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The mantle dynamical repertoire: plates, plumes, overturns and tectonic evolution

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In the present mantle, plates and plumes are the main active components. However, calculations of the thermal evolution of the mantle based on purely thermal convection with plates and plumes yield very smooth changes, which are difficult to reconcile with the apparently episodic accumulation of continental crust and with observed changes in tectonic style through Earth history. The effects of composition on the density of the lithosphere, both at the surface and after subduction, can change the dynamics of plates and mantle convection, resulting in a repertoire of mantle behaviour that may better explain the observed tectonic evolution. Thus major mantle overturns may have occurred during the Archaean and possibly the Proterozoic, with dramatic tectonic and magmatic effects, and plate tectonics may not have worked in its modern form when the mantle was hotter. Plumes are not an alternative to plates: they come from a different thermal boundary layer. Plumes have probably played a significant but usually secondary role throughout Earth history.

Introduction

The Earth's tectonic regime is governed by the way the mantle gets rid of its internal heat. Currently, this is by plate tectonics, which is a form of thermal convection: the plates and their associated 'plate-scale' flow are driven by the negative thermal buoyancy of the plates, and the cycle of plate formation at a spreading centre, cooling, subduction and reheating in the mantle accounts for most of the heat lost from the mantle. Thus the plate-scale flow is the dominant form of mantle convection at present.

The density of the plates, and hence their 'negative buoyancy', is affected by composition as well as by temperature. There are two ways in particular in which this might affect the dynamics of the plates and the mantle. One is through the lower density of the oceanic crust, which makes plates positively buoyant until they are about 15 m.y. old. The other is through effects on phase transformations in the mantle transition zone, which might enhance or inhibit the penetration of mantle convection through the transition zone.

The result of the first effect may be that plate tectonics, in its modern form, has not always been viable. The lithosphere would then have had to behave in a different way, which would result in a different form of tectonics. The second effect may have caused the mantle to become layered, either intermittently or for long periods. The breakdown of such layering, if it then occurred, would probably have had dramatic tectonic and magmatic consequences.

The role of plumes may also have changed, though perhaps not dramatically. However if the mantle was ever layered, then plume heads would have been smaller and plumes probably more numerous.

Plumes cannot be a substitute for plate tectonics, because plumes arise from a lower thermal boundary layer, whereas plates involve the upper thermal boundary layer. A change in the way the upper thermal boundary layer operates does not imply any change in the way plumes operate. If plate tectonics is not available as a way to remove heat from the mantle, then the upper thermal boundary layer will have to find another behaviour that does.

The present mantle

Robust inferences about mantle convection can be made from well-established observations, principally the large-scale topography of the sea floor and heat flow through the sea floor. Here I will simply summarise the main points of the arguments about the present mantle; the details can be found elsewhere (Davies & Richards 1992). The seafloor heat flow is well explained as due to cooling and thickening of the oceanic lithosphere by conduction (Fig. 1). This heat comes from the passive upwelling of hot mantle under spreading centres. The cooled lithosphere is eventually subducted, and ultimately reheated from the internal heat of the mantle, which is renewed by radioactivity. The net result of this process of formation of lithosphere, cooling, subduction and reheating is that heat is removed from the mantle. The ultimate source of this motion is the negative buoyancy of the cooled lithosphere (forces such as 'slab pull' and 'ridge push' exist because the lithosphere is cold and dense). The process is therefore a form of convection; in other words, heat is transported by fluid motions that are driven by buoyancy.

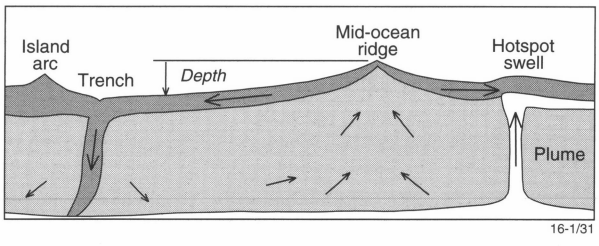

Figure 1. Sketch of the present dynamical system of the mantle, as deduced from topography and heat flow observations (Davies & Richards 1992). The active components are plates and plumes. The negative buoyancy of plates drives a large-scale flow that involves a passive return flow under mid-ocean ridges. Plumes are driven independently by the buoyancy of material from a hot thermal boundary layer at the base of the mantle. Each active component generates characteristic topography. The mid-ocean ridge topography is due to the cooling and thermal contraction of the plates. Hotspot swells are due to the buoyancy of plume material arriving under the lithosphere. The plates and the plate-scale flow are dominant in terms of heat flow, mass flow and topography, and the plumes are secondary, corresponding to about 10% of the Earth's heat budget, which is consistent with them coming from the base of the mantle rather than from the transition zone.

The plates strongly control the structure of mantle convection, since upwelling clearly must occur under spreading centres and subduction zones are the dominant source of active downwelling: the strength of the plates inhibits 'dripping' off their lower surface away from subduction zones. If there were a significant amount of such dripping, it would be evident through widespread topographic and gravity lows, and no such signals are evident.

