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Mantle convection, plates, and hotspots

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ABSTRACT

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More than 20 years after their discovery, plates and hotspots are still awaiting a consistent dynamic model. Both imply large viscosity variations in the mantle. Sustained efforts have been targeted on modeling (on computer as well as in the laboratory) convection in fluids with temperature- and pressure-dependent viscosity. The main results of these studies are reviewed and summarized in a few simple rules. Several properties of plates, such as ridge-push or heat-flow distribution, are well accounted for by convection models with temperature-dependent viscosity. However, the process of subduction is not properly explained by these models. One of the simple rules deduced from the study of convection and temperature-dependent viscosity states that the viscosity drop across the lower boundary layer adjusts to a low value (less than 10). This constrasts strongly with the viscosity variations required by current hotspot formation models, which are typically 100 times larger. It is suggested that this paradox can be solved if subduction is properly included in the convection models.

Introduction

More than 20 years have elapsed since the formulation of "plate tectonics" as a quantitative theory for describing motions at the surface of the Earth (McKenzie and Parker, 1967; Le Pichon, 1968; Morgan, 1968). Plate tectonics became the frame to which all domains of earth sciences refer. Long before that, some had proposed that thermal convection was active in the Earth's mantle, carrying the heat produced by radioactive isotopes up to the surface (Holmes, 1928; Pekeris, 1935). But, here again, only with the advent of plate tectonics did it become possible to achieve a quantitative analysis (Turcotte and Oxburgh, 1967). It was during that same breathtaking episode of the history of earth sciences that "hotspots" were first identified (Wilson, 1963). Meanwhile, Tozer (1972) was showing how important it was, in order to build a realistic convection model of the mantle, how viscosity of the Earth constituents varies with temperature.

The mere existence of the "plates" (defined as "rigid" bodies over a deformable asthenosphere) is an illustration of the existence of a strong temperature-induced viscosity variation in the mantle. In contrast, hotspots appear to be made of material much less viscous than the surrounding mantle they rise through. A better understanding of convection with temperature-dependent viscosity should therefore help us explain these two major tectonic phenomena: plates and hotspots.

The present celebration of the twentieth anniversary of plate tectonics is a good opportunity to try to summarize in simple words the major results that have been obtained on convection with variable viscosity, and see what constraints they put on the Earth's dynamics. Several important characteristics of the plates are found to be well explained by these convection models. However, plate tectonics, with its subducting oceanic lithosphere, appears to be at best atypical of convection with variable viscosity, and remains largely unexplained. The paradox is even worse for hotspots: the common assumption that hotspots originate in a thermal boundary layer with very low viscosity seems to violate the rules formulated from the study of convection with variable viscosity...

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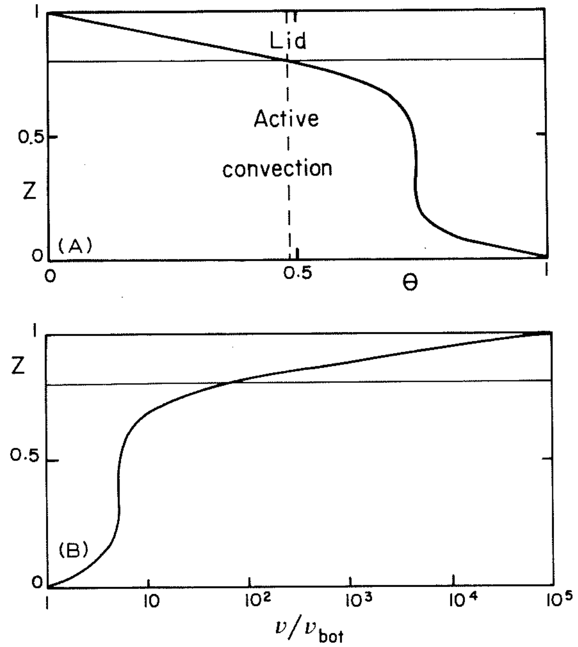

Fig. 1. (A) Vertical profile of the horizontally averaged temperature in a convecting layer with temperature-dependent viscosity. The viscosity at the top (cold) is 10^5 times larger than at the bottom (hot). (B) Vertical profile of viscosity. Most of the viscosity variation is absorbed in the lid that forms at the top, while the viscosity drop across the bottom boundary layer is only a factor of 8. (From Richter et al., 1983.)

