parameter. In two-dimensional computer models (Christensen & Yuen 1984) it was found that, for chemical density contrasts of the order of 5% or larger, descending slabs would be stopped at the discontinuity, bend around, and feed into a return flow in the upper mantle (Fig. 1). The slab causes a depression of the discontinuity in the subduction zone by about 100 km. When the chemical density contrast is less than 5%, slabs would deeply penetrate into the lower mantle (Fig. 2), and when it would be less than about 2% rapid mixing of the two different reservoirs of upper and lower mantle should occur, destroying the chemical layering quickly. Kincaid & Olson (1987) found semi-quantitatively the same behaviour in laboratory models of slab subduction, using corn syrup as a model fluid.

For the influence of phase transformations on the dynamics of convection the important parameter is, besides the density change, the Clapeyron slope $\gamma = dp/dT$ of the phase boundary. A positive slope will enhance convection, whereas a negative slope retards it, and, in the extreme case, will cause separate convecting layers above and below the phase boundary. The principal effect is demonstrated in Fig. 3. Inside a cold slab, a positive dp/dT causes an elevation of the high-pressure-phase to shallower

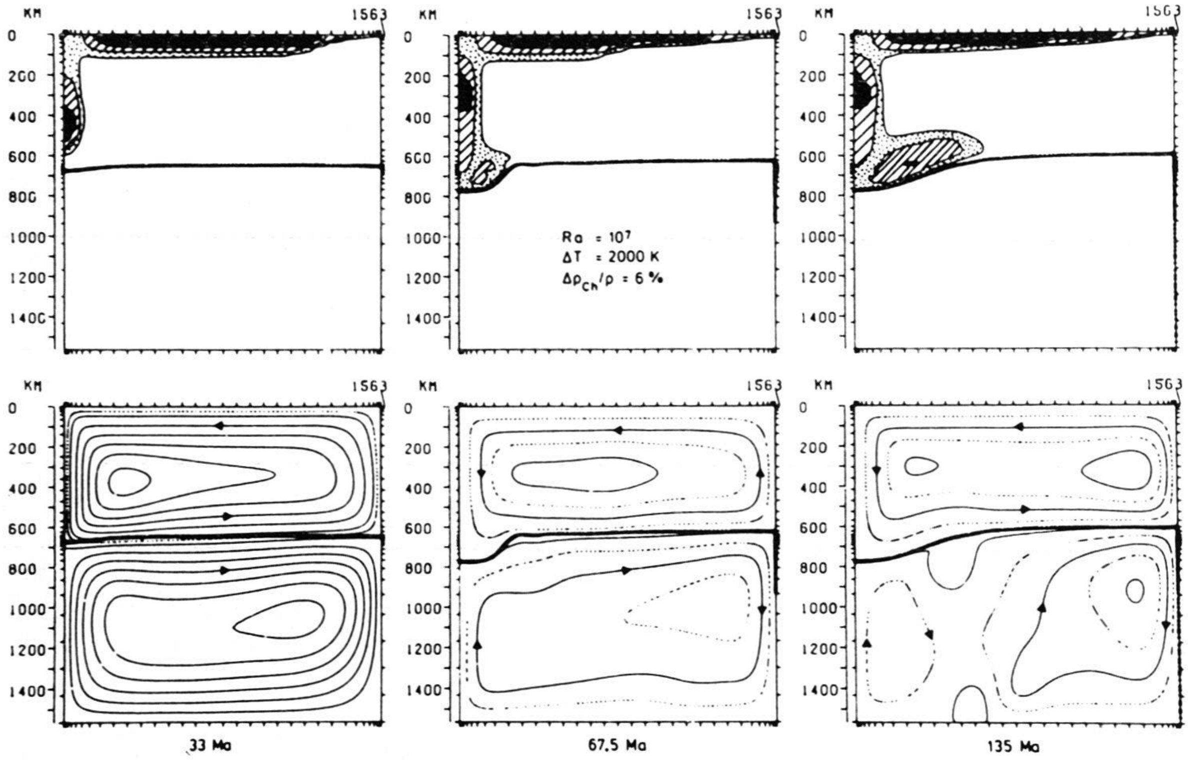

Fig. 1. Two-dimensional convection model including a subducting lithospheric slab with non-linear temperature-dependent rheology. A chemical boundary with 6% density contrast is at about 650 km depth. Three snapshots in time are shown. In the upper diagram the effective viscosity is shown, where the slab is easily identified, in the lower diagram streamlines.

