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Models of Mantle Convection: One or Several Layers

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Models of mantle convection: one or several layers

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[Plate 1]

Numerical model calculations are used to determine if convection in the Earth's mantle could be organized in two or more layers with only limited mass exchange in between.

The seismic discontinuity at 670 km depth and the top of the D″-layer at the bottom of the mantle are considered as candidates for internal boundaries. If the 670 km discontinuity is caused by an isochemical phase transition, it has to have a Clapeyron slope of $\mathrm{d}p/\mathrm{d}T \leqslant -6\ \mathrm{MPa\ k^{-1}}$ to prevent convection currents from crossing; this value is improbably low. If the discontinuity represents a chemical boundary, the intrinsic density difference has to exceed 3% to prevent subducted lithospheric slabs from penetrating deeply into the lower mantle; also the condition is possibly hard to meet. The least improbable mechanism for a mid-mantle barrier for convection currents would be a combination of endothermic phase transition and chemical change. The boundary between upper and lower mantle would show considerable topography, and a limited material exchange is to be expected at any rate.

The possibility of a downward segregation of former oceanic crust, transformed to dense eclogite, is studied in a further model series. It requires a region of low viscosity, as the D″-layer probably is, and is faciliated by the decrease of the thermal expansion coefficient with pressure. About 20% of subducted oceanic crust could accumulate at the core–mantle boundary. The dense material would concentrate underneath rising thermal plumes, and some of it is entrained into the plumes, possibly affecting their geochemical signature.

Introduction

Although convection in the Earth's mantle is accepted as a concept, its detailed structure is poorly understood. An outstanding question is the amount of mass exchange between various parts of the mantle. The idea of separately convecting layers above and below the seismic discontinuity at 670 km depth, as opposed to whole-mantle convection, has been discussed for two decades now, without a consensus being reached. The evidence for seismology (Creager & Jordan 1986; Giardini & Woodhouse 1984), from mineral physics (Jeanloz & Thompson 1983; Bukowinski & Wolf 1988) and from geochemistry (Allègre *et al.* 1983; Davies 1984) is ambigious. In this paper, numerical model results are reviewed, which approach this problem from the viewpoint of the mechanism that might enforce the separation of convecting layers.

Recently it was also discussed whether the D″-layer, the lowermost 200 km of the mantle, is chemically distinct from the overlying material (T. H. Jordan, personal communication). Several sources for the origin of such a dense bottom layer are considered: chemical interaction between core and mantle, a relic from the core-formation process, or downward segregation of dense heterogeneities in the convecting mantle. Hofmann & White (1982) proposed for geochemical reasons that subducted oceanic crust, being transformed into dense eclogite, could accumulate at the bottom of the mantle, being stored for about 1–2 Ga, and then rise again in mantle plumes. In the third section, preliminary numerical model results are presented, which indicate that such a process is indeed feasible.

Layering at the 670 km discontinuity

The transformation of the spinel structure of olivine, $(Mg,Fe)_2SiO_4$ into a mixture of perovskite, $(Mg,Fe)SiO_3$, and magnesiowüstite, $(Mg,Fe)O$, occurs at about the right pressure to explain the second major discontinuity in the mantle. This transition is likely to be endothermic, i.e. it has a negative Clapeyron slope $\mathrm{d}p/\mathrm{d}T < 0$, and it causes an increase of density of the order 6–10%. Some doubt is cast on the interpretation of the 670 km discontinuity as a pure phase boundary by the observation of reflected and converted seismic waves from that depth, implying that the discontinuity is sharp. This is more typically expected for a compositional boundary. A higher magnesium:silicon ratio than in the upper mantle is assumed in most petrological models of a stratified mantle (see Liu 1979). However, the uncertainties in seismic and equation-of-state data are too large to permit a clear decision between uniform or stratified composition. Thus the 670 km discontinuity may represent a pure phase boundary, a compositional boundary, or perhaps a combination of both.

The main influence of a solid–solid transition on convection currents is caused by the deflection of the boundary in hot or cold plumes. For example, in a cold subducted lithospheric slab an exothermic boundary (like the olivine–spinel boundary at 400 km depth) will be elevated to lower pressure because of its positive $\mathrm{d}p/\mathrm{d}T$ slope. The elevated dense phase exerts a strong gravitational force in a downwards direction, which helps to pull down the slab. For an endothermic reaction, a depressed region of the low-density phase occurs, which is buoyant and opposes subduction. If the Clapeyron slope is sufficiently negative, the phase boundary may thus enforce two-layer convection. A further effect of a phase boundary is the release of latent heat. It is important at the critical Rayleigh number (Schubert *et al.* 1975). However, at the high Rayleigh number of mantle convection, its only influence is to modify the adiabat locally. For the question of single- against double-layer convection it is of little consequence (Christensen & Yuen 1985).