The first section of this paper describes the major results of the study of variable viscosity convection, in the context of the preceeding paradox. I then discuss some characteristics of the plates and their dynamics, review major models of hotspots formation, and finally, I propose a few ways of solving our paradox.

Convection with temperature-dependent viscosity

The results of laboratory experiments, and of analytical and numerical modeling of convection in fluids with temperature-dependent viscosity can be summarized in a few simple rules. First, let us consider convection beneath a rigid top boundary (i.e., the "plate" cannot move), the situation which is easiest to model in the laboratory. Let us impose the temperatures at the top and bottom horizontal boundaries. What does the average "geotherm" then look like? Figure 1 shows that it consists of a "lid" at the top, which is very viscous and immobile. Beneath it, a nearly isoviscous fluid layer is actively convecting. Two parameters are needed to describe the "geotherm" (in fact the depth profile of the horizontally averaged temperature): the temperature in the heart of the convecting region, and the temperature gradient in the boundary layers. All available results indicate that the interior temperature is determined from the value of the viscosity drop at the base of the layer. The first simple rule is then: "the interior temperature adjusts so that the viscosity drop at the base of the layer is a factor between 6 and 10" (depending upon the actual viscosity law). This rule can be derived from scaling analysis (Morris and Canright, 1984; Nataf, 1986), and has been checked both numerically (Christensen, 1984, 1985) and in the laboratory (Richter et al., 1983). It has strong implications, since it excludes geotherms such as those proposed by Kenyon and Turcotte (1983), which produce viscosity drops of more than 1000 in the lower boundary layer. We define a convecting sub-layer, as in Fig. 1, so that the temperature profile is symmetrical across it. The viscosity at the top of that sub-layer is used to define its Rayleigh number:

$$Ra = \frac{\alpha g \, \Delta T d^3}{\kappa \nu}$$

where α is the coefficient of thermal expansion, g is the acceleration due to gravity, d is the thickness of the sub-layer, ΔT is the temperature drop across it, κ is the thermal diffusivity, and ν is the kinematic viscosity.

The temperature gradient in the boundary layers is related to the heat transport efficiency of the convecting system. The Nusselt number is the ratio of the actual temperature gradient to the gradient there would be if only conduction was present. The second rule states: "the Nusselt number of the convecting sub-layer obeys the rule of ordinary isoviscous Rayleigh-Bénard convection". Here again, this rule rests upon dimensional analysis (Stengel et al., 1982; Morris and Canright, 1984; Nataf, 1986) and has been confirmed by numerical and laboratory experiments (Booker, 1976; Richter et al., 1983; Christensen, 1984, 1985). In ordinary Rayleigh-Bénard convection, the Nusselt number and the Rayleigh number are related by:

$Nu = aRa^b$

Typical values are $a = 0.185$ and $b = 0.281$. Once the interior temperature has been deduced from the first rule, the remaining unknown is the thickness of the sub-layer. A value is chosen, and the Nusselt number of the sub-layer is calculated from the second rule. The temperature gradient in the top boundary layer of the sub-layer is thus obtained. It should match the gradient in the lid, which also depends on the thickness of the sub-layer. If this is not the case, the process is iterated, until a proper thickness is found. The overall Nusselt is then readily obtained. Other rules have been proposed (Tozer, 1972; Booker, 1976; Booker and Stengel, 1978). Most can be shown to reduce to the preceeding ones (Nataf, 1986). On the other hand, those based on an a priori cut-off temperature at the base of the lid can lead to large overestimation of the viscosity drop at the base of the convecting system, as mentioned by Nataf and Richter (1982) and Nataf (1986).

I will now focus on this question of the viscosity drop at the base on the convecting system. Our first rule states that it is small (of the order of 10). How robust is that assertion? What are the consequences of it? Let us consider one additional ingredient: the dependence of viscosity on pressure. At some point the question arose of whether the temperature profile would then correspond to an isothermal or an isoviscous interior. Fowler (1983) proposed that convection would indeed be active only in a region where viscosity varies little, and that the temperature would adjust to make that region as thick as possible. No laboratory experiments have been carried out using pressure-dependent viscosity, but numerical experiments (Torrance and Turcotte, 1971; Fleitout and Yuen, 1984; Christensen, 1985) all indicate that the interior remains isothermal, even though viscosity may vary by several orders of magnitude across it. The explanation is simple: even where viscosity is large, as soon as the advection of temperature is sufficient to overcome diffusion, the temperature is homogenized. However, one effect of the augmentation of viscosity with depth is to increase the thickness of the lower boundary layer and the associated viscosity drop across it. How large is that increase?