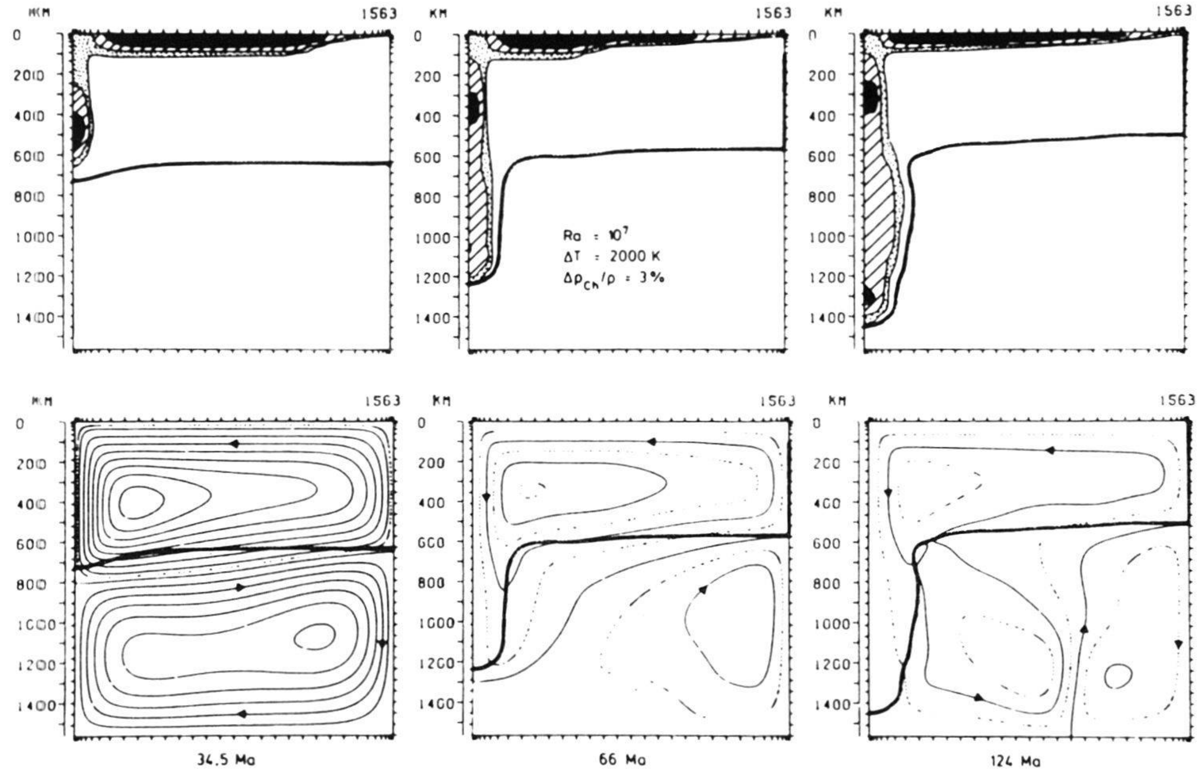

Fig. 2. As Fig. 1, but for a chemical density contrast of 3%.

depth, i.e. to lower pressure. The strong gravitational body force connected with the elevated dense material helps to pull the slab down. When the Clapeyron slope is negative, a depression of the lower-density phase is created whose buoyancy opposes subduction. The relevant transition from the spinel-form of Mg_2SiO_4 to the perovskite-plus-magnesiowüstite assemblage at about 670 km depth has a negative Clapeyron slope; high-pressure experiment arrive at values of dp/dT in the range of −2 to −3 MPa/K. Again, to determine if this would be sufficient to cause layered convection, numerical convection calculations have been performed (Christensen & Yuen 1984, 1985). We found that a critical value of dp/dT in the range of −4 to −8 MPa/K is required to inhibit a strong flow across the boundary. Some weak "leaking" of material across the boundary still occurs in the layered regime, and when dp/dT is close to the critical value intermittent layering and penetration across the phase boundary is possible. Because the critical Clapeyron slope seems not to be reached for the 670-km-discontinuity the question arises if a combination of the phase transition and a slight superimposed change in the composition could cause layered convection. In Fig. 4 all the data from the various model calculations are combined. Due to uncertainties about various important material parameters which play a role for convection, the uncertainty in the location of the boundaries between the convective regimes is probably of the order of 30−50%. If we should indeed have separate convection in the upper and the lower mantle, the most plausible mechanism might be a compositional density change of at least 2−3% superimposed onto the phase transition with a Clapeyron slope of around −3 MPa/K. Recently Machetel & Weber (1990) found in