The depth of the sea floor increases in proportion to the square root of its age (Fig. 1), and this can be explained by thermal contraction due to conductive cooling of the lithosphere. The observed decline of heat flux in inverse proportion to the square root of seafloor age supports this model. A clear implication is that mantle upwelling under normal mid-ocean ridges is passive. If there were active, buoyant upwellings under ridges, the ridge crest would be higher, and the sea

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floor would subside more steeply as it migrated off the buoyant upwelling. (This is observed where there are plumes rising under ridge crests, but elsewhere it is not true.) The implication is that the upwelling under 'normal' ridges is a passive 'return flow' complementing the active downwelling at subduction zones.

Hotspot swells are the only clearly identified manifestation of active, buoyant upwelling (Fig. 1). The size of the swells reflects the amount of buoyant mantle maintaining them. Taking account of the rates of plate motion over plumes, it is possible to calculate the flux of heat transported by plumes, and this turns out to be less than 10% of the Earth's heat budget (Davies 1988, Sleep 1990). This estimate is not dependent on assumptions about the temperature of plumes or the viscosity of the mantle. Also, it is consistent with the amount of heat expected to be conducting from the core into the base of the mantle. This heat would generate the hot thermal boundary layer that gives rise to plumes.

The topography of the sea floor provides constraints also on the possible layering of the mantle. If the mantle were layered at present, then there should be much stronger buoyant upwellings reaching the base of the lithosphere, and these would generate topography rivalling the mid-ocean ridge system in scale and magnitude. This argument proceeds as follows. There is very little radioactivity in the upper mantle (Jochum et al. 1986), so most of the heat emerging at the surface must come from below the upper mantle. Cooling of the upper mantle can account for only a small fraction of the total. Therefore, if there is a barrier to flow in the mantle transition zone, most of the Earth's heat budget must conduct through it. This will generate a hot thermal boundary layer at the base of the upper mantle, and from this will arise hot upwellings. When these reach the base of the lithosphere, they will raise it and so generate topography. There is a proportionality between heat flux, buoyancy flux and topography, so the uplifts would be comparable to the mid-ocean ridges (which are the expression of the same heat *leaving* the upper mantle).

The mantle may be close to the condition where flow through the transition zone is blocked. Seismic tomography seems to show some subducted slabs penetrating the transition zone and others not (Grand 1994, van der Hilst et al. 1991, 1997), although some of the slab contortions can be explained by local trench motions (van der Hilst 1995, van der Hilst & Seno 1993). Numerical models show that old, stiff slabs and large plume heads can probably penetrate a phase transition barrier, but that young, thin slabs and some plume tails may not (Davies 1995a).

The picture of the present mantle that emerges from these arguments is that there is a dominant 'plate-scale' mode of mantle convection, a secondary plume mode driven by a lower thermal boundary layer, and substantial flow between the upper mantle and the lower mantle, though possibly with localised and transient interruptions.

There have been arguments made that geochemical evidence requires the mantle to be layered (Allegre et al. 1987, Jacobsen & Wasserburg 1979, O'Nions 1987). The evidence certainly requires the source of mid-ocean ridge basalts (which must be the shallow mantle) to be different from sources represented by hotspots, which must be deeper. However, there is little else that can be concluded solely on the basis of the geochemistry. It is plausible, though not proven, that the geochemical heterogeneity of the mantle can be explained by slow mantle mixing at depth due to higher viscosity, separation of the heavy oceanic crust component at the base of the mantle, and inhibition of flow through the transition zone (Christensen & Hofmann 1994, Davies 1990, 1995a). Neither the geochemistry (Hofmann 1997) nor the geophysics (above) supports a simplistic picture with a depleted upper mantle and a primitive lower mantle.

Thermal evolution of the mantle: simple smooth case

It is possible to calculate the rate of change of mantle temperature, and thus to calculate the thermal history of the mantle. The results depend on how the mantle is being cooled. It is thus possible to calculate the implications of different assumptions, and to use the record of the Earth's tectonic history to test the models. Conversely, we will see here that the calculations suggest possibilities that have not been apparent until recently, and these may provide alternative ways to interpret the rock record. The objective is to gain the kind of fundamental understanding of the Earth's past tectonic regime(s) that we think we have for its present regime of plate tectonics.

I will not present equations here. These are available in the references. The rate of decrease of mantle temperature depends on the net rate of cooling. This is the cooling due to the action of plate tectonics, as described above, minus the heat input due to radioactivity and due to heat coming from the core. The latter is assumed here to be evident in the heat carried by plumes, which assumes that plumes come from the base of the mantle and carry the heat conducted from the core into the base of the mantle.

The key to these calculations is knowing how much heat is transported by convection (i.e. by plates and plumes). There is a simplified 'boundary layer theory' of convection that relates the rate of heat transport to the temperature difference between the surface (top or bottom) and the interior fluid (Davies 1993, Stacey & Loper 1984). The resulting expression also involves the mantle viscosity (which obviously affects the rate of convection), and this is a strong function of temperature: an increase of 100°C causes a viscosity decrease by about a factor of ten.