Christensen & Yuen (1984, 1985) determined under which conditions a phase boundary would lead to two-layer convection in two-dimensional numerical model experiments. The important parameter is

$$P = \frac{\Delta\rho\gamma}{\rho^2\alpha g h}. \tag{1}$$

Here α is the thermal expansion coefficient, ρ the density, $\Delta\rho$ the density contrast between the two phases, $\gamma = \mathrm{d}p/\mathrm{d}T$ the Clapeyron slope, g the gravity acceleration, and h the height of the mantle. P is a non-dimensional number. Single-layer convection breaks down when P is lower than a certain critical value. Experiments at various values of P indicate that the critical value of P depends slightly on the Rayleigh number Ra (figure 1)

$$P_{\mathrm{crit}} \approx -4.4\, Ra^{-\frac{1}{5}}. \tag{2}$$

Although the circulation pattern was found clearly to be that of two-layer convection for $P < P_{\mathrm{crit}}$, there is still a considerable mass flux across the boundary (figure 1). At a Rayleigh

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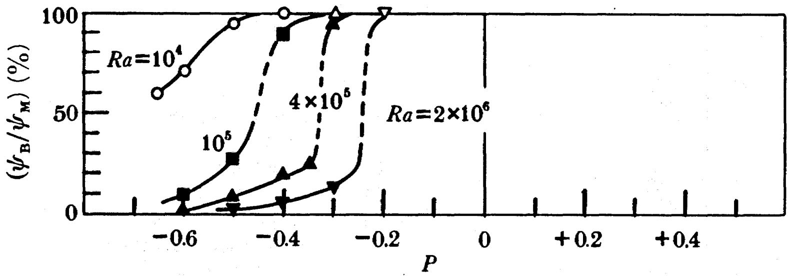


FIGURE 1. Results from convection models in a square box with a phase boundary at mid-depth. The ratio of the stream-function maximum at the phase boundary to the absolute maximum is plotted against the parameter P (equation (1)); it is a measure for the degree of mass exchange between both layers.

number appropriate for present-day mantle convection, say 10^7, $P_{crit} \approx -0.18$. Assuming plausible values for the various parameters that enter into equation (1), one finds that dp/dT has to be less than about -6 MPa K^{-1} for the spinel → perovskite + magnesiowüstite transition to cause layered convection. Preliminary experimental values are only of the order of -2 MPa K^{-1} (Ito & Yamada 1982). Therefore it seems unlikely that the phase change would lead to layered convection unless future experiments would indicate a much lower value of dp/dT. It is interesting to note that for the large icy satellites of the outer planets the story could be different. For the ice-II–ice-V transition in Ganymede, $P = -1.6$, thus the critical value is by far exceeded and two-layered convection must be expected if the temperature is in the stability field of these phases.

Christensen & Yuen (1984) also considered the influence of a chemical or a combined chemical and phase boundary in numerical convection models including subducted lithospheric slabs. The results are summarized in a domain diagram (figure 2) for the style of convection. We estimated the uncertainties in the exact limits of the domains to be on the order of some tens of percent, because of uncertainties in parameter values and unmodelled effects, for example the influence of the driving buoyancy of the olivine–spinel transition. According to our model results, the slab would be stopped at the boundary and bend sideways when the compositional density difference exceeds about 4–5%. The impact of the slab causes a depression of the boundary between upper and lower mantle of 100–200 km. It is not very likely that the compositional density difference can exceed a few percent. When it is less than 4% the slab was found to plunge deeply into the lower mantle and with less than 2% it would reach the core mantle boundary and lead to large-scale mixing of upper- and lower-mantle material. Kincaid & Olson (1987) found, at least semi-quantitatively, the same behaviour in laboratory experiments using corn syrup to model slabs in a stratified convecting system.