Let us consider two cases: whole-mantle convection, and layered convection with the upper mantle convecting separately above the lower mantle. Note that, because of compressibility, the temperature rises with depth, even in a constant viscosity mantle. Our analysis still applies if one considers the "potential temperature" instead of the actual temperature. In the Boussinesq approximation, the potential temperature is the actual temperature minus the temperature of the adiabatic system. An "isothermal" profile in the lab becomes an "adiabatic" profile in the Earth.

In the case of whole-mantle convection, the temperature profile should be adiabatic, except in the lithosphere and at the core–mantle boundary (CMB). According to our first rule, the viscosity drop at the base of the mantle should be of the order of 10. The pressure-dependence of viscosity could be large in the upper mantle (about 10^3), but this has little effect on the temperature profile near the base of the mantle. In the lower mantle, the pressure-dependence cannot be predicted directly from material properties measured in the laboratory. However, the variation of the activation volume for creep with depth has been investigated. Because of the compressibility of the lattice, the pressure-dependence of viscosity decreases as depth increases. Poirier and Liebermann (1984) have conducted a detailed analysis, based on thermodynamic theory, using in situ elastic properties derived from seismology. They found that an adiabatic interior leads to a variation in viscosity with depth no greater than a factor of 100 across the lower mantle. In other words, the viscosity variation due to pressure is about a factor of 100, which does not much affect the temperature profile. All the viscosity variation is absorbed in the lithosphere. The viscosity drop across the lower boundary layer (at the CMB) should be small (less than 100, when taking into account the weak pressure dependence of viscosity).

The second case is layered convection. We will not consider here the details of the coupling between the two layers (see Cserepes and Rabinowicz, 1985; Nataf et al., 1988), but concentrate on the gross features of the geotherm. In the upper mantle, the viscosity variation due to temperature is very large. It can be described in terms of an Arrhenius law:

$$\nu(T) = A \exp\left[\frac{E}{R}\left(\frac{1}{T} - \frac{1}{T_0}\right)\right]$$

where the activation energy E is about 420 kJ/mole (we ignore the pressure effect for the moment, and assume a Newtonian rheology). A temperature increase from 900 °C to 1000 °C lowers the viscosity by a factor of 22. Here again, most of the viscosity variation is absorbed in the lithosphere. A remaining viscosity drop of about a factor of 10 is expected at the base of the upper mantle from our first rule. The increase of viscosity with pressure slightly modifies this prediction; for a reasonable value of the activation volume, Fleitout and Yuen (1984) obtain a viscosity drop of 100, and a corresponding temperature increase of 300 °C across the lower boundary layer of the upper mantle. What are the characteristics of the geotherm in the lower mantle? From our analysis, one could expect a thick viscous lid to develop at the top of the lower mantle, analogous to the lithosphere beneath the surface (Spohn and Schubert, 1982). This is actually not necessarily the case because the overall viscosity variation due to temperature is not very large across the lower mantle, if it is still described by the preceeding Arrhenius law. Indeed, when the mean temperature is 2000 °C, a 100 °C temperature increase now lowers the viscosity by a factor of 2.6 only (of course this statement needs revision if the activation energy of the lower mantle material is larger). The total variation would be less than a factor of 1000, a value which is too small for a well-defined lid to form. We have seen that the viscosity increase caused by pressure was also moderate in the lower mantle. The two effects partly cancel out, so that convection in the lower mantle would look almost isoviscous. Again, we would predict a small viscosity drop at the base of the lower mantle.

I have deliberately left aside several important issues, such as the actual temperature in the lower mantle and at the CMB (see Jeanloz and Morris, 1986), or the validity of the layered model versus the whole mantle convection model (see Davies, 1984; Silver et al., 1988; Nataf, 1989), in order to concentrate on one single issue: the value of the viscosity drop in the lower thermal boundary layers.

The conclusion of the above analysis is that this viscosity drop should be small (less than 100), regardless of the type of layering.