An example of the results of this kind of calculation is shown in Figure 2 (after Davies 1993). The top panel shows the temperatures at the top of the mantle (below the lithosphere) and at the top of the core. A schematic temperature profile through the mantle and core is shown in Figure 3. The core temperature is much higher because there is an adiabatic increase of temperature through the mantle of about 1000°C, as well as the difference in temperature between the deep mantle and the top of the core. The calculation in Figure 2 started with the lowest mantle at the same temperature as the core. The mantle cools rapidly, because it is initially hot and has a low viscosity, so convective cooling is fast. After about 0.5 Ga, the rate of mantle cooling slows, because of the effect of radioactive heating.

The lower panel of Figure 2 shows the contributions to the heat budget of the mantle. The heat loss ('plates') drops rapidly at first, like the mantle temperature, until it approaches the rate of radioactive heat generation ('heat gen'). Thereafter it tracks the heat generation. The mantle temperature drops only slowly in this phase (top panel). This is because of the strong temperature-dependence of the viscosity: it takes only a modest change in temperature to accommodate a large change in heat transport.

Initially there is no temperature difference between the deep mantle and the top of the core, by assumption. This means there will be no heat conduction from the core into the mantle, so there will be no hot thermal boundary layer at the bottom of the mantle and no plumes generated. Thus, initially, plumes transport no heat. As the mantle cools, the bottom thermal boundary layer becomes established and plumes can begin. In Figure 2, the lower panel shows that the heat transported by plumes rises rapidly from zero early in the calculation. Thereafter the plume flux is fairly constant. This is because both the mantle and the core are cooling only slowly, so the temperature difference between them does not change much, and the rate of plume generation does not change much.

The results in Figure 2 illustrate important features of the Earth's internal thermal system in a fairly simple case. An important point is that the cooling of the mantle is governed mainly by the decline of radioactive heating. This changes by about a factor of three since 3.5 Ga, but is accommodated by a change in mantle temperature of only about 200°C.

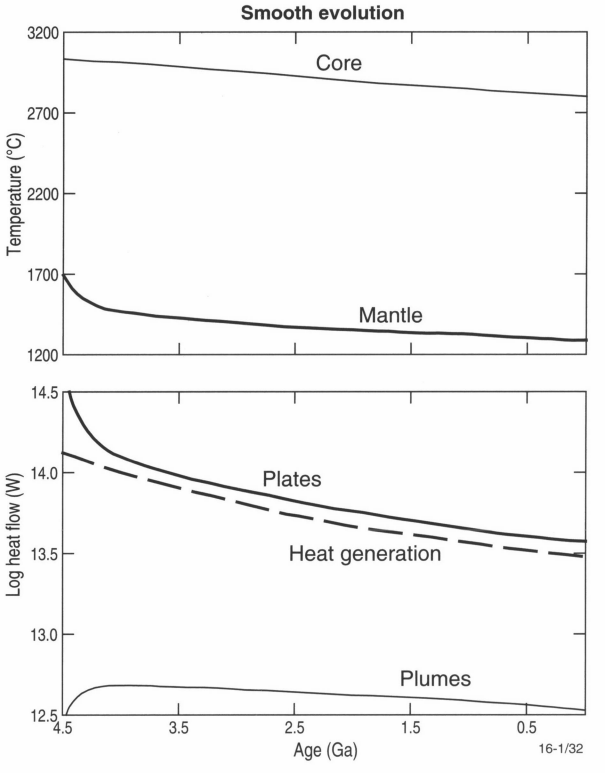

Figure 2. Thermal evolution of the core and mantle, assuming whole mantle convection, plate tectonics, and plumes coming from the core-mantle boundary. Top panel: temperatures at the top of the mantle (below the lithosphere) and at the top of the core. Bottom panel: heat flows in and out of the mantle. The 'plates' curve is the rate of heat loss from the mantle. 'Heat gen' is the rate of radioactive heat generation, and 'plumes' is the rate of heat input into the mantle in the form of plumes from the core-mantle boundary (from Davies 1993).

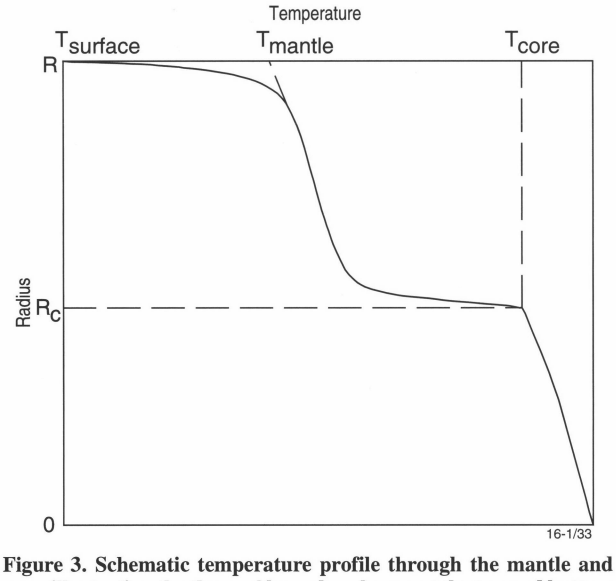

Figure 3. Schematic temperature profile through the mantle and core, illustrating the thermal boundary layers at the top and bottom of the mantle, the adiabatic temperature increase through the mantle, and the temperatures used to characterise the evolution of the mantle and core in the model of Fig. 2.