The least improbable mechanism for two-layer convection (in a narrow sense) is a combination of a distinctly negative dp/dT of the spinel → perovskite + magnesiowüstite transition and a superimposed compositional change. As can be seen from figure 2, 3% chemical density difference and a Clapeyron slope of -3 MPa K^{-1} would lead to layered convection. Both are still plausible values. Of course, having a chemical change and a phase boundary at the same depth must appear as an improbable coincidence, unless someone can put forward a plausible explanation for the development of such a situation, for example within the scenario of an early magma ocean. The depression of the boundary between upper and lower mantle at the site of slab subduction would be of the order of 100 km. It should be possible to detect such a depression by seismological means. However, one would expect travel-

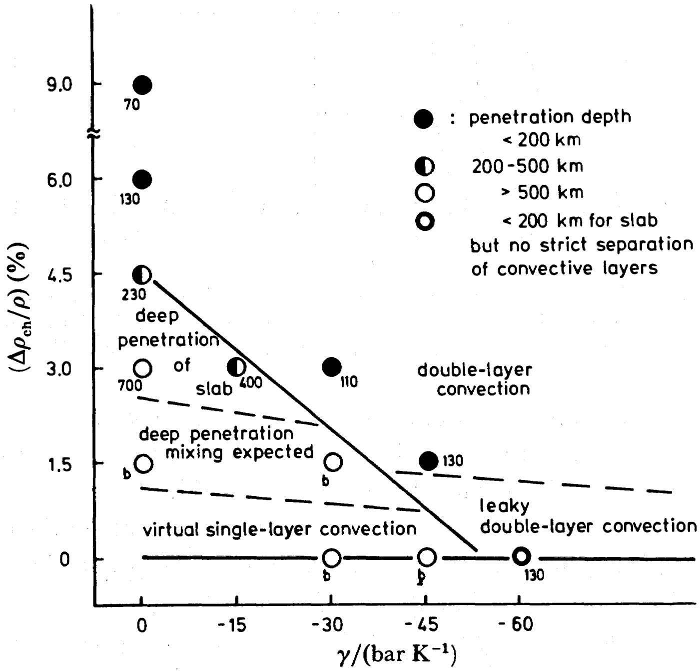


FIGURE 2. Domain diagram for convection including subducting plates. It indicates the mode of convection depending on the chemical contribution to the density differences at the 670 km discontinuity and the p/T-slope of the phase transition contribution. Numbers indicate the penetration depth of slabs below 670 km, 'b' means penetration to the bottom of the mantle. $(\Delta\rho/\rho)_{tot} = (\Delta\rho/\rho)_{ph} + (\Delta\rho/\rho)_{ch} = 9\%$. 1 bar = 10^5 Pa.

time anomalies from deep-focus earthquake that show more or less the opposite sign to what has been found (Creager & Jordan 1986).

In summary, from the point of view of the mechanism for dynamical layering, whole-mantle convection is preferred. On the other hand, the conditions for having separate convection layers above and below the 670 km discontinuity are not so extreme that this possibility could be ruled out entirely. Further evidence could come from mapping the boundary in the region of slab subduction. It should be emphasized that, whatever the mechanism for possible layering, the mass exchange between upper and lower mantle would still be significant.

SUBDUCTED CRUST AND A CHEMICAL D″-LAYER

Recently it has been speculated that the D″-layer at the bottom of the mantle is a chemical boundary layer (T. H. Jordan, personal communication) to explain the strong seismic lateral heterogeneity in this region and the inferred topography of the core–mantle boundary, which is on the order of ± 6 km (Morelli & Dziewonski 1987). Among the various proposed mechanisms for its formation, perhaps the least speculative is the segregation of subducted oceanic crust. At least we know that large amounts of basaltic crust are actually subducted and that it transforms under pressure into eclogite and other dense phases. According to Ringwood (1982) the former crust keeps an excess density of about 5% compared with normal mantle through all the phase transitions that occur on the way to the lower mantle. Recent work (Irifune & Ringwood 1987) indicates that it may be buoyant in a limited depth interval around 670 km; however, if slabs can penetrate below this zone it would be of little importance for the processes considered here. An obstacle for segregation is that the crust lies above a layer of depleted harzburgite, which is compositionally the complement to the crust and less dense than normal mantle by 1–2 %. Both layers combined are about neutrally buoyant (except for thermal buoyancy). Another obstacle is that the crustal layer is thin, about 6–7 km. Stokes-sphere estimates show that a body of eclogite of 5 km diameter would sink only at 0.1 mm a^{-1} for a typical viscosity value of the mantle of 10^{21} Pa s. This is much less than the expected velocity of convection, and mixing rather than segregation would be expected. Model calculations with constant viscosity by Gurnis (1986) indeed show for realistic parameters not much of a settling effect for heavy tracer particles inserted into a convective flow. However, because the viscosity is strongly temperature dependent, localized regions exist in the mantle where the viscosity is much below its mean value, namely the hot lower boundary layer of convection, or hot rising plumes. Here the separation of 'crust' and 'harzburgite' may take place with the former accumulating at the bottom.