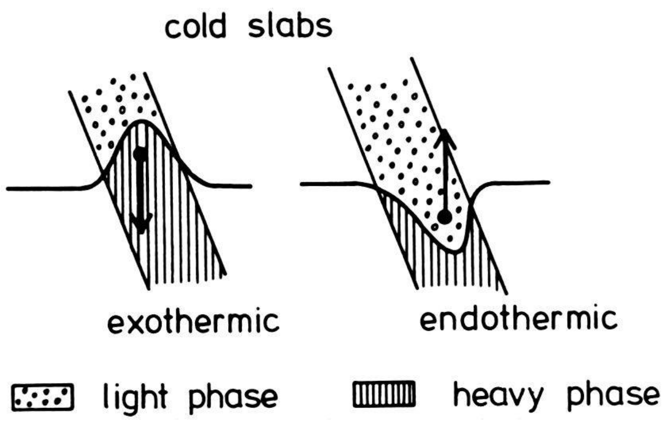

Fig. 3. Cartoon showing the principal dynamic influence of an exothermic phase boundary (dp/dT > 0) and an endothermic boundary (dp/dT < 0) on a descending lithospheric slab.

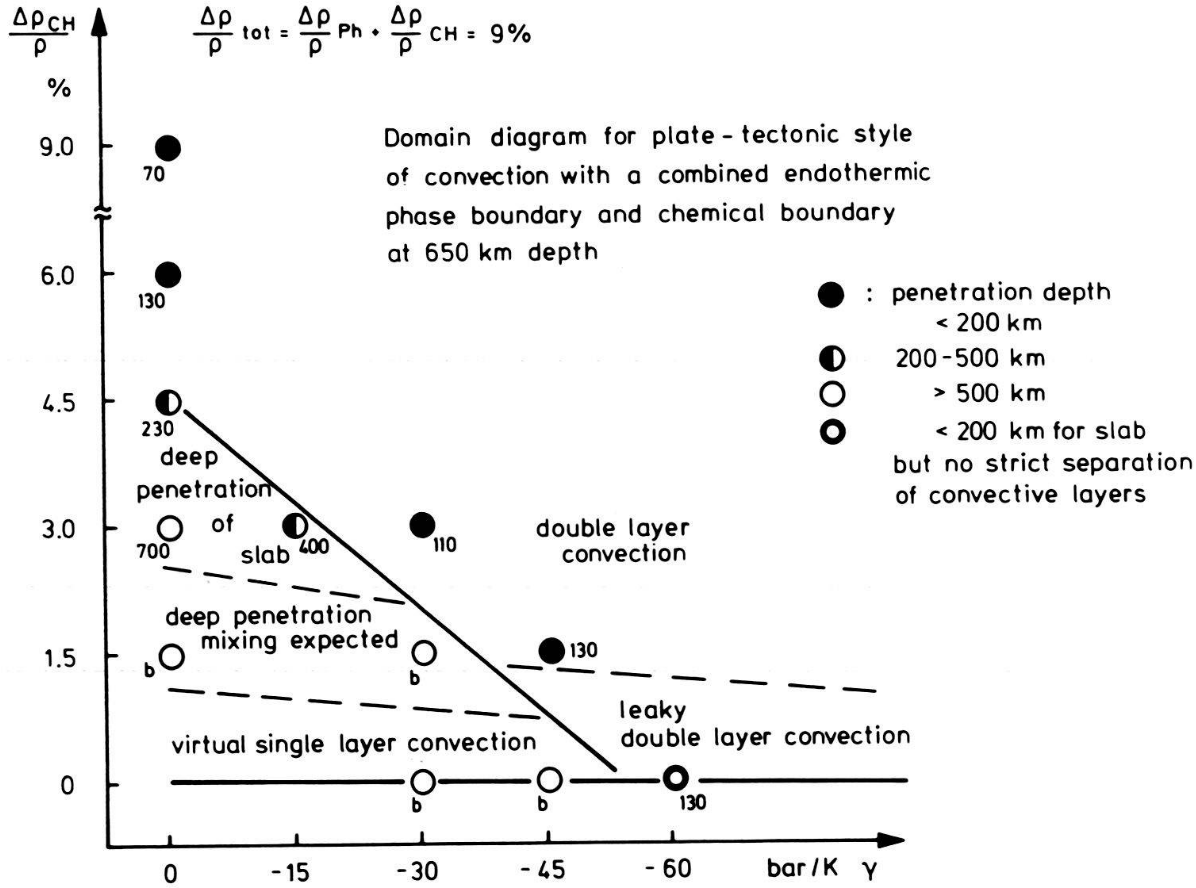

Fig. 4. Domain diagram for the style of mantle convection, summarizing the results from numerical model calculations with a descending slab. Parameters are the chemical contribution $(\Delta p/p)_{Ch}$ to the density contrast across the 670-km-discontinuity, $(\Delta p/p)_{Ph}$ is the contribution of the phase change to the density contrast and both combined are taken to be 9%, $\gamma = dp/dT$ is the Clapeyron slope of the phase transition [10 bar/K = 1 MPa/K].

numerical models with a more realistic flow geometry than in our calculations (where descending slabs are guided vertically by the side boundary of the model, compare Figs. 1 and 2) and assuming a somewhat lower coefficient of thermal expansion, that the intermittent flow regime could occur already for a Clapeyron slope of −2 MPa/K. This suggests that the parameter range for a hybrid form of convection might be broader than the range indicated in Fig. 4.

Comparing calculated density from experimental equation-of-state data for the relevant minerals with the density in the earth inferred from seismology, it had been concluded that the lower mantle must be more rich in iron (KNITTLE & JEANLOZ 1987). However, according to more recent data a uniform mantle composition seems preferred (CHOPELAS & BOEHLER 1989). More precise data from high-pressure research will probably enable us in the near future to decide conclusively whether or not the composition of the lower mantle is the same as in the upper mantle. So far, two-layer convection cannot be ruled out from considerations of the dynamical mechanism which might prohibit strong mass flux across the 670-km-discontinuity. On the other hand, we have no indication from seismological studies for a strong downward deflection of the discontinuity in the region of slab subduction (RICHARDS & WICKS 1990), which must necessarily occur if slabs are stopped at the boundary between upper and lower mantle. From this point of view the end-member model