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# MANTLE DYNAMICS: THE STRONG CONTROL OF THE SPINEL–PEROVSKITE TRANSITION AT A DEPTH OF 660 KM

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**Abstract**—The strong control that the endothermic phase change from spinel to perovskite and magnesiowüstite at a depth of 660 km has on mantle convection is discussed. The phase transition determines the morphology and length scales of upflow and downflow structures and, through retardation of sinking slabs, can cause an avalanche phenomenon involving rapid flushing of cold upper mantle material down to the base of the lower mantle. The phase change significantly heats plumes that rise from the lower mantle and penetrate into the upper mantle. The exothermic phase change from olivine to spinel at a depth of 400 km in the mantle mitigates the effects of the dynamically and thermally dominant endothermic phase transition.

## INTRODUCTION

The most significant step forward in our understanding of mantle convection in recent years has been the realization that the phase changes of the transition zone, particularly the transition from spinel to perovskite and magnesiowüstite at a depth of 660 km, can have a profound influence on the dynamical state of the mantle. There has been an explosion of activity on this aspect of mantle convection following the initial paper by Machetel and Weber (1991). Several developments account for this, including recent laboratory experiments identifying the seismic discontinuity at 660 km depth in the mantle as a phase change and quantifying the thermodynamic properties of the transition, recent seismological studies revealing the nature of the interactions of descending slabs with the phase change at 660 km depth, and recent numerical simulations of a convecting mantle containing the phase change. In this paper we review this recent activity and both describe and explain the strong control that the spinel–perovskite phase transition has on mantle dynamics.

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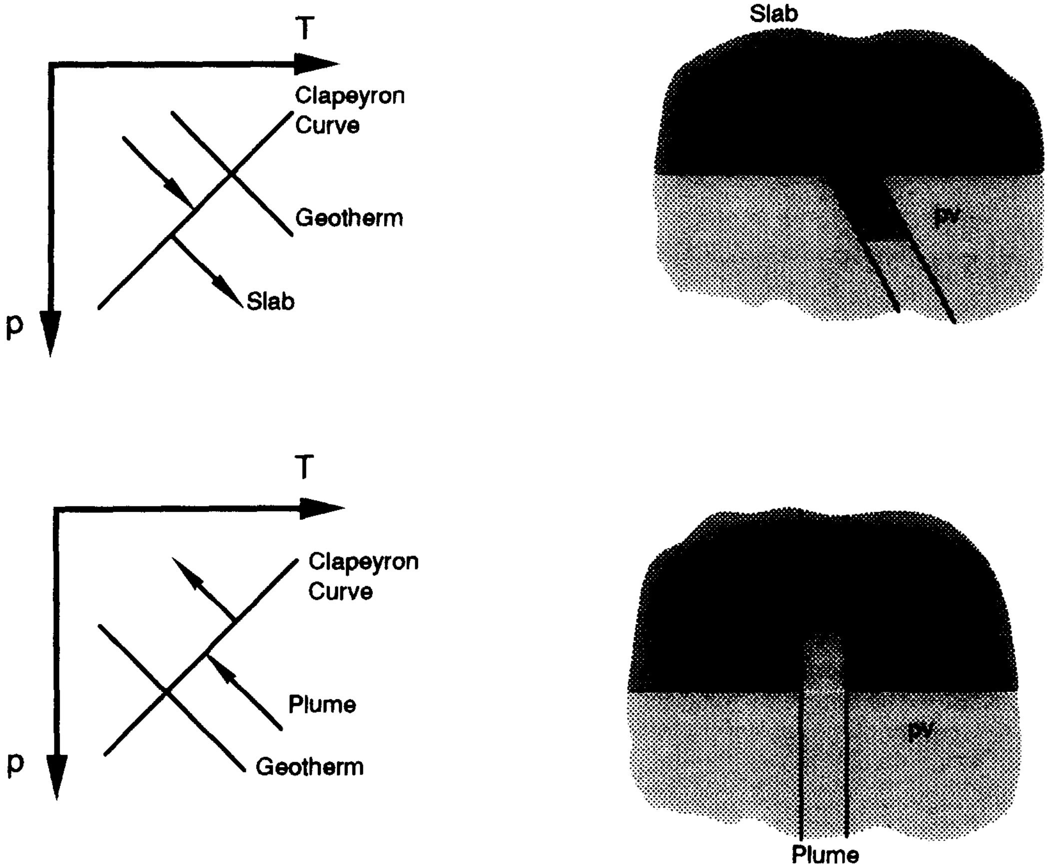