A number of numerical model experiments have been performed to determine whether such a process is indeed feasible with realistic parameters. Into time-dependent models of variable viscosity convection, a chemically layered 'lithosphere' is inserted at time zero, consisting of a dense thin crust and a thicker buoyant layer underneath. Both combined are neutrally buoyant. Technically the compositional differences are modelled by statistically distributed tracer particles. To avoid spurious effects, their number must be so large that the influence of a single tracer is negligible and only their statistical density effects the flow. Up to 10000 tracer particles have been used. The effective Rayleigh number is of the order of 3×10^5, about 1 or 2 orders of magnitude lower than what is realistic for whole-mantle convection. This poses the question of how to scale various parameters to get similar behaviour as in the mantle. The two conditions, which are met in the calculations, are that the ratio of chemical to thermal density anomalies is the same as for the Earth

$$\Delta\rho_{\text{chem}}/\rho\alpha\Delta T \approx 1, \tag{3}$$

and that the ratio of the thickness of the chemical boundary layer (the oceanic crust) to the thermal boundary layer (the lithosphere) is correct:

$$\delta_{\text{chem}}/\delta_{\text{therm}} \approx 0.1. \tag{4}$$

A last point is how to translate model time into real time. Here a timescale based on the mean flow velocity is taken. The time that a particle needs to move a distance equal to the depth of the convection layer is called the transit time. For the Earth, one transit time is about 60 Ma. This figure is used to translate the time lapse in the model into geological time.

The standard parameters in the model comprise an exponential decrease of the thermal expansion coefficient with depth by a factor of three (Stacey 1977), a decrease of the viscosity due to its temperature dependence from top to bottom by a factor of 16000, and an increase due to pressure by a factor of 64. This combines into a viscosity difference between top and bottom by a factor of 250. No internal heating is considered and all boundaries are stress-free. A model run with these parameters in a box of aspect ratio 1.5 shows in fact a strong separation effect in the lower boundary layer (figure 3, plate 1). The former harzburgite layer rises in a buoyant chemical diapir, and the former crust accumulates in a dense pool underneath the rising thermal plume. This pool is not entirely crust, it contains also some normal mantle and a little harzburgite and has about half the excess density of the crust. The pool is leaking;

continuously some material is entrained into the thermal plume. However, even after 18 transit times (1 Ga) about half of the original crustal material resides in the pool.

This model experiment is unrealistic in the sense that subduction must occur at the impenetrable sidewall of the box. This guides the lithosphere directly into the hot low-viscosity bottom layer, where it lies upside down, which highly favours the separation effect. However, before considering a more realistic model, the influence of two variations of the standard parameters were studied. When the thermal expansion coefficient is not depth-dependent, but set to a mean value, again a pool of segregated crust forms. However, this time the pool is leaking much more heavily and has lost most of its substance after seven transit times (400 Ma). In another experiment, the viscosity, not the expansion coefficient, was fixed at a mean value. This time no separation at all was found; rather the compositional layers stayed together and were almost passively torn around by the flow. This emphasizes the importance of the local viscosity reduction due to temperature dependence. For example, when a pool has formed, the low viscosity reduces the shear forces of the external flow that act on the pool such as to drag out material.

To avoid the unrealistic geometrical constraint on subduction at a sidewall of the computational box, model runs have been done in a domain of aspect ratio three with two cells (initially), where subduction takes place in the middle. In this configuration the flow is strongly time-dependent due to boundary instabilities and the subducting chemically layered lithospheres does not go straight to the bottom but is first turned sideways, although eventually some of its material enters the lower thermal boundary layer. This time only 5% of the subducted crust was found to form a tiny dense pool under one of the thermal plumes, whereas the rest was scattered throughout the cells. This seems discouraging; however, in the model a crustal layer was only once subducted, whereas on the Earth crustal subduction is a continuous process and new crust could be added to pre-existing dense pools.

The presence of previously segregated material at the bottom seems also to enhance the addition of new crust, although the reason for this is not clear. In two model runs that were started with a compositionally stratified lithosphere and some dense material already at the bottom, about 20–25% of the subducted crust was found to join into the bottom pools. In figure 4, plate 1, it can be seen how part of the subducted crust merges with the pre-existing dense pool.