of strictly layered convection appears to be somewhat unlikely. Besides the simple whole-mantle model, a hybrid form of single and double-layer convection caused by moderate chemical density gradients below the 670-km-discontinuity (e.g. SILVER et al. 1988) or a sufficiently negative Clapeyron slope of the phase transitions (MACHETEL & WEBER 1991) could be possible and deserves further investigation. Even more complex models with additional chemical stratification within the transition zone of the upper mantle (ANDERSON & BASS 1986) or near the 670-km-discontinuity (RINGWOOD & IRIFUNE 1988) have been proposed, however, the stability of such internal layer in a dynamic mantle appears to be doubtful (e.g. CHRISTENSEN 1988).

When there is no strong barrier for vertical mass flux between upper and lower mantle, mantle plumes are likely to arise from a hot thermal boundary layer above the core-mantle boundary (D″-layer). Thus the basalts erupted at hot-spots may sample the deepest part of the earth's mantle, in contrast to mid-ocean ridge basalts, which sample the upper mantle. Numerical and laboratory experiments indicate that under conditions relevant for the mantle convection is unsteady (e.g. OLSON 1989). Especially the hot lower thermal boundary layer is expected to be unstable; it ejects thermal diapirs which rise through the mantle followed by thin trailing pipes feeding the diapir with more hot material (Fig. 5). The diapiric head of the plume would finally disperse in the asthenospheric low-viscosity channel underneath the plates, and the trailing pipe would remain as a quasi-steady plume causing the chain of volcanic islands forming a hot-spot track. Recently evidence has been presented that continental flood basalts could be caused by the arrival of the diapiric head of a mantle plume (e.g. RICHARDS et al. 1989). In the conduit-like plumes disturbance may propagate as solitary waves, i.e. there is a local enhancement in plume width and material flux, which travels upwards along the conduit without changing the shape. Such solitons would trap material which would be brought from the core-mantle boundary to the surface without contamination with adjacent mantle (WHITEHEAD & HELFRICH 1988).