Fig. 1. Sketch of the trajectory for a descending slab and a rising plume in a $p$–$T$ diagram. The location of the endothermic phase transition in the background state is determined by the intersection of the Clapeyron curve and the geotherm. The phase transitions in the plume and the slab occur at the intersections of the Clapeyron curve with the trajectories of the plume and slab. The phase change in the plume and slab actually occur over a narrow two-phase region. pv—perovskite; sp—spinel.

## DYNAMICS OF AN ENDOTHERMIC PHASE CHANGE

The solid–solid phase change from spinel to perovskite and magnesiowüstite at 660 km depth in the mantle (Ito and Takahashi, 1989; Shearer, 1991; Shearer and Masters, 1992) is unusual in that it is an endothermic phase change, i.e. heat is absorbed when the less dense spinel converts to the more dense perovskite and magnesiowüstite. This endothermic phase change has a negative slope in a pressure $p$–temperature $T$ diagram as sketched in Figure 1. The slope of the line separating the light and heavy phases in a $p$–$T$ diagram is the Clapeyron slope $\gamma$ given by the Clausius–Clapeyron equation

$$\gamma = \frac{dp}{dT} = \frac{L\rho_{\mathrm{l}}\rho_{\mathrm{h}}}{T(\rho_{\mathrm{h}} - \rho_{\mathrm{l}})} \tag{1}$$

where $L$ is the latent heat of the transition (positive if heat is liberated when the light phase transforms to the heavy phase), $T$ is the temperature at which the transition occurs, and $\rho_l$ and $\rho_h$ are the densities of the light and heavy phases, respectively. In this description we are approximating the phase change at 660 km depth as if it were a univariant phase change, a good approximation since the phase transition actually occurs over a depth interval of only several kilometers (Ito and Takahashi, 1989; Benz and Vidale, 1993).

Consider now what effect a cold descending slab would have upon encountering the phase change (Fig. 1). Since the temperature in the slab is less than in the surroundings, the phase transition from spinel to perovskite would occur at a higher pressure or greater depth in the slab than in the surroundings, as shown in Fig. 1. The opposite would occur for an upwelling plume encountering the phase change, i.e. the transition from perovskite to spinel would occur at a smaller pressure or shallower depth in the plume than in the surroundings (Fig. 1). The effect of phase boundary displacement is to make the slab lighter and the plume heavier than their immediate surroundings in the vicinity of the phase change. The descending slab and ascending plume thus experience inhibitive forces in their attempts to traverse the phase change. This is the main dynamical effect of the endothermic phase change at 660 km depth on mantle convection. There are other effects associated with the latent heat release or absorption that will be elaborated below.

The inhibition of mantle convection by an endothermic phase change has been understood for some time (Schubert *et al.*, 1975; Christensen and Yuen, 1984). What has happened recently to bring the potential importance of this effect to the forefront of mantle convection studies is that the value of $\gamma$ for the spinel–perovskite phase change has been determined experimentally (Ito and Takashi, 1989; Ito *et al.*, 1990; Akaogi and Ito, 1993) to be sufficiently negative ($-2$ to $-4$ MPa $K^{-1}$) that the force opposing flow through the phase change is dynamically important (this force is proportional to the Clapeyron slope $\gamma$).