Further model experiments were concerned with the dynamics rather than with the formation of a chemical bottom boundary layer. They were started by putting dense material right onto the bottom. It was found that the pools (which quickly formed) can wander around and follow the upwelling currents. They can also split up and merge again, analogous to their

Description of plate 1

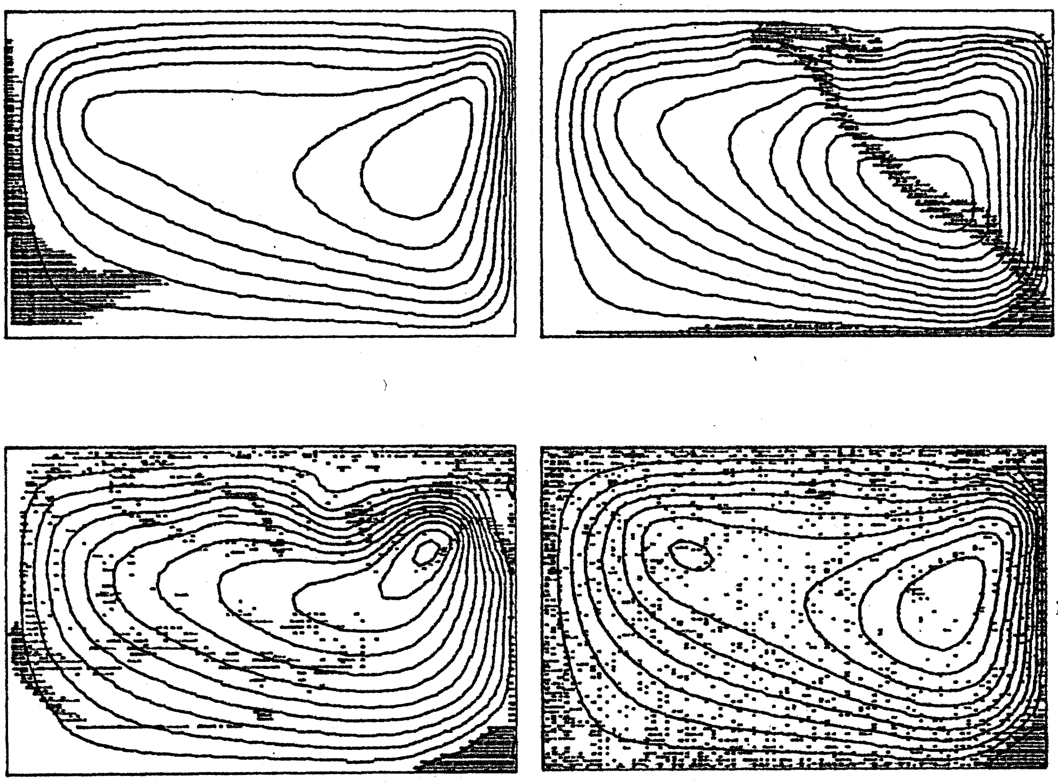
Figure 3. Four snapshots of the temporal evolution of a convection model starting with a chemically stratified top layer with dense crust in red (dark shading) and buoyant harzburgite in blue (faint shading). The instantaneous stream-lines are drawn, numbers at the margin indicate the time lapse in transit time units.

Figure 4. Two instances in time of a convection model with subduction initiated in the center of the model box with a chemically stratified top layer and an already existing dense bottom layer. At transit time 2.5 the subducted former lithosphere is torn and distorted. Some of the former crust merges with the dense pool on the left side. At time 4.1 two distinct pools, sitting underneath rising currents are found; they have consumed 25% of the subducted crust and they leak into the rising plumes.