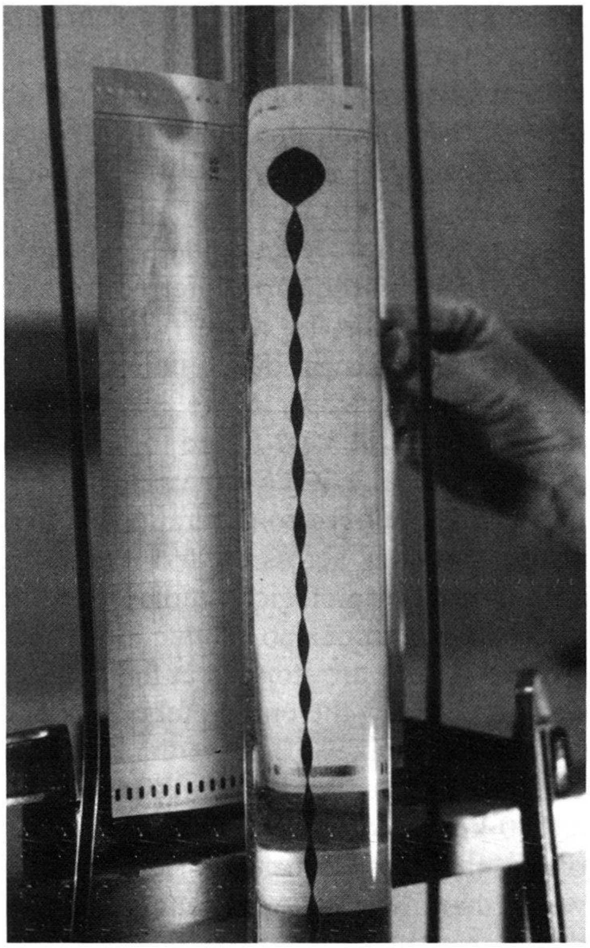

Fig. 5. Diapir obtained by injecting a buoyant liquid of low viscosity into a column of corn syrup. Note the pearl-string instability rising as a special kind of solitary wave in the feeding pipe underneath the diapiric head. Although in this experiment a chemical plume was studied, thermal plumes in the mantle may behave in a similar manner.

The difference in various isotope ratios (e.g. WHITE 1985) and trace element concentrations between hot-spot basalts and mid-ocean ridge basalts makes it unlikely that a simple whole mantle convection model can give a complete description of the structure and dynamics of the mantle. Although there is some controversy about the efficiency of mixing by mantle convection, it appears likely that whole mantle convection would erase any strong isotopic differences (CHRISTENSEN 1989a). A new facet in the set of possible scenarios is the hypothesis that the D″-layer is not simply a thermal boundary layer but also a chemical boundary layer of dense silicates, which seems compatible with the seismic evidence which is so far available (LAY 1989, WEBER & DAVIS 1990). For the origin of chemical heterogeneity at the base of the mantle several hypotheses have been put forward. It might be primordeal, it could be a "dreg"-layer of dense material segregated from the mantle, a "slag"-layer of silicates expelled by the cooling core, or the result of chemical reactions between core and mantle. The second source seems the most plausible one, although all might contribute in various degrees.

The subduction of oceanic crust is the main possible source of mantle heterogeneity that we know of. At lower mantle pressures, a rock of basaltic or eclogitic com-