## RESULTS FROM NUMERICAL EXPERIMENTS

Numerical studies of the effects of the spinel–perovskite phase transition on mantle convection have been carried out in two-dimensional planar and axisymmetric geometries (Christensen and Yuen, 1984; Machetel and Weber, 1991; Liu *et al.*, 1991; Zhao *et al.*, 1992; Peltier and Solheim, 1992; Weinstein, 1993; Solheim and Petier, 1994) and in three-dimensional cartesian (Honda *et al.*, 1993; Yuen *et al.*, 1994) and spherical shell (Tackley *et al.*, 1993, 1994) geometries. Our discussion here will focus on the three-dimensional spherical shell results. The main effects of the endothermic phase change on the style of mantle convection are shown in Figs 2–4. Figure 2 shows the morphology of cold downflow structures which consist of a network of interconnected sheets in the upper mantle and cylinders in the lower mantle. The sheet-like downflows do not extend below 660 km depth. Material in these sheets cannot penetrate the phase change; instead, this material flows laterally on top of the phase change and accumulates in quasi-cylindrical piles at the intersections of the sheets. When enough cold and heavy material piles up to overcome the inhibitive force of the phase boundary, the pile breaks through the phase boundary and sinks into the lower mantle in an event described as an avalanche (Tackley *et al.*, 1993). The cylindrical avalanches in the lower mantle spread out like pancakes when they fall onto the lower or core–mantle boundary. The avalanches provide efficient conduits for rapidly flushing large quantities of cold material all the way to the core–mantle boundary where the spreading cold material effectively cools the core.