3

2.1 3.9

5.3 17.7



4

2.1

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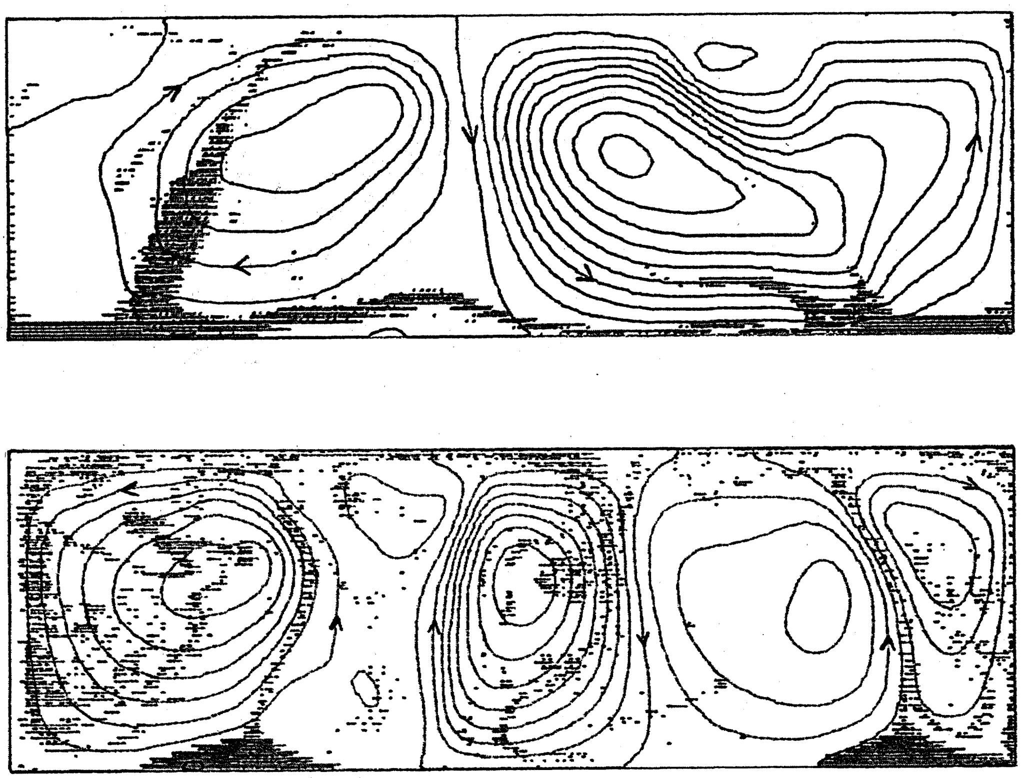

FIGURES 3 AND 4. For description see opposite.

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counterpart at the Earth's surface: the continents. They show internal circulation with the opposite sense of rotation to the external flow. They do not only leak into the rising plumes, but they also entrain ambient mantle. Thus they become progressively diluted and destabilized and finally dissolve rapidly after about 1 Ga. In the real Earth there may, however, be a dynamical equilibrium between dilution by entrainment, loss by leaking into plumes, and replenishment by the addition of new subducted crust. In a few examples the bottom topography has been calculated. As would be expected there is a local low under the dense pools. However, this depression is not larger in amplitude than the variation that is typically produced by thermal convection alone (see also Davies & Gurnis 1986). This can be understood in the following way: the topographic depression is more or less related to the height of the dense pool. Pools form because the dense material is swept laterally by thermal convection currents. How high the material can pile up depends only on the strength of thermal convection; without it the pools would flow flat like pancakes, especially in the low viscosity of the bottom layer. Thus the topographic variation of the bottom depends in an indirect way on the thermally driven flow.

Conclusions

Two kinds of layering in mantle convection have been studied in numerical models, considering the mechanisms for stabilization or formation of such layers. Separate convection in the upper and lower mantle, divided by the 670 km discontinuity, could be caused by a very endothermic nature of the spinel → perovskite + magnesiowüstite transition, by a gross difference in the bulk chemistry of upper and lower mantle, or a combination of both. However, the required parameter values are somewhat extreme and with our present knowledge it is not very likely that the conditions are met. Whole-mantle convection is the preferred mode of convection. This does not exclude the possibility that the mass exchange between upper and lower mantle could be lower than previously assumed, for example because of high viscosity in the lower mantle (Davies 1984).