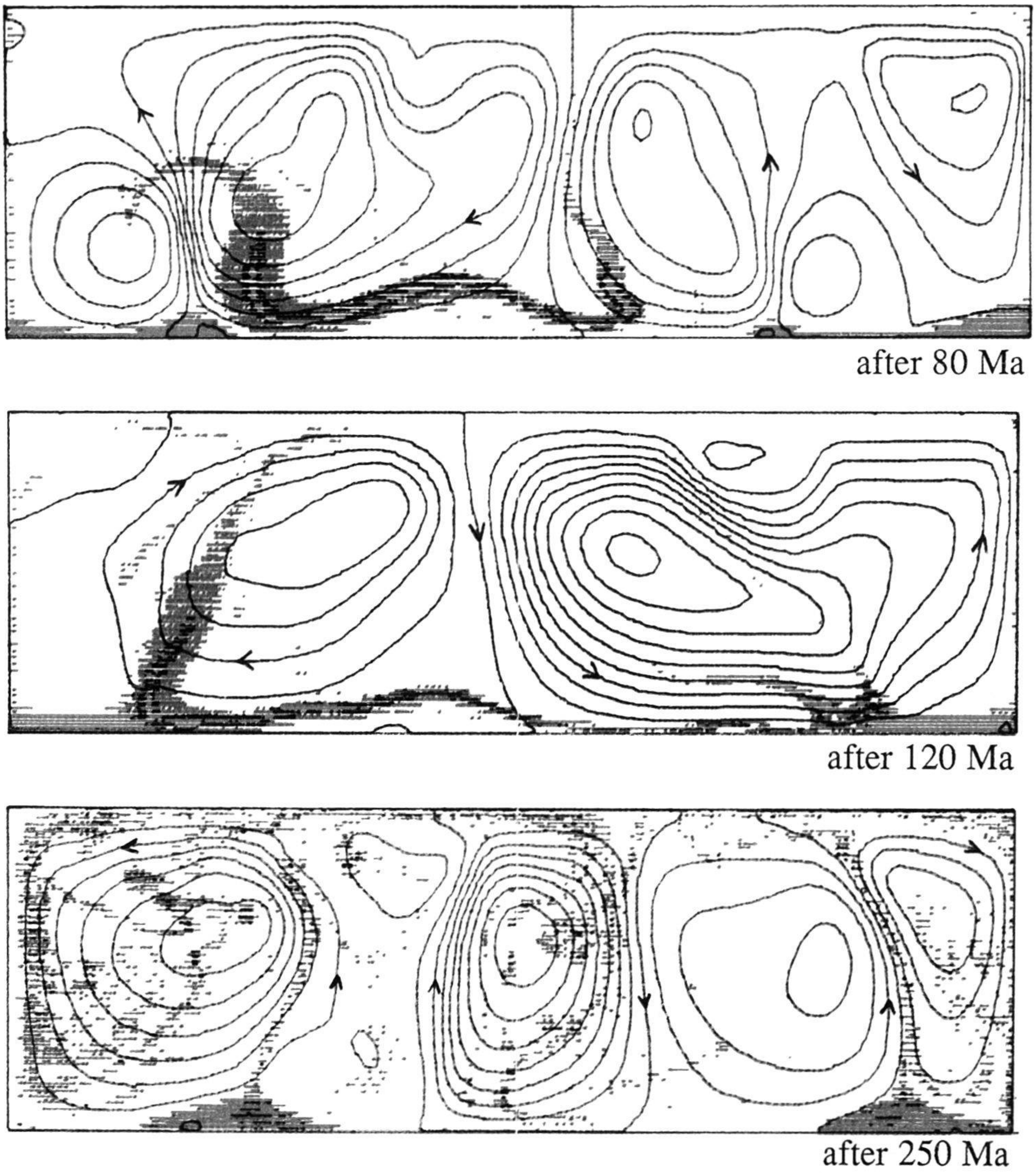

Fig. 6. Numerical convection model with subduction of a chemically layered lithosphere and an initial dense bottom layer. In red dense (eclogitic) crust, in blue buoyant harzburgite. After 80 Ma: Subduction is completed and the initially homogeneous bottom layer is swept into isolated pools. After 120 Ma: Part of the subducted crust is added to the pool to the left. After 250 Ma: The harzburgite and the remainder of the crust get mixed into the convection cells. The pools can be seen to leak into the rising currents.

position is probably denser by a few percent compared to normal mantle (IRIFUNE & RINGWOOD 1987). This suggests the possibility of a downward gravitational segregation of the former crust which could accumulate at the core-mantle boundary. Some arguments against such segregation have been made: that the crust is too thin for being able to sink independently from the ambient mantle at a significant velocity despite its excess density, and that the crust is tightly connected to a layer of buoyant material of harzburgitic composition, which is the residue of basalt extraction at the mid-ocean ridge, and which causes the entire assemblage to be neutrally buoyant.

Numerical model experiments simulating the subduction of a chemically layered lithosphere have been performed (CHRISTENSEN 1989b), with a thin layer of crust, being dense after subduction and transformation to eclogite, which overlies a thicker layer of buoyant harzburgite. In cases where the viscosity is assumed to be constant, no crustal segregation and accumulation is found. However, when the viscosity in the lower thermal boundary layer (i.e. the D″-layer) is reduced by a strong temperature-dependence of the rheology, part of the subducted crust was found to separate from the harzburgite and settle to the bottom. The segregation effect is enhanced when the thermal expansion coefficient is taken to decrease with depth (CHOPELAS & BOEHLER 1989) and when there is already a layer of previously segregated crust at the bottom (Fig. 6). Although the physical key parameters, like density contrast and viscosity, are notoriously uncertain in the lowermost mantle, the model calculations suggest that the mantle dreg origin of a chemical D″-layer is a viable hypothesis, and that 3–30% of subducting oceanic crust may feed into the chemical boundary layer.