Mantle phase transformations

The mantle transition zone, between about 400 km and 670 km depth, is inferred to be a region in which upper mantle minerals go through a series of pressure-induced transformations to denser structures. Because the pressure at which a transformation occurs depends on temperature, the depth of the transformation is different in cold subducted lithosphere from the depth in surrounding hot mantle.

The transformation of Mg_2SiO_4 from the spinel structure to $MgSiO_3$ (perovskite structure) plus MgO (sodium chloride structure) occurs at a depth of about 660 km, and experiments indicate that its transformation pressure decreases as the temperature increases (Akaogi & Ito 1993). This is often expressed by saying the transformation has a negative Clapeyron slope, the Clapeyron slope (or more strictly, the Clausius-Clapeyron slope) being the temperature derivative of the transformation pressure.

The effect of this negative slope in subducted lithosphere is illustrated schematically in Figure 4. Within the slab, the lower temperature causes the phase boundary to occur deeper, where the pressure is higher. As a result there is a region (shaded) where the low-pressure phase exists within the slab at the same depth as the high pressure phase exists outside the slab. This region has a lower density, and so it is buoyant. This buoyancy (broad arrow) opposes the descent of the subducted lithosphere. If this buoyancy were sufficiently large, compared with the negative thermal buoyancy of the slab, it might prevent the slab from penetrating the phase transformation zone.

It was first shown by Machatel & Weber (1991) that this could cause intermittent layering in numerical models of a convecting fluid. Whether this might be true in the mantle has remained an open question because of the simplifications in many numerical models. More recent, somewhat more realistic models suggest that it is not so likely in the present mantle (Davies 1995a), which would be consistent with other evidence, summarised above, that the mantle is not layered at present.

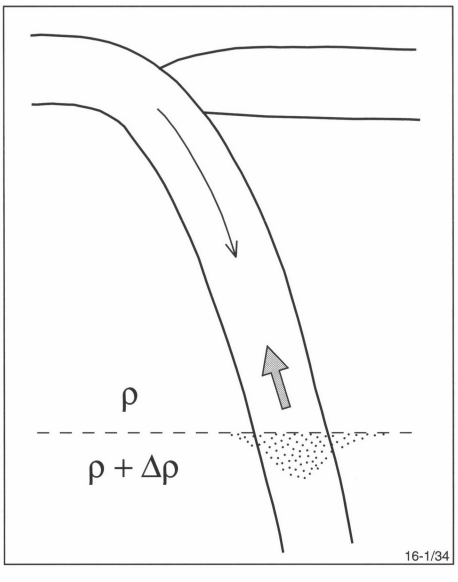

Figure 4. Sketch showing the deflection of a phase transformation boundary to greater depth in the subducting lithosphere because of the lower temperature in the subducting lithosphere and the temperature-dependence of the transformation pressure. In the zone of deflection (shaded) the less-dense low-pressure phase exists, and this region is buoyant as a result (broad arrow).

However, this mechanism would have been more able to induce layering in the past, when the mantle was hotter. Because the plates would have moved faster in that case, they would have been younger and thinner when subducted, and less able to penetrate a phase barrier. If layering was induced in this way, the plate cycle would cool the upper mantle. As the upper mantle cooled, there would come a point at which a plate was thick enough to penetrate the phase barrier, and the layering might then break down.

Numerical models indicate that the layering can also be broken down by another mechanism, which amounts to there being a limit on the temperature difference that can exist between the layers. As the upper mantle cools relative to the lower mantle, thermal boundary layers are formed on either side of the interface between them as heat conducts through the interface. These thermal boundary layers generate convection, and if this becomes sufficiently vigorous it might pull material through the interface. This becomes more likely as the temperature difference between the layers increases, for two reasons: first, the local convection is more vigorous, and second, thermal contraction of the upper layer increases the likelihood that the cooler upper mantle material can break through into the lower layer (and *vice versa* for the lower layer).

The effect of these rather complicated ideas is that layering will break down if one of two conditions is reached: (a) the upper mantle cools to a critical temperature (at which plates begin to break through); (b) the temperature difference between the upper mantle and the lower mantle reaches a maximum value (at which internal breakthrough may occur). The actual value of the critical temperatures is far from certain at present, but they can be estimated roughly, well enough for possible consequences to be explored.

One example of a possible mantle thermal evolution with these effects included is shown in Figure 5 (Davies 1995b). A smooth ('whole mantle') evolution similar to that shown in Figure 2 is included for comparison. The first impression of the effect of the phase barrier must be that it can produce dramatically different behaviour: the temperatures of the upper mantle and lower mantle undergo large swings through the earlier parts of Earth history, ultimately to die out and approach the smooth "whole mantle" curve. Before looking at the details of this model, two main aspects of the character of the behaviour can be noted. First, the model is episodic, more like the apparently episodic history of additions to the continental crust. Second, there are three long phases of behaviour, reminiscent of the three tectonic eras of Earth history.