Plates

What we called the "lid" in the previous section of course ressembles the "lithosphere" a lot. It would perfectly describe the lithosphere in one-plate planets, such as the Moon, Mars, or Mercury. However, the Earth is not a one-plate planet. Plates move with different velocities at the surface of the Earth. They do so because some of them can subduct back into the mantle. Can we predict such behaviour from the analysis of convection with temperature-dependent viscosity? Certainly not with what we have seen above since we discussed results obtained with a *rigid* upper surface. If we now relax this restriction, we can consider the case of convection beneath a free upper surface. No laboratory results are yet available for this but numerical experiments have been performed, in particular by Christensen (1984, 1985). What happens when the upper surface is made free? Not much if the viscosity of the fluid at the top is so large that it plays the role of the now removed rigid boundary. The velocity of the lid essentially depends on the Rayleigh number of the layer, defined using the value of the viscosity *at the top*, as first noted by Daly (1978), and confirmed by Christensen (1984). What makes the lid move? Is it the shear induced by the underlaying circulation at the base of the lid, or rather the horizontal density gradient within the lid? This gradient is due to the slope of the isotherms: high above an uprising current, low above a downwelling. It was shown by Fowler (1985) for a particular case, and by Nataf (1986) for a more general one, that the latter effect is always the dominant one. An essential feature of plate dynamics is therefore obtained form our simple model of convection with temperature-dependent viscosity beneath a free upper surface. The importance of "ridge-push" as a driving mechanism for plate tectonics is now well established (Forsyth and Uyeda, 1975; Richardson et al., 1979; Vigny et al., this issue). "Slab-pull" is of course also important, even though it is largely exhausted in maintaining the descent of the slab (Richter, 1977).

Even if the velocity of the lid is much less than the convective velocities beneath it, its contribution to the heat flux at the surface can be much larger than what would be conducted across it if it were immobile. Using arguments introduced by Morris (1983), Nataf (1986) shows that the Nusselt number above which the motion of the lid is the dominant heat exchange mechanism can be written as:

$$Nu_{50/50} = 10 \frac{\sqrt{\nu_{top}/\nu_{bot}}}{\left[\ln(\nu_{top}/\nu_{bot})\right]^{5/2}}$$

where ν_{top} and ν_{bot} are the viscosities at the top and bottom of the fluid layer, respectively. Even for a total viscosity ratio as large as 250,000, a Nusselt number of 10 is sufficient for reaching this fifty-fifty partition.

Here again, a fundamental property of plates is explained. It is indeed well known that most of the heat gets out of the Earth through the cooling of the drifting oceanic plates (e.g., Sclater and Francheteau, 1970). A word of caution however; this property has been used by some to classify the importance of convective processes in the mantle. For example, Davies (1988b) states that "the plate-scale mode is dominant, plumes are secondary, and other modes are minor". Not only is this ranking rather subjective, but is can also be somewhat misleading: our approach shows for example that convective velocities beneath the plate could be larger than plate velocities. Also, the interior temperature is largely controlled by sublithospheric convection.

This last statement raises an important question: what are the thermal consequences of the penetration of the cold lid into the convective layer beneath it? It is obvious that the cold sinking lid will lower the temperature of the layer. If the bottom temperature is fixed, this will translate into a decrease in the interior temperature. The temperature profile will become more symmetric, and the temperature jump at the base will increase. The question is: by how much?

For a given viscosity ratio, the higher the Nusselt number, the thinner the lid, and therefore the easier the motion of the lid. Morris (1983) estimates that a symmetric temperature profile is reached when the Nusselt number is of the order of:

$$Nu_{sym} = \frac{\nu_{top}/\nu_{bot}}{\ln(\nu_{top}/\nu_{bot})}$$

For a total viscosity ratio of 250 000, the required Nusselt number would be 20 000! If one assumes a more reasonable value for the Nusselt number, say 30 (see Jeanloz and Morris, 1986), a symmetric temperature profile is obtained only if the total viscosity ratio is less than 150 across the layer. The viscosity drop at the base of the layer would then only be the square root of 150, i.e. 13 (assuming that viscosity is an exponential function of temperature).