Unless the amount of dense material at the bottom of the mantle is large, it would be swept into isolated pools (or "anti-continents") residing at points of the core-mantle boundary from which thermal plumes rise (Fig. 6). The rising plumes are found to entrain some of the dense pool material, which therefore influences the geochemical

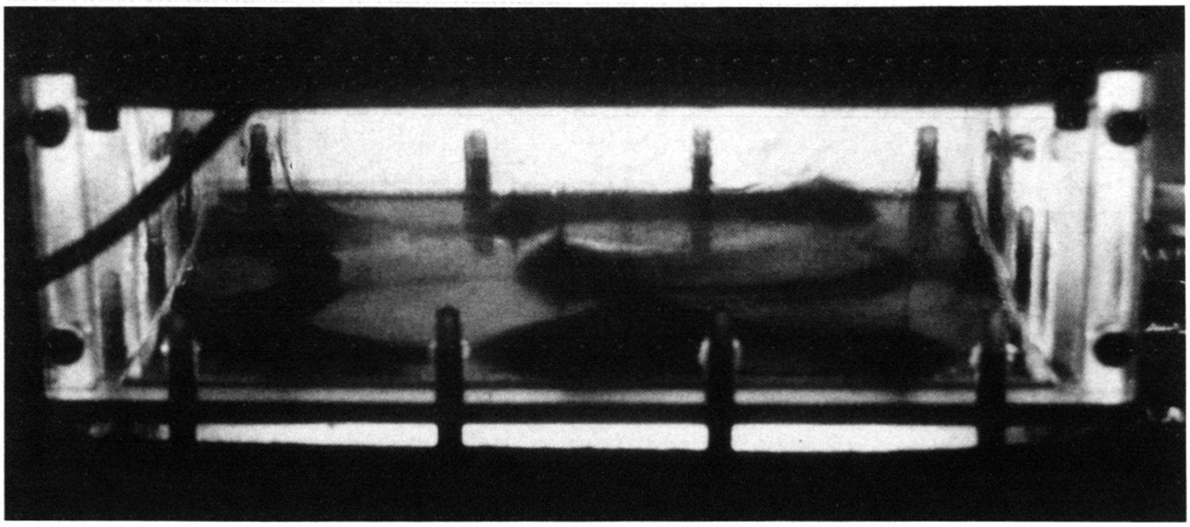

Fig. 7. Convection experiment with corn syrup. The main layer of syrup is transparent. Initially a thin layer of syrup with increased density (due to the addition of $CaBr_2$) was put onto the bottom and made visible by adding a darker dye. Some part of the heating plate is swept clean of the bottom layer, and upon close inspection "trails of smoke" rising from the bottom layer are visible, which indicate the entrainment by thermal plumes.

signature of the plumes (HOFMANN & WHITE 1982). The formation of separate patches of the dense bottom dreg and the entrainment into rising hot plumes is also observed in laboratory experiments using corn syrup (Fig. 7). The entrainment rates could be of the order of 3–10%, although the influence of the entrained material on the trace element and isotopic composition of hot spot basalts could be higher, because it is enriched in the incompatible elements. The dense pools show internal circulation and entrain ambient mantle, thus they become progressively diluted and may finally dissolve. On the other hand, they could become replenished by fresh subducted oceanic crust, and there may exist a dynamical equilibrium between leaking into plumes, dilution by digesting ambient mantle, and replenishment by ongoing crustal subduction. The residence time of segregated crust at the core-mantle boundary could be on the order of one or several billion years. A patchy distribution of anomalous material at the core-mantle boundary is also suggested by the seismic observations of a discontinuity above the core-mantle boundary, which is not a global feature, but present in some regions and absent in others (LAY 1989, WEBER & DAVIES 1990).

From these model calculations and laboratory experiments it seems that the assumption of oceanic crust accumulating at the core-mantle boundary and being entrained into rising plumes can at least not be easily discarded as dynamically implausible. Still, at this stage the evidence to support this hypothesis is very scanty, and further understanding of the dynamics of the processes is needed as well as more seismological data on the D″-layer and better constraints on the physical properties of the various petrological compositions under very high pressure.

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