The model starts with the upper mantle and the lower mantle both at a high temperature. In this condition layering occurs. The upper mantle cools, because of the cycle of plate formation and subduction. The lower mantle heats, because it is heated by radioactivity but not cooled by the plates. As their temperatures diverge there is some heat transfer between them, involving convection driven by the internal thermal boundary layers (see Davies 1995b for details). However it is not sufficient to stop the divergence, and a breakdown of the layering occurs when the temperature difference reaches the critical value. At this point it is assumed, for the purposes of the calculation, that hot material from the lower mantle replaces the cooled material, and the latter is mixed with the remaining lower mantle material. Thus the temperature of the upper mantle is reset to the temperature the lower mantle had reached, while the temperature of the lower mantle is reset to the mean value of the former upper mantle and the remaining lower mantle. The evolution is then continued.

In this model, four such mantle overturns occur, the last at about 2.7 Ga. The following overturn, nearly as large, is part of the second series (discussed below; Fig. 5). After each overturn, the upper mantle cools to a lower temperature. At

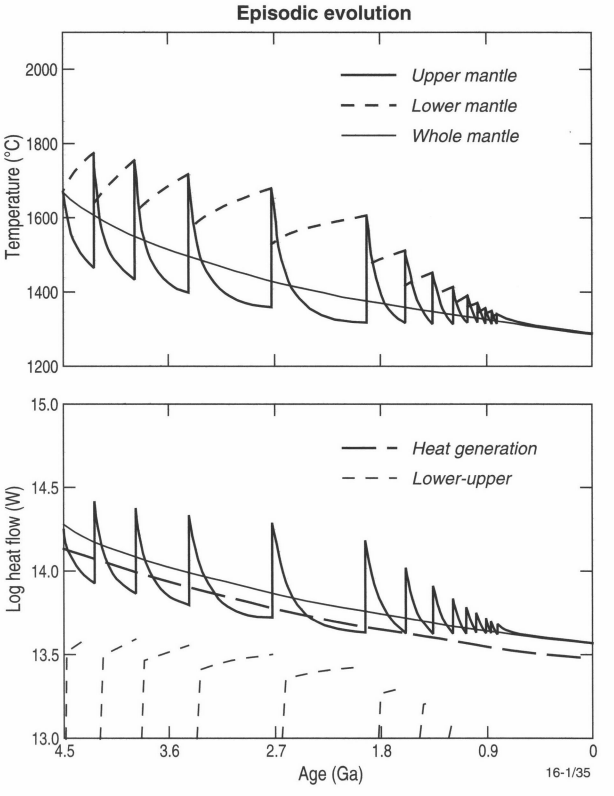

Figure 5. An example of a thermal evolution in which a phase transformation barrier causes transient layering of the mantle, which then convects separately in the upper mantle and the lower mantle. An evolution assuming whole-mantle convection, similar to that in Fig. 2, is included for comparison. When the mantle is layered, the upper mantle and lower mantle temperatures diverge. When the layering breaks down (see text), the cooled upper mantle material is replaced by hot lower mantle material, and the remaining lower mantle material is assumed to mix with the cool material displaced from the upper mantle. Lower panel shows heat flows, including the heat generated by radioactivity and the heat transferred across the interface between the upper mantle and the lower mantle ('lower-upper'). After (Davies 1995b). For this case, the transition zone heat transfer parameter is $b_z = 0.6$ and plate penetration occurs at an age of 72 m.y. ($b_u = 3.25$).

about 1.9 Ga in this model, it has cooled to the critical temperature at which plates can start to penetrate the phase barrier, and another overturn is assumed to occur. There follows a second series of overturns triggered by plate penetration. These occur after progressively smaller intervals and with progressively small temperature variations, because the temperature of the lower mantle is being progressively reduced. Eventually, by about 700 Ma, the lower mantle temperature is reduced to the same as the upper mantle, and whole mantle convection is assumed to occur thereafter.

There are other possible courses of this type of thermal evolution, and in fact the parameters of this one, although plausible, have been adjusted to correspond to particular features of the Earth's tectonic history. Some of the other possibilities are illustrated in Figure 6. In Figure 6(a), the ability of subducted lithosphere to penetrate the phase barrier has been increased by assuming it can penetrate by the time it is 25 m.y. old (the 'buckling parameter' $b_u = 5.5$ (Davies 1995b)), compared with penetration at 72 m.y. ($b_u = 3.25$) in Fig. 5. In this case the sequence of overturns triggered by plate penetration begins and finishes much earlier, and all overturns are confined to the Archaean.