The model results are more favourable for another kind of hypothetical 'layering', the formation of mantle dregs in the D″-layer, where former subducted oceanic crust could partially accumulate. The model calculations show the great importance of temperature-dependent viscosity for the segregation, and, to a lesser extent, of the decrease of the thermal expansion coefficient with depth. The ancient crust would form local pools underneath rising thermal currents (mantle plumes), which entrain some of the dense material. A sufficiently large amount of dense material would form a continuous cover around the core. Only in this case it would thermally insulate the core to a significant degree, but not if part of the core–mantle boundary were swept clean of it. The model calculations presented here are somewhat preliminary and leave a number of questions. It is necessary to expand the parameter range to define more clearly the conditions for the segregation effect. A number of important points can, in principle, be addressed in numerical calculations, for example the question of the average residence time of material in D″, or if the amount of material entrained into plumes would be large enough to affect significantly the geochemical signature of hot-spot basalts. Entrainment of significant amounts of ancient oceanic crust could explain some, but not all, of the geochemical differences between MORBS (mid-ocean ridge basalts) and hot-spot basalts. The high $^{3}He/^{4}He$ ratios of some oceanic islands warrant another explanation. The influence on core–mantle topography and heat flow, and the effect on seismic travel times could be systematically investigated in numerical models. The present model calculations have indicated that the segregation process can operate and leave it to future work to quantify the details.

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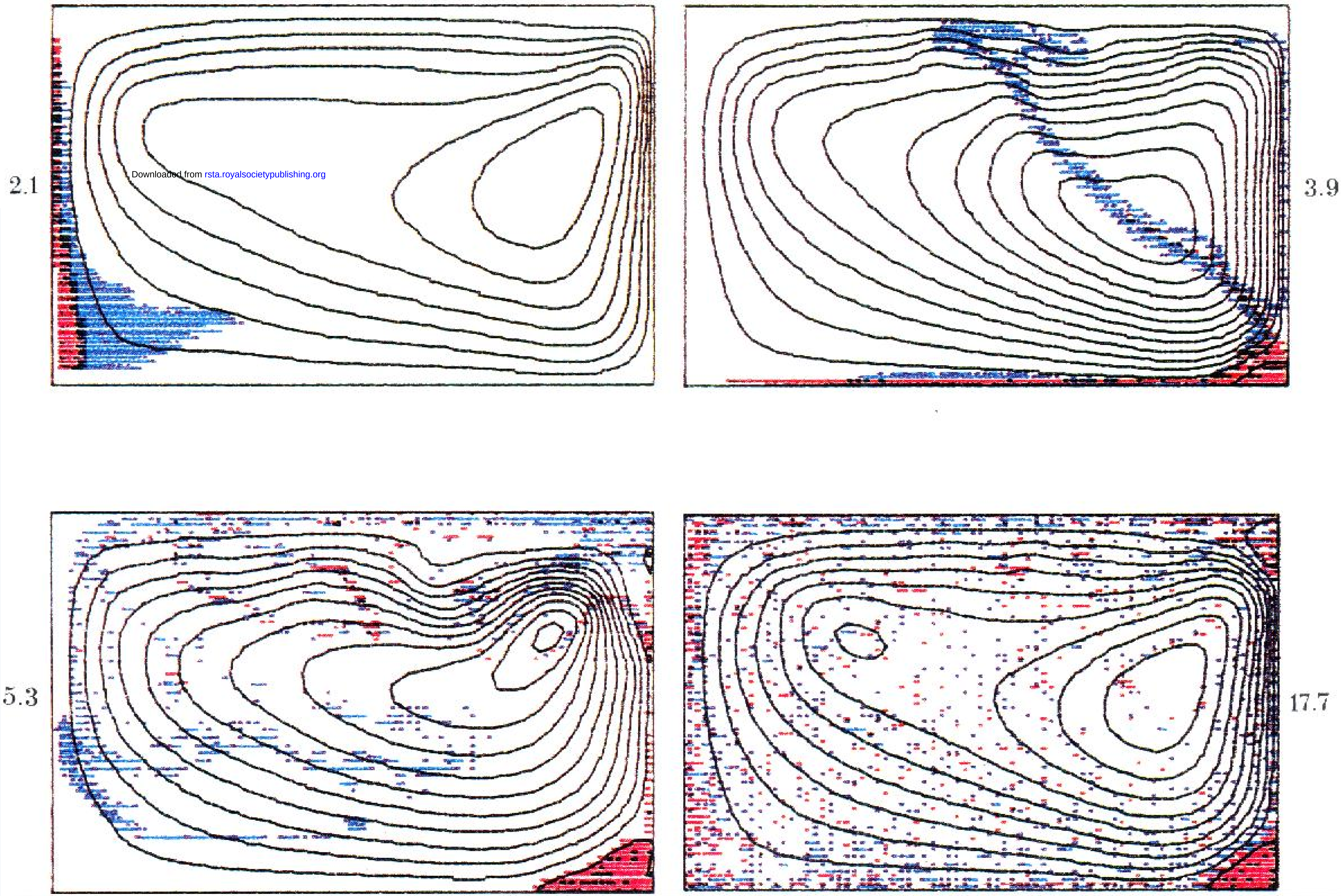