The viscosity drop at the base of the layer is probably the best indicator of the thermal influence of the sinking cold lid. Christensen (1984, 1985) has performed a very systematic study of convection with various mechanical boundary conditions, and in particular with a free upper surface. In Fig. 2, his result have been used to illustrate the evolution of the viscosity drop at the base of the layer, as a function of the vigor of convection, for different values of the total viscosity variation (no pressure-dependance is considered here). As we have seen in the first section, the viscosity drop is constant and equal to about 6 when both the upper and lower boundaries are rigid (R/R). If the lower boundary is free but the upper one rigid (R/F), the viscosity drop is even lower, and decreases when the vigor of convection is increased, as predicted by Morris and Canright (1984). The interesting case is when the two boundaries are free (F/F). The striking result is that for all the computations performed by Christensen, the viscosity drop at the base never gets above a factor of 8, even though total viscosity ratios as large as 250 000 are considered. For a given Ra/Ra_c, the larger the viscosity ratio, the larger the potential for building a large viscosity drop at the base, but also the smaller the Nusselt

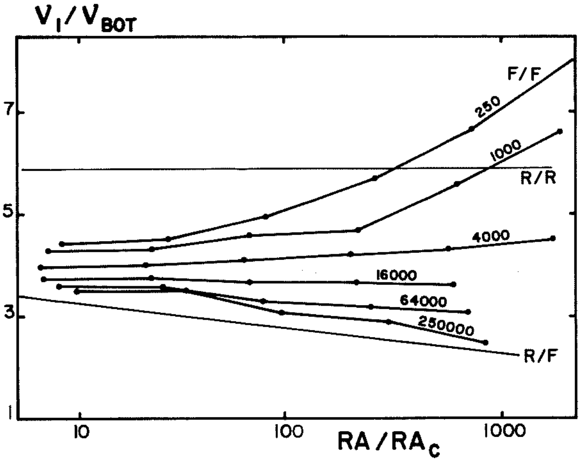

Fig. 2. Viscosity variation across the bottom boundary layer as a function of the vigor of convection for various mechanical boundary conditions. R/R = rigid boundaries at the top and bottom: the viscosity drop remains constant and equal to about 6 (see our first rule); R/F = rigid top and free bottom: the viscosity drop decreases when the vigor of convection increases; F/F = free boundaries at the top and bottom. The six curves correspond to different values of the imposed viscosity variation across the entire layer. For large values (250 000), the lid moves little, and the behaviour is that of the R/F case. For lower values (250), the cold lid is circulated into the layer, and lowers the interior temperature, thereby increasing the viscosity drop across the bottom boundary layer. Note that it remains however smaller than 8 in all cases shown. The curves were calculated from the results of the two-dimensional numerical experiments of Christensen (1984).

number, and therefore the smaller the participation of the lid. It is therefore reasonable to consider that, for a *given Nusselt number*, the viscosity drop at the bottom cannot be larger than a certain value, whatever the total viscosity variation across the layer. The results discussed above suggests that for a Nusselt number of 30 the viscosity drop at the base would never be larger than a factor of about 20.

What we have seen so far is that several fundamental properties of the lithospheric plates are well explained in the frame of convection with temperature-dependent viscosity. In particular, the main driving force was identified as the "ridge-push" (i.e., the force due to lateral temperature variations within the lid), and it was found that the heat flux is primarily due to the cooling of the plate away from the ridge, in agreement with observations. However, we never got close to explaining a very important feature of the plates: the fundamental asymmetry of the subduction process. The cold oceanic lithosphere subducts entirely beneath another lithosphere, which remains at the surface. Such a process, just like transform faults, cannot be explained in the frame of a Newtonian viscous rheology. No model has been proposed where asymmetric subduction spontaneously develops in a convective system. The paradox is then that although we are able to explain several important features of the plates no model incorporates the phenomenon of subduction in a proper way. The symmetric sinking mode observed in convection models might, however, be important in intracontinental deformation. It is probably responsible for cold mountain roots in collision chains (Fleitout and Froidevaux, 1982). An example is the root of the Transverse Range (Humphreys et al., 1984). However, these chains only seem to develop after a subduction-driven convergence phase. In the last section we will discuss what could be the thermal consequences of the asymmetric subduction process.

Hotspots

Wilson (1963) had shown that island chains, such as the Hawaiian–Emperor chain, could be viewed as the tracks left on the moving plate by a fixed source beneath it. Morgan (1971) suggested that the fixed source could in fact be a thermal plume originating in a thermal boundary layer at the base of the mantle. This suggestion is te basis of all recent models. Everybody agrees that the plume should originate in a thermal boundary layer. Which one? This remains an open question: some prefer the hypothetical boundary layer at the base of the upper mantle to Morgan's original suggestion. I will now briefly review the main properties of hotspots that are least questionable.