Figure 6(b) shows the effect of more efficient heat transfer between the upper and lower mantles (parameter $b_z = 0.8$, compared with $b_z = 0.6$ in case (a); Davies 1995b). In this

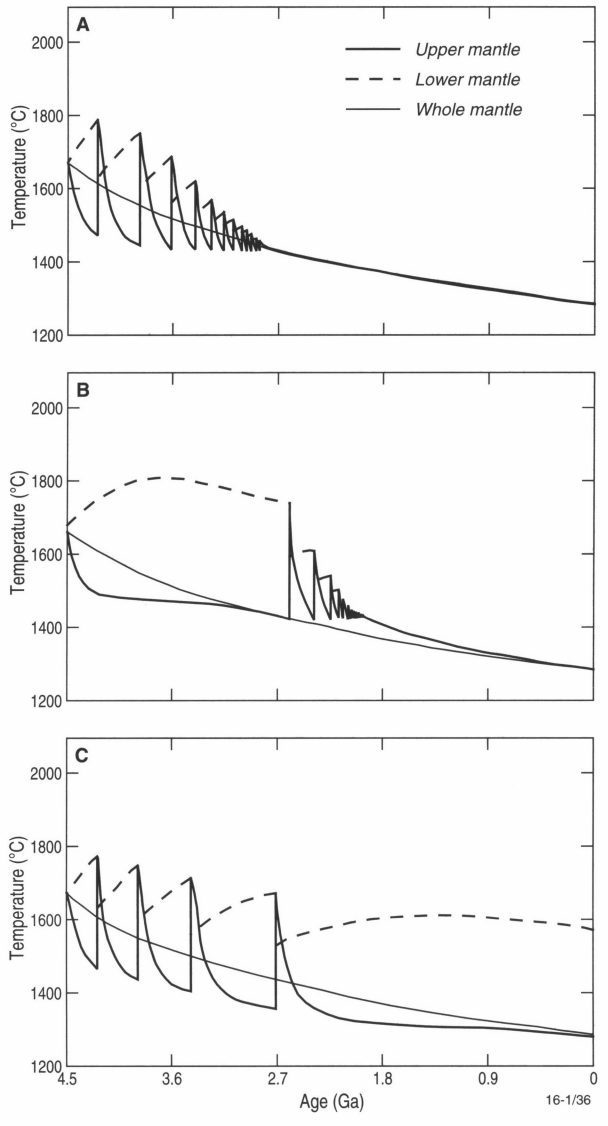

Figure 6. Further examples of thermal evolution models with episodic layering. (a) Plates penetrate at an age of 25 m.y. ($b_u = 5.5$). (b) Heat transfer between layers is more efficient ($b_z = 0.8$; $b_u = 5.5$). (c) Plate penetration is assumed not to occur (with $b_z = 0.6$). This model is less plausible (see text).

case sufficient heat is transferred to keep the temperature difference between the upper and lower mantles from reaching the critical value, and the internally triggered overturns never occur. The mantle is stably layered for a long period through the Archaean. Overturns triggered by plate penetration begin at 2.6 Ga and whole-mantle convection is achieved by 1.9 Ga.

Figure 6(c) is like the case in Figure 5, except that plate penetration is, in effect, turned off. This again shows a long period of stable layering, after the internally triggered overturns have ceased. However this model seems less plausible, since it is necessary to assume that plates as old as 100 m.y. cannot penetrate the phase barrier.

The most important point here is not the details of a particular calculation, but the overall character of this type of thermal history.

Implications of mantle overturns

The task of assessing whether the Earth has undergone an evolution of this type may not be easy. A preliminary discussion has been given elsewhere (Davies 1995b). Here, only some key points will be mentioned. The potential geological implications of the kind of mantle overturn that occurs in the model of Figure 5 are difficult to predict in any detail, but it is easy to show that they could be dramatic.

The arrival in the upper mantle of material 200°C or more hotter than the material it displaces would generate a major magmatic event. Flood basalts are believed to be generated by the arrival of a plume head 100–200°C hotter than ambient upper mantle (Campbell & Griffiths 1990, Richards et al. 1989), but a typical flood basalt province covers less than 1% of the Earth's surface. Thus a mantle overturn event might be like 100 flood basalt events occurring simultaneously. There would probably be substantial tectonic effects too, as mantle flow rates would be much larger during an overturn and the hotter mantle would have lower viscosity: both factors would be likely to cause plates to move much faster. It is plausible that an overturn event could generate and aggregate large amounts of mafic crust, and that such crust would be substantially reworked either in the same or subsequent overturn events (Davies 1995b).

Whether such an event might correspond to one of the major crust-forming episodes, such as the 2.7 Ga or 1.9 Ga episodes, will not be easy to determine. The main point here is to demonstrate the possibility, so that it can be considered along with other hypotheses. At present, the main hypotheses for episodic events would seem to be mantle overturns, large plume heads (Campbell & Griffiths 1990, Campbell & Hill 1988), or continental breakup and aggregation (Condie 1995, Hoffman 1988, 1991).