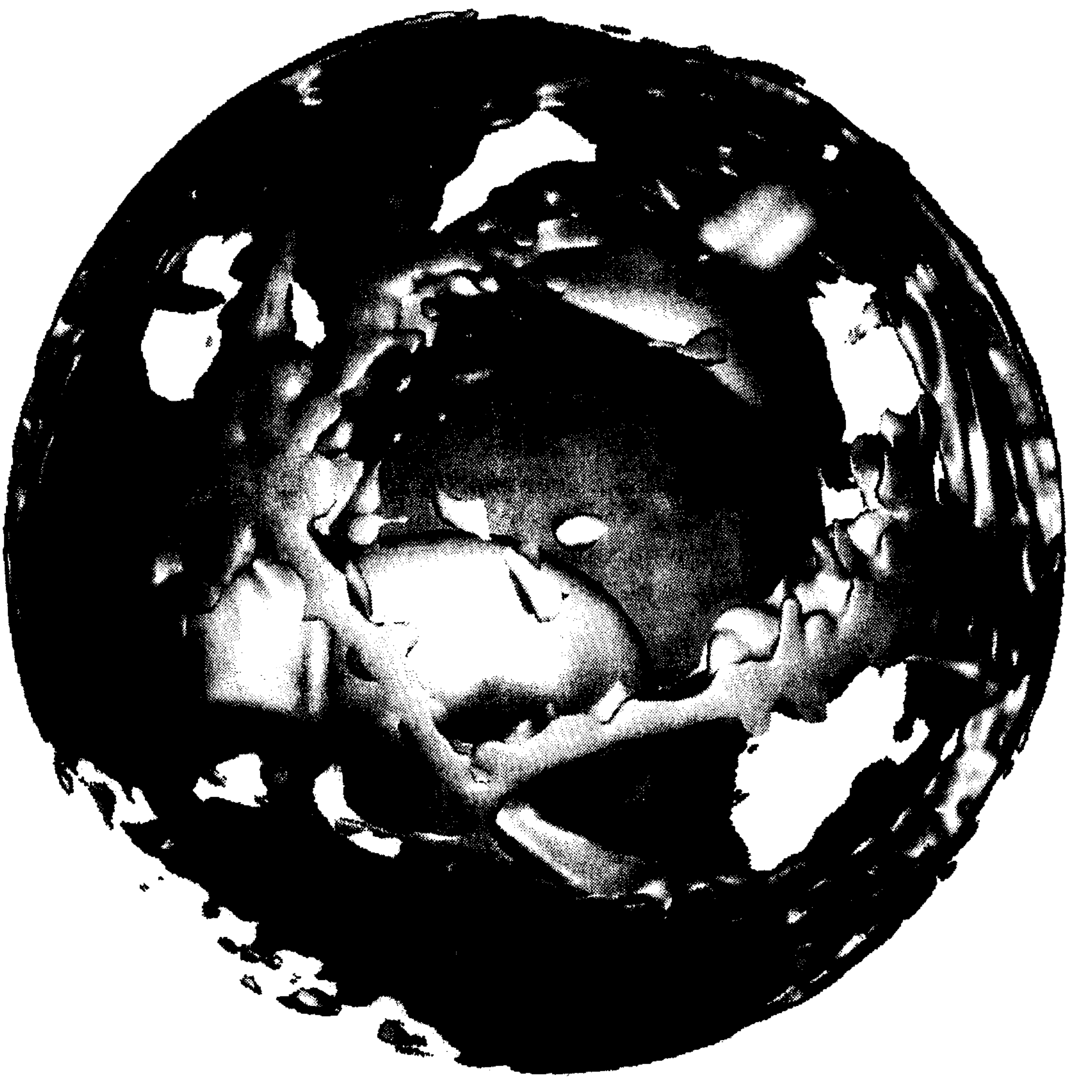

Fig. 2. An isosurface showing where temperature is 110 K lower than the horizontally average value. The inner sphere is the core. A network of interconnected linear downwellings is seen in the upper mantle. Three large cylindrical downwellings or avalanches in the lower mantle spread like cold pancakes on the core–mantle boundary (after Tackley *et al.*, 1993).

Fig. 3. A hot isosurface corresponding to the cold isosurface of Fig. 2. Temperature on the surface is 110 K higher than the reference state adiabat. A single plume from the core–mantle boundary feeds a hot region of the upper mantle; many hot regions in the upper mantle are not directly connected to the lower mantle. Ridges of hot material form a connected network on the core–mantle boundary (after Tackley *et al.*, 1993).

The picture of mantle convection provided by Fig. 2 is a snapshot at one particular time. The scene changes dramatically in detail with time, the positions and numbers of upper mantle sheets and lower mantle cylinders change, but the overall character is unchanged. There are always upper mantle sheets and lower mantle cylinders. Avalanches occur sporadically in space and time. At any one time there are always several avalanches in progress. Convection is therefore of the whole-mantle type since there is always substantial mass exchange between the upper and lower mantles. Some two-dimensional models (Peltier and Solheim, 1992; Weinstein, 1993; Solheim and Peltier, 1994) on the other hand