The duration of most hotspots is at least 100 Ma. Their relative velocities are much smaller than 1 cm/a. Their spacing can be as low as 500 km (Fleitout and Moriceau, 1990). They often tap "primitive" material, i.e. material that has never been processed in geochemical cycles at the surface (e.g., DePaolo and Wasserburg, 1976). They occur beneath oceans and continents. Their conduit in the mantle has never been detected by seismologists. Finally, they seem to be made of material whose viscosity is several orders of magnitude lower than that of the surrounding mantle. Direct evidence for this last statement is provided by the existence of channels that seem to link some hotspots to a neighbouring mid-ocean ridge (Morgan, 1978; Schilling, 1985).

All these properties taken together suggest that hotspots originate in a thermal boundary layer deep in the Earth, and rise rapidly through the mantle in rather narrow conduits.

It quickly became apparent that temperature-dependent viscosity was an essential ingredient for explaining such plumes (Yuen and Schubert, 1976). More recently, "cavity plumes" have been shown to present several attractive features (Olson and Singer, 1985; Loper, this issue). Such a plume is shown in Fig. 3. It is characterized by a large spherical head trailing a very narrow conduit, in which fluid rises very rapidly. After the head has reached the surface, the hotspot lives on fluid that rises through the already "opened" conduit. In the case of the La Reunion hotspot, it has been proposed that the very large volume of the head was responsible for the huge Deccan Traps eruptions, 65 Ma ago (Courtillot et al., 1986). The conduits of cavity plumes develop instabilities, when submitted to horizontal shear. These instabilities can help to explain the apparent fixed nature of hotspots, even if they rise through an actively convecting mantle (Olson and Singer, 1985).

If plumes are a consequence of temperature-dependent viscosity, how do they fit in the general description of convection that we have presented above?

In a recent paper, Sleep et al. (1988) emphasize that plume formation is possible only if the viscosity of the plume's material is at least two orders of magnitude less than their surroundings. For cavity plumes to form, the viscosity ratio must be even larger (Olson and Singer, 1985), probably as large as three orders of magnitude (Loper, this issue). Considering our lengthy discussion of the viscosity drop at the base of the convecting layer, we are faced with the following paradox. What we know of temperature-dependent viscosity convection indicates that the viscosity drop across the lower thermal boundary layer should never be larger than a factor of 100, whereas plume models require a viscosity contrast from 100 to 10^4 to form.

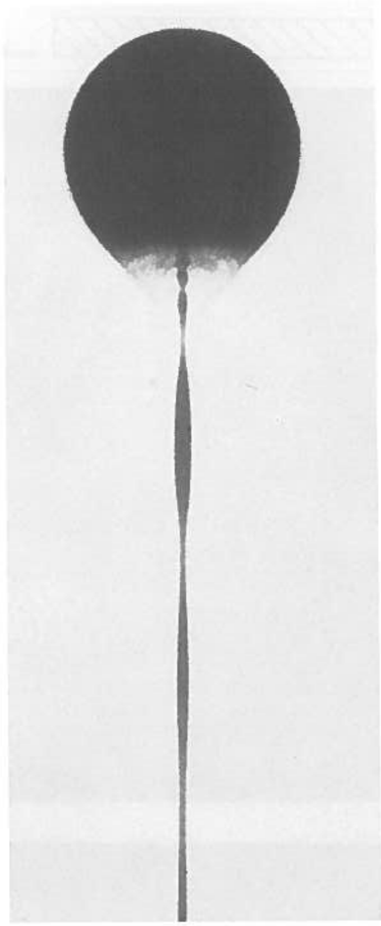

Fig. 3. A cavity plume from Peter Olson's laboratory. Low viscosity low density fluid (in black) rises through high viscosity high density material (transparent). The cavity plume is characterized by a spherical head, which contains most of the fluid, fed by a narrow conduit, in which fluid rises very quickly. The viscosity contrast required for these plumes to form is of the order of 1000. (Reproduced from Olson and Singer (1985), by courtesy of the authors.)