Buoyancy of the oceanic crust

Because the oceanic crust is less dense than the mantle (2.9 g/cm^3 compared with 3.3 g/cm^3), the oceanic lithosphere must cool for about 15 m.y. before its negative thermal buoyancy overcomes the buoyancy of the oceanic crust. The mantle residue from melting at mid-ocean ridges is also less dense than normal mantle (Ringwood & Irifune 1988) (Figure 7a). In combination, these compositional buoyancies mean that oceanic lithosphere has a net positive buoyancy until it is about 20 m.y. old. Since it is the negative buoyancy of the lithosphere that ultimately drives plate tectonics, this means that plates younger than about 20 m.y. could not be driven.

If the mantle was hotter in the past, two things would be different, and each would amplify this effect. First, a hotter

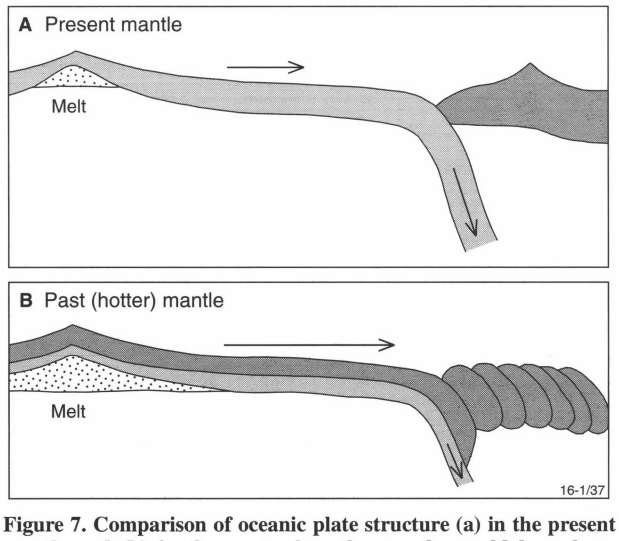

Figure 7. Comparison of oceanic plate structure (a) in the present mantle and (b) in the past, when the mantle would have been hotter. A hotter mantle would produce more melting under mid-ocean ridges and hence thicker oceanic crust. In addition, the viscosity of a hotter mantle would be lower, so convection would go faster and the plates would be younger when they reached a subduction zone. It is possible that the thick, buoyant oceanic crust would resist subduction and accumulate at the surface.

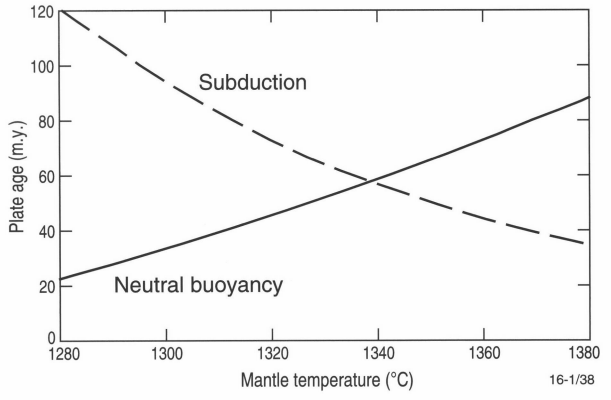

Figure 8. Age at which oceanic lithosphere becomes neutrally buoyant (solid) and mean age of plates upon subduction (dashed) as functions of mantle temperature. If the mantle were hotter, the resulting thicker oceanic crust (Fig. 7) would make the lithosphere positively buoyant until the age given by the solid curve, and the lithosphere would arrive at a subduction zone sooner (Fig. 7), when its age was that given by the dashed curve. At temperatures about 60°C above the present mantle temperature the curves cross, meaning the plates would, on average, be only neutrally buoyant when they arrived at a subduction zone, so there would be no force to pull them down. At higher temperatures, the plates would have to move more slowly in order to be old enough to subduct (following the solid curve), or they might delaminate the crust and mantle parts, the crust accumulating at the surface and the mantle part subducting (Fig. 7). In either case they would remove less heat from the mantle than was necessary to cool it (after Davies 1992).

mantle would generate a thicker oceanic crust, due to more melting at mid-ocean ridges (Fig. 7b). This would increase the compositional buoyancy. Second, the mantle viscosity would be lower, so mantle convection and plates would go faster, and plates would be younger on average when they reached a subduction zone. This would reduce their negative thermal buoyancy. Figure 8 shows how these two effects vary as a function of mantle temperature (Davies 1992). It shows that if the mantle is about 60°C hotter than at present, the oldest plates would only be neutrally buoyant when they reached a trench.

At higher mantle temperatures, plate tectonics could still be driven, but more slowly than is implicit in calculating the 'subduction' curve in Figure 8. The result would be that plate tectonics would no longer remove heat fast enough to cool the mantle. This leads to a paradox: it implies that the mantle would be hot and getting hotter, and could not have cooled to its present temperature.

The paradox can be avoided only by assuming that the plates were not behaving as they do today, or that something other than plate tectonics was operating. Two possibilities can be mentioned here. One is that the transformation of the basaltic oceanic crust to dense eclogite might keep plates moving. This would depend on the transformation proceeding at a sufficient rate, which in turn depends on what controls the transformation. If it is limited by reaction kinetics at low temperature, then it might proceed only slowly, and might not be viable. If it is possible to force the transformation by overdriving the pressure, then it might proceed more rapidly. In any case, there is a threshold pressure, equivalent to roughly 60 km depth, which the crust must reach before it can transform.