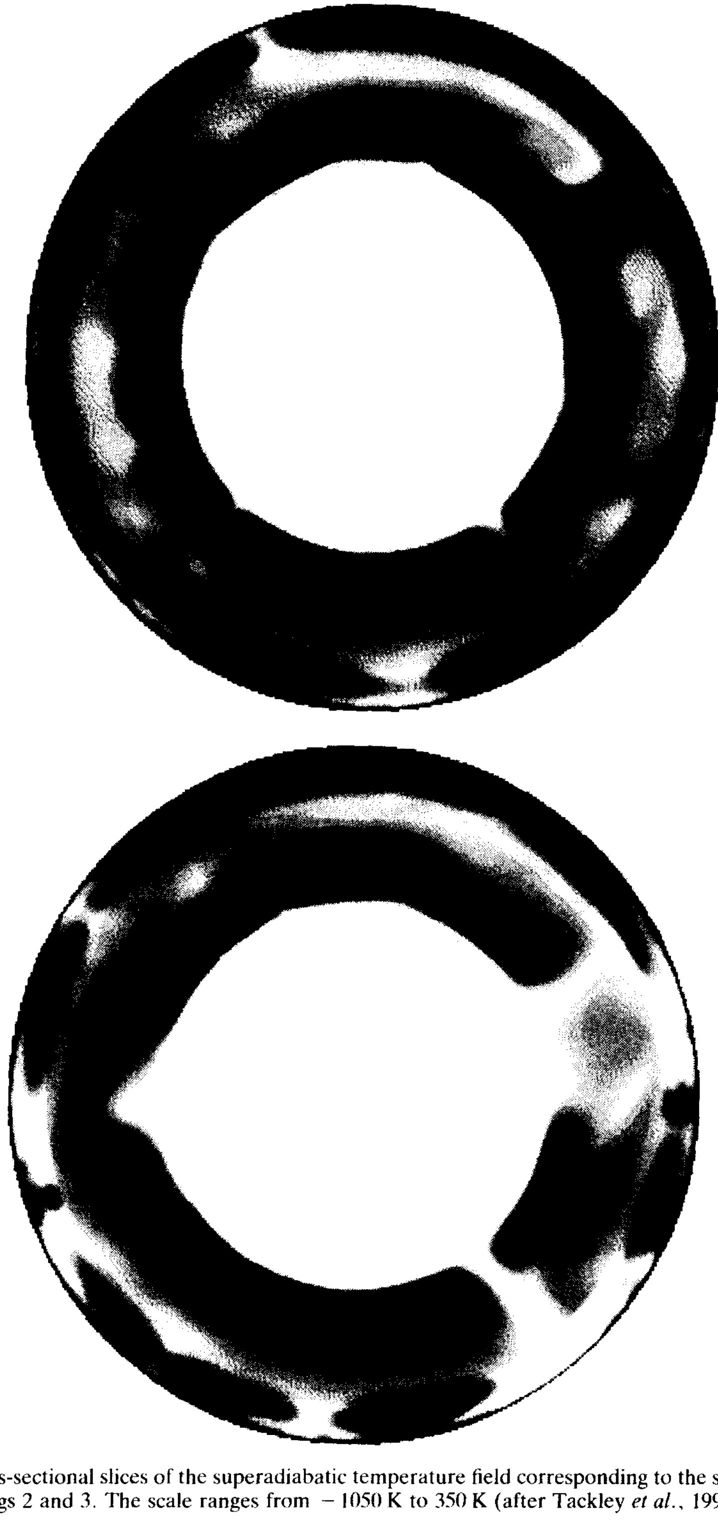

Fig. 4. Two cross-sectional slices of the superadiabatic temperature field corresponding to the solutions shown in Figs 2 and 3. The scale ranges from − 1050 K to 350 K (after Tackley *et al.*, 1993).

predict long periods of time in which the endothermic phase change cuts off any mass transfer across the transition, forcing the mantle into a two-layer mode of convection which is eventually terminated by a massive avalanche of material that has piled up on the phase boundary. Intermittent whole-layer and two-layer mantle convection on a global scale is not encountered in three-dimensional models.

The hot upwellings that are complementary to the cold downflows of Fig. 2 are shown in Fig. 3. There are broad hot regions in the upper mantle, hot ridges at the core–mantle boundary, and a plume of hot material connecting the core–mantle boundary with one of the hot upper mantle regions. This picture is again a snapshot of a highly time-variable situation. The hot ridges on the core–mantle boundary are swept around and form and reform in response to the avalanches of cold material falling sporadically on the core–mantle boundary. Plumes of hot material erupt irregularly from ridge intersections. The number of plumes in this simulation is relatively small because of the thorough blanketing of the core by the cold avalanches. Most of the upflow occurs in a broad featureless rise of material characteristic of internally heated convection.

Both the cold downflows and hot upflows are simultaneously visible in the temperature cross-sections of Fig. 4. There are horizontally broad regions of both hot and cold material in the upper mantle in addition to small-scale convection cells. The broad regions of cold material correspond to the pile-up of downflows on the endothermic phase change. Some of the broad hot regions are fed by hot material from mantle plumes. There is strong heterogeneity in the upper mantle at large horizontal dimensions as a consequence of the inhibitory effect of the phase change on the sheet downwellings. There is also broad horizontal structure at the core–mantle boundary which is almost surrounded by cold avalanche material. One of these avalanches connects the upper mantle to the core–mantle boundary. Another avalanche is in its earliest stages, before the sinking cold material has fallen onto the core–mantle boundary. A couple of well-developed mantle plumes are seen to emerge from the core–mantle thermal boundary layer; a few incipient plumes are also visible.