A more concrete demonstration of the above paradox is given by Loper's experiments. Loper (this issue) wanted to study cavity plumes originating at the interface between a low viscosity–low density fluid below, and a high viscosity–high density fluid above. This could never work because the two fluids underwent rapid long wavelength Rayleigh–Taylor instability. The very clever

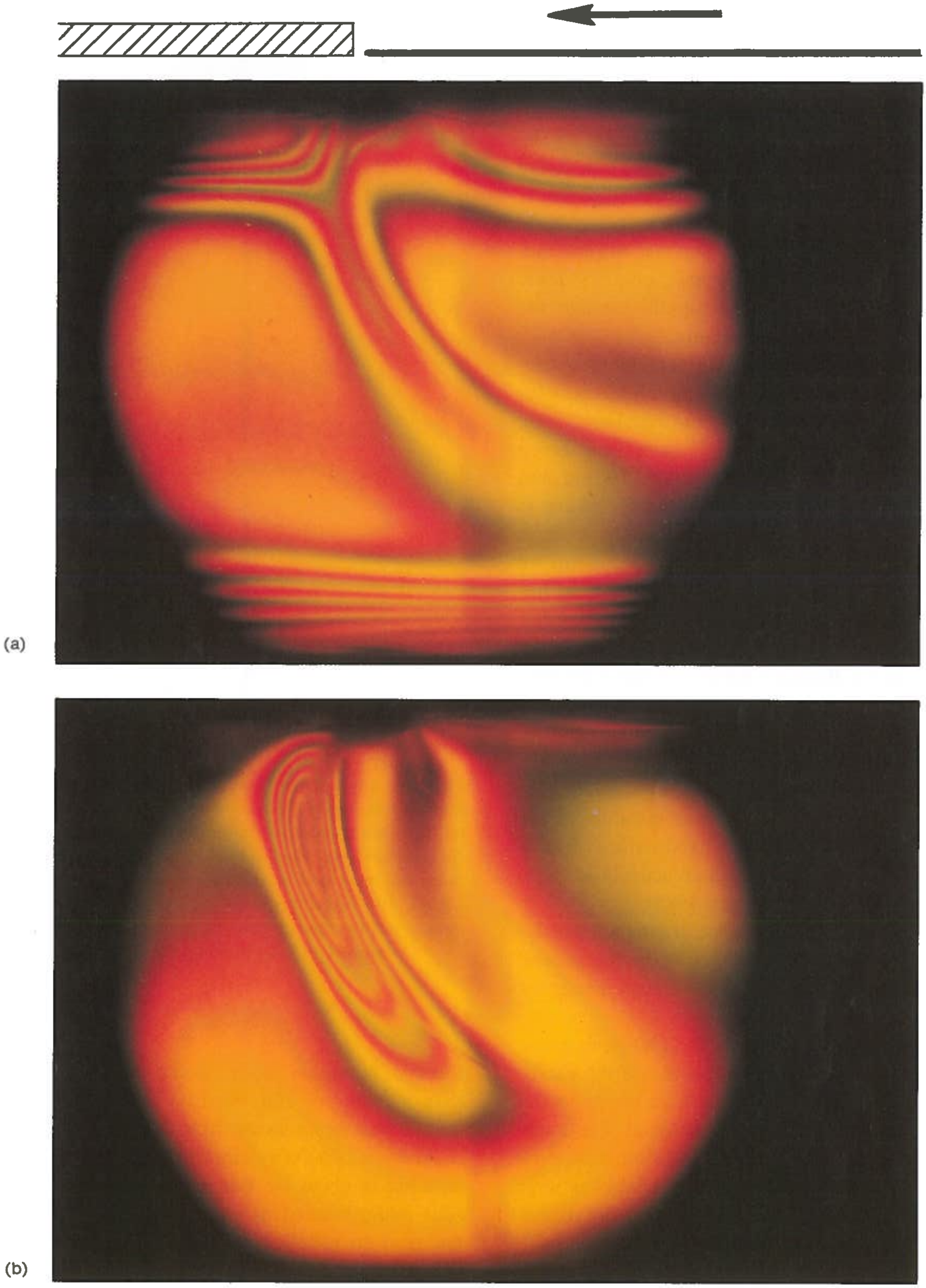

Fig. 4. (a) Vertical and (b) horizontal gradients of temperature in a subducting "slab". Fringes are lines of equal gradient, produced by a differential interferometer. The fluid is Polybutene Oil. Its viscosity strongly depends on temperature. The arrow indicates the sense of motion of the moving lid.

idea of Loper was then to inhibit this instability by placing a silk membrane at the interface. In this way he was able to obtain individual rising cavity plumes. What our analysis confirms is that the long-wavelength instability is the essence of temperature-dependent convection.