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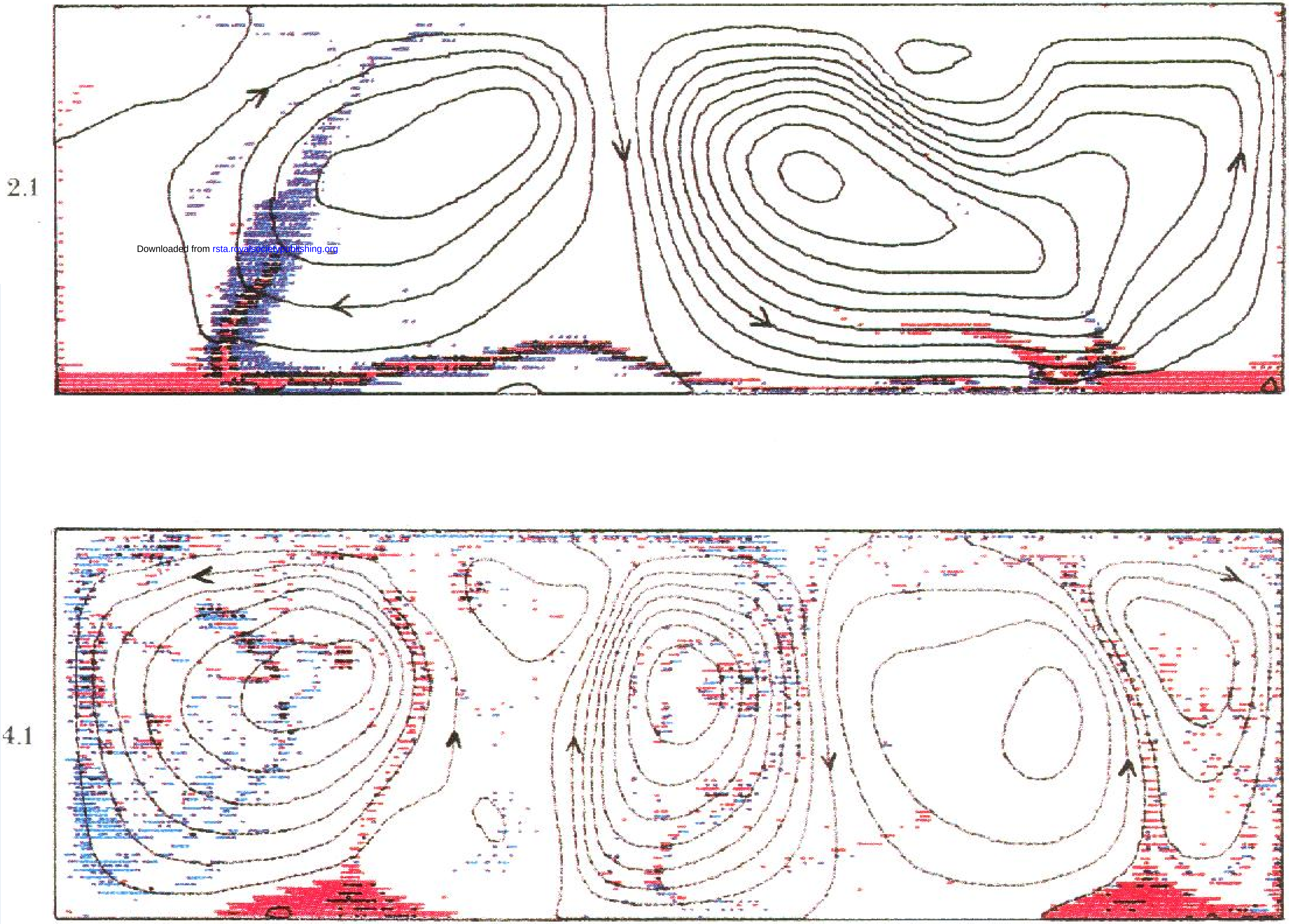

IGURE 4. Two instances in time of a convection model with subduction initiated in the center of the model box with a chemically stratified top layer and an already existing dense bottom layer. At transit time 2.5 the subducted former lithosphere is torn and distorted. Some of the former crust merges with the dense pool on the left side. At time 4.1 two distinct pools, sitting underneath rising currents are found; they have consumed 25% of the subducted crust and they leak into the rising plumes.