The above results emphasize the role of the endothermic phase change in producing avalanches and controlling the dominant length scales in different regions of the mantle. The behaviour of the phase change responsible for these results is the displacement of the phase boundary in response to changes in temperature. Though the latent heat of the phase change has less important dynamical effects, it is still significant for the thermal state of plume material reaching the upper mantle as shown in Fig. 5 (after Schubert *et al.*, 1995). The figure compares the thermal anomalies of two plumes; one that has passed through the endothermic phase change and one that has not. The plume that has traversed the phase change is 100 K hotter in the upper mantle than the plume that did not pass through the phase change. Passage of a plume through the phase change at a depth of 660 km heats the plume through the absorption of latent heat released by the endothermic phase change as more dense perovskite

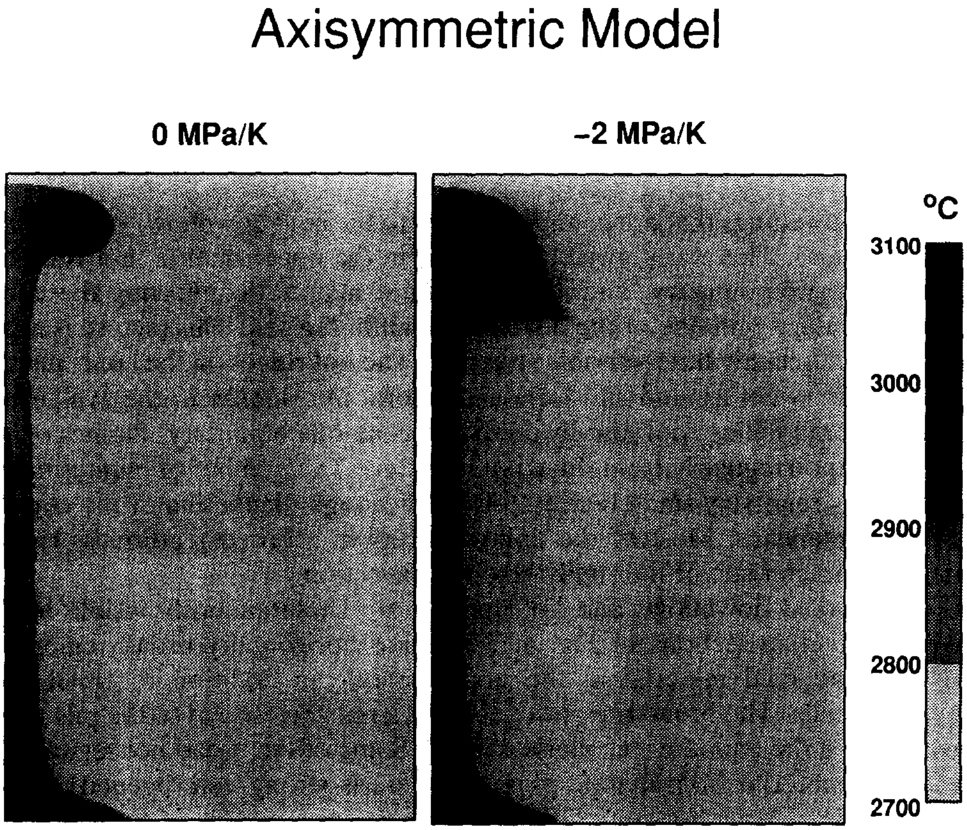


Fig. 5. Temperature field of two axisymmetric plumes, one that has passed through an endothermic phase change (right) and one that has not (left). The plume that went through the phase change is about 100 K hotter than the other plume (after Schubert *et al.*, 1995). The temperature scale is in °C.

converts to less dense spinel. The additional plume heat has important consequences for the manifestation of the plume as a hotspot. Heat release in regions of material upwelling through the endothermic phase change can also result in the formation of diapiric structures in the upper mantle (Liu *et al.*, 1991).