Discussion and conclusion

Linked in the history of earth sciences, hotspots and plates are also linked by the fact that both are violating some of the simple rules deduced from temperature-dependent viscosity convection. Could it be that the asymmetric subduction process, unexplained by solely temperature-dependent viscosity models, is capable of building a large viscosity contrast across the lower boundary layer, thus resolving our paradox for hotspot formation? One of the consequences of the symmetric subduction process is that the entire cold lithosphere sinks into the mantle at "full speed". Its ability for cooling the interior of the mantle is thus enhanced and the viscosity drop at the base of the layer is increased. Luce Fleitout (pers. commun., 1988) recently proposed that the cold subducted plate spreading at the bottom of the convective layer could inhibit the Rayleigh–Taylor overturn of the lower thermal boundary layer and, in a sense, play the role of Loper's silk membrane.

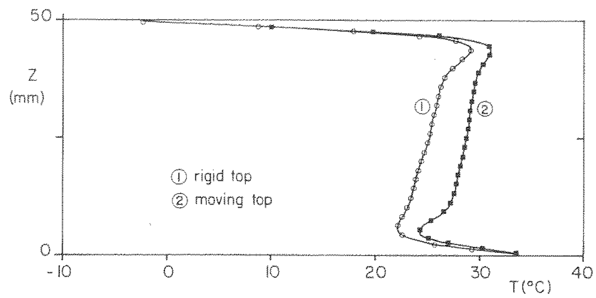

Fig. 5. Preliminary vertical profiles of horizontally averaged temperature. The overall viscosity variation is 100. Profile 1 is with a fixed rigid top boundary. Profile 2 was obtained with a moving top boundary that forced subduction. The velocity of the lid is approximately five times less than the maximum convective velocity. Note the lower temperature above the bottom boundary in profile 2. The shift towards high temperature at mid-depth for profile 2 is a bias of the averaging procedure. It disappears for larger velocities and/or larger viscosity ratios.

A link between subduction and hotspots has been advocated by several authors (Chase, 1979; Anderson, 1982). It is usually assumed that the cold slabs hamper plume formation (Richards et al., 1988), but our thesis, on the contrary, is that the presence of the cold slabs is vital for the formation of hotspots.

Even though asymmetric plate subduction may be difficult to model, it would be interesting to take this as a fact, and to examine the thermal consequences it produces. This approach has been taken by Jacoby and Schmeling (1981) and Davies (1988a), among others. Unfortunately, the total viscosity ratio in these computations was probably too small to build a large enough viscosity variation across the bottom boundary layer. This problem is currently being studied experimentally in our laboratory. Figure 4 shows a subducting slab in one of the experiments. We used a fluid with highly temperature-dependent viscosity. We entrained the viscous "lid" at the surface at a constant velocity. An obstacle forced it to subduct at a given position. More details concerning the experimental set-up will be presented elsewhere. Preliminary temperature profiles are given in Fig. 5. The profile with forced subduction is compared to the profile with a rigid top. The effect of subduction is to decrease the temperature above the lower boundary, thus forming a sort of cold horizontal tongue. This results in a larger viscosity contrast across the lower boundary layer, as predicted in the above analysis. In this preliminary experiment, the overall viscosity variation is only 100. It is too small for narrow plumes to form. Experiments with viscosity ratios up to 10^5 are now being conducted in our laboratory.

Are there other ways to solve the hotspot paradox? One possibility is to obtain a larger viscosity drop in the lower thermal boundary layer. An increase of viscosity with pressure in the lower mantle by three or four orders of magnitude could do it. The presence of very viscous material in the transition zone (Claude Allegre, pers. commun., 1988; Ricard et al., 1989) would also help if hotspots originate in the upper mantle. Also, since the mantle is not in a thermal steady-state (McKenzie and Weiss, 1975; Jeanloz and Morris, 1986), the lower boundary layer might not have had time to lower the viscosity of the interior of the mantle. A final alternative consists in hotspot models that do not require a very large viscosity contrast to form. Bercovici et al. (1989) argue that spherical geometry can help in obtaining stable plumes, while Rabinowicz et al. (1989) show that a low-viscosity zone beneath the lithosphere stabilizes such plumes with respect to plate motion.

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