A vigorously subducting plate might maintain its subduction, through the deeper part pulling the shallow, buoyant part after it. However, if such subduction were ever interrupted, it might be difficult to restart. This might involve accumulating a large pile of basaltic crust until its root became deep enough and hot enough to transform and sink. The result might be a more episodic form of plate tectonics.

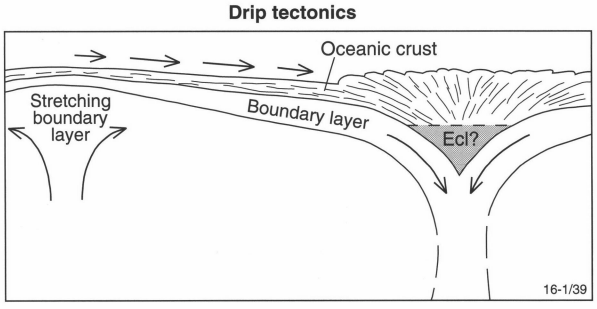

Figure 9. Sketch of possible non-plate behaviour of the mantle's upper thermal boundary layer that might have occurred when the mantle was hotter. The mantle part of the boundary layer is too thin and weak to behave as a rigid plate, so it stretches over a broad zone of upwelling and descends symmetrically under a convergence zone. Thick oceanic crust piles up over the convergence zone. The lower part of this pile might transform to eclogite ('Ecl') and sink episodically (after Davies 1992).

A second possibility, if conventional subduction did not work, might be that the crustal component of the lithosphere separated from the mantle component and accumulated at the surface. A scenario along these lines is illustrated in Figure 9 (Davies 1992). Unfortunately, this mode of tectonics would be less efficient at cooling the mantle than plate tectonics, because the coldest part is left at the surface. It is not clear that this is a viable way of cooling the mantle.

It is not immediately obvious how these possibilities can be quantified in order to estimate the resulting plate velocities and rates of mantle cooling, and this has not yet been done, hence the qualitative and conjectural nature of this discussion. It is clear, however, that there are good reasons of physics to expect that tectonics would have operated differently when the mantle was hotter. What is not clear is whether the tectonics would have been a modified version of plate tectonics or something quite different, perhaps like that sketched in Figure 9.

This means that the thermal evolution calculations presented earlier may not be accurate for those times when the mantle was substantially hotter than at present. What can be said is that the mantle clearly has been able to cool to its present temperature, that the rate of cooling is likely to depend strongly on mantle viscosity and hence on mantle temperature, and that as a result the character of the calculations presented above is likely to be retained as our understanding improves, though details are subject to substantial revision.

Plumes

The existence of mantle plumes can be inferred with little uncertainty from the existence of the volcanic hotspots and their associated hotspot swells (Davies 1988, Davies & Richards 1992). The swells are not due to thickened crust, and the lithosphere is not strong enough to hold up such broad features (about 1000 km across), so the only remaining possibility is that they are held up by buoyant mantle under the lithosphere. As noted above, it can be inferred from the swells that mantle plumes carry no more than about 10% of the Earth's heat budget. The only plausible source of plumes is a hot thermal boundary layer at depth. This is taken here to be at the base of the mantle, for the reasons given earlier.

As shown earlier, if whole mantle convection applied throughout Earth history, then the plume flux would probably not have changed very much (Fig. 2). However, if there were periods in which the mantle was layered, then, by assumption, plumes from the base of the mantle would not have reached the surface. Instead, a thermal boundary layer would form at the base of the *upper* mantle, and smaller, more numerous plumes would probably arise from this. The flux of such upper

mantle plumes could be greater than the present plume flux. This can be seen from Figure 5, in which the 'lower-upper' heat flow is the heat passing through the interface between the upper mantle and the lower mantle. It is this heat that would be transported by the upper mantle plumes.

Plumes were not included in the calculations of episodic thermal evolution presented above. This was justified as a first approximation by noting that the plumes transport only a fraction of the Earth's heat budget and thus will not exert the dominant control on mantle temperatures (Davies 1995b). However, it is possible that plumes may be able to play a role in triggering mantle overturns. This possibility has not yet been explored. It is one more reason that the details of Figures 4 and 5 are still conjectural.

It is sometimes suggested that 'plume tectonics' could substitute for plate tectonics, either in the Earth's earlier history or on other planets, particularly Venus. This betrays a misunderstanding of the role of plumes. The existence and vigour of plumes depends on the presence and strength of a hot thermal boundary layer at the base of the convecting layer. It follows that the existence of plate tectonics does not directly affect the role of plumes. Plate tectonics is the current expression of the upper, cool thermal boundary layer of the mantle. If this boundary layer operated differently in the past, then something other than plate tectonics may have prevailed, but plumes would have existed anyway, so long as the mantle was cooler than the core, or the upper mantle was cooler than the lower mantle. Plumes and plates play different roles. Plumes are an expression of heat entering the mantle from below, while plates are an expression of heat leaving the mantle at the top.

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