## DISCUSSION

Comparison of the above model results with new seismic observations of the mantle shows that the spinel–perovskite phase change at 660 km depth holds the key to explaining the observations and understanding mantle convection. The seismic tomographic images of van der Hilst *et al.* (1991) and Fukao *et al.* (1992) suggest that the slabs beneath the southern Kuril, Japan, and Izu Bonin arcs meet resistance at the 660 km depth and tend to flatten and move horizontally at this depth, while the slabs beneath the Mariana and Kuril–Kamchatka arcs plunge through 660 km depth into the lower mantle. The flattening out of the slabs may be analogous to the pile-up of cold sheet-like downflows on the

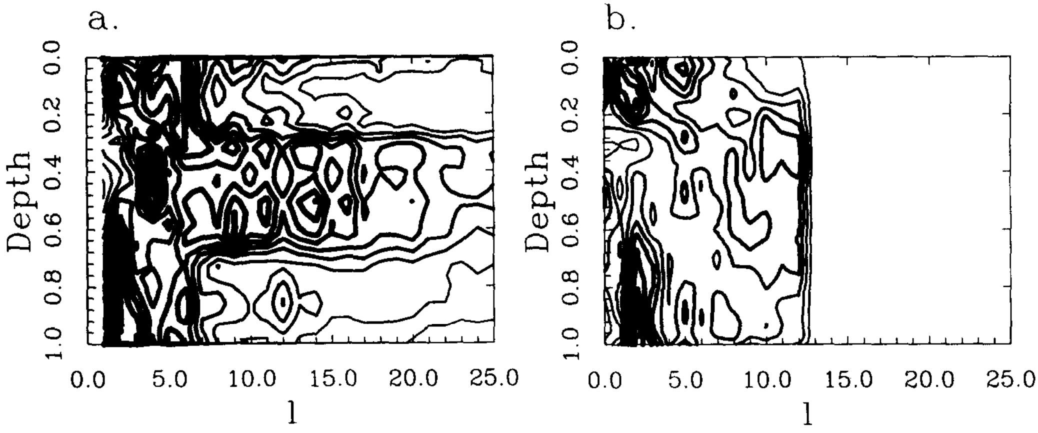


Fig. 6. A spectral heterogeneity map for thermal anomalies in the calculation of Figs 2–4 (left) and seismic velocity in the seismic model SH12/WM13 of Su *et al.* (1992) (right). The depths are normalized to the total thickness of the model shell or the mantle thickness, as appropriate. The parameter l is the degree of the spherical harmonic. The plots are normalized by the total root mean square amplitude at each radius (after Tackley *et al.*, 1994).

endothermic phase boundary in the model calculations, a consequence of the downward displacement of the phase change caused by the cold temperatures of the downwellings. Why some real slabs are able to penetrate the 660 km depth phase change while no model slabs are capable of this may be due to differences in the numerical models and the Earth, of which there are many. For example, real slabs are relatively rigid due to the strong dependence of mantle viscosity on temperature, while the three-dimensional spherical numerical models are not yet able to account for variable viscosity. The rigidity of real slabs may also affect how accumulations of slabs on the 660 km discontinuity (if there really are any) might ultimately sink under their weight into the lower mantle because slab material is not able to flow horizontally along the strike of a slab as can the less viscous material in the model calculations. Cylindrical accumulations of material at slab intersections above the 660 km depth phase transition may not be able to form in the Earth, and avalanches, if they occur in the Earth, may be unlike those in the numerical models. Some real slabs may be able to penetrate the 660 km depth phase change because of the angle at which they impinge on the transition; in the numerical model the sheets fall vertically on the phase change, the most favourable angle for penetration.

Global seismic tomography (Su *et al.*, 1992) has shown the structure of mantle heterogeneity to be characterized by relatively strong long wavelength variations in the upper mantle, relatively weak variations over a broad wavelength spectrum in the upper part of the lower mantle, and relatively strong long wavelength variations near the base of the mantle (Fig. 6). The spectrum of thermal heterogeneity in the model calculations of Figs 2–4, also shown in Fig. 6, bears a striking resemblance to the observed spectrum of seismic heterogeneity.

There is a similar concentration of power at long wavelengths in the upper mantle and in the bottom of the lower mantle of the model. The endothermic phase change in the model is essential in establishing these predominant wavelengths in the different regions. The piling up of cold downflows on the phase boundary creates the long wavelength upper mantle heterogeneity, while the spreading of cold downflow avalanches on the core–mantle boundary creates the long wavelength heterogeneity at the bottom of the lower mantle. Models that do not incorporate the endothermic phase change (Tackley *et al.*, 1993) yield a spectrum of thermal heterogeneity unlike the observed spectrum of seismic heterogeneity (Su *et al.*, 1992).

We have seen in Fig. 5 that in transforming from perovskite to spinel mantle, plumes receive a strong input of heat that can raise plume temperature by 100 K. When such a plume arrives at the base of the lithosphere its effectiveness in thinning the lithosphere, producing partial melts due to pressure release melting in the plume, and melting the material of the lower lithosphere, is enhanced. This conclusion rests on the assumption, valid in the calculations of Fig. 5, that the plume surroundings are not also heated by upflow through the endothermic phase change. If a plume is part of a broader large-scale upwelling, then its surroundings would also be heated by passage through the phase transition and the thermal anomaly of the plume relative to its surroundings would be reduced. In the chaotic convective regime of the Earth's mantle it can be expected that plumes will impinge on the phase transition and enter into regions of the upper mantle that are both relatively hot and cold owing to previous episodes of upwelling and downwelling. Therefore the importance of plume heating by the phase change in determining the magmatic and volcanic consequences of a plume's arrival near the surface will depend on the state of the upper mantle at the time. Stated differently, if the mantle in the vicinity of the phase transition were adiabatic, then the latent heat would be included in the adiabat, and a plume rising through the transition would not be heated more than its surroundings. However, owing to the accumulation of cold downflows above the phase transition and the episodic flushing of this material into the lower mantle, the mantle in the vicinity of the phase transition may be nonadiabatic.

The volumes of hotspot swells have been used to estimate the heat flux coming from the Earth's core on the assumption that the heat content of mantle plumes (reflected in the swell volume) derives from the core (Davies, 1988; Sleep, 1990; Davies and Richards, 1992). However, if a substantial fraction of the plume thermal content comes from the phase change heat pulse then the inferred core heat flow would be an overestimate. Use of hotspot swell volume to infer core heat flow faces another problem revealed by the numerical calculations discussed above. There is a large component of conductive heat loss from the core of the model that goes into the reheating of avalanche material that blankets the core; this heat loss from the core is not carried to the surface by mantle plumes which initially derive their hot material from those parts of the core–mantle boundary covered by a hot thermal boundary layer.

Plumes are not the only way to bring anomalously hot mantle material to the base of the lithosphere. The occurrence of a mantle avalanche can have the same effect. When an avalanche takes place, a large amount of cold material that was piled up on the 660 km phase change is rapidly flushed into the lower mantle and is replaced by hotter mantle material from below. While anomalously cold upper mantle material sinks into the lower mantle during an avalanche, anomalously hot lower mantle material rises into the upper mantle to conserve mass. Though an avalanche is fundamentally a downflow phenomenon, its consequence for the upper mantle can be analogous to the arrival of a lower mantle plume, i.e. the inflow of anomalously hot material. Thus it is possible that major tectonothermal events normally associated with mantle plumes, e.g. the massive outpourings of continental flood basalts, could instead be caused by avalanche events in the mantle.

We have not yet discussed the effects of the other major phase change of the mantle transition zone, i.e., the olivine–spinel phase change at a depth of about 400 km. This phase change is an exothermic one whose dynamical effect due to thermally induced phase boundary distortion is to promote convection across it (Schubert *et al.*, 1975). A mantle plume rising through the 400 km phase change would be cooled by the absorption of latent heat. Thus the 400 km phase change acts to mitigate against the dynamical and thermal effects of the 660 km phase change. However, the magnitudes of the Clapeyron slope and the density change are smaller for the 400 km phase change than the 660 km phase change which dominates the dynamics and thermics. Numerical models of the combined effects of both phase transitions on mantle convection have been carried out by Zhao *et al.* (1992), Steinbach and Yuen (1992), Solheim and Peltier (1994), and Tackley *et al.* (1994). The calculations of Tackley *et al.* (1994) show that the 400 km phase change reduces the overall amount of material pile-up on the 660 km phase change, by pushing material through this phase change, resulting in smaller and more frequent avalanches with a broader range of morphologies.

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