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Punctuated tectonic evolution of the earth

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Abstract

The potential of a phase transformation barrier to cause mantle layering has been incorporated into calculations of the thermal evolution of the earth's mantle based on parameterised convection theory. A range of possible behaviors is demonstrated, depending on parameter values, including episodic layering, long-term layering or no layering. Novel findings are 1–2 Ga phases that might correspond with major tectonic eras, and that early mantle overturns may have caused global magmatic and tectonic convulsions. For the more plausible parameter values, the models are initially layered, but typically the layering becomes unstable and breaks down episodically via mantle overturns. Subsequently the models evolve into whole-mantle convection due to the increasing ability of subducted plates to penetrate the phase barrier as the mantle cools, consistent with geophysical evidence against strong layering of the present mantle. The early layering-overturn cycles may occur on timescales of a few hundred million years. The overturns replace cooler upper mantle material with hotter lower mantle material, and would cause global convulsions that potentially correspond with episodes of crust formation. Such models permit plate tectonics to operate in the Archean between overturns, though the early crust may record mainly the effects of overturns, which could generate, aggregate and/or rework large volumes of mafic crust in a short time. They would help to explain present degrees of depletion of the mantle in incompatible elements as well as strong upper mantle depletion in the early Archean. The early convulsions may have controlled the composition of the atmosphere and frustrated the development of life.

1. Introduction

Our present picture of the earth's tectonic history, derived from many decades of geological and geochronological observation of the continental crust, is one with three major eras, each punctuated by briefer episodes of tectonic activity [1,2]. The first era, the Archean, terminated with what seems to have been the most strongly recorded tectonic pulse, peaking at about 2.7 Ga, and its tectonic style is distinctly different from that of the later phases. The 2.7 Ga episode may have been global, though this is not definitively established. The middle, Proterozoic era seems to be transitional, but it has some distinct tectonic and geochemical differences from both the first and third eras [3,4]. The earliest substantial Proterozoic activity occurrea around 2.1 Ga [5], with a strong pulse, possibly global, at about 1.9 Ga, followed by less distinct, possibly only regional pulses [6,7]. The Phanerozoic era is defined in terms of metazoan fossils, but its tectonic style and geochemical signatures seem to be dominated by the effects of plate tectonics, with mantle plumes playing an important subsidiary role. The Phanerozoic features regional episodes, possibly with some global waxing and waning, but with less impession of strong global pulses. Some of the Phanerozoic and possibly Proterozoic pulses or phases of activity

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have probably been associated with the aggregation and breakup of supercontinents [8–10].

Although the causes of these tectonic episodes and eras have been the subject of a great deal of discussion, much of it has remained in the realm of unquantified conjecture. In particular, we don't know whether the Archean tectonic regime was a plate-tectonic regime (slightly or strongly modified) or something quite different [11], we don't know why the Archean had such a pronounced climax followed by an apparent substantial tectonic minimum lasting about 400 Ma, the nature and timing of the transition into modern-style plate tectonics is necessarily indistinct [12], and we are not sure how important mantle plumes were before the Phanerozoic [13]. We have, until recently, lacked a well-defined and plausible mechanism that might account for the briefer pulses. It has not even been clear whether the age-clustering of continental crust reflects actual tectonic pulsing of the earth or is an artifact of uneven preservation or limited sampling [14].

Recently it has been demonstrated that the existence of a pressure-induced phase transformation with a negative Clapeyron slope (the temperature derivative of the transformation pressure) can induce episodic layering in constant-viscosity convection models, alternating with overturns in which the layering breaks down [15–17]. It has been suggested that such behavior of the mantle could account for tectonic pulses, due to the negative Clapeyron slope of the transformation of the spinel structure to perovskite plus magnesiowustite at a depth of 660 km. On the other hand, more recent models incorporating high-viscosity plates show that an old, thick subducting plate can penetrate such a phase barrier more easily than constant-viscosity downwellings, and that older plates plausibly can penetrate into the lower mantle at present [18,19]. Recent models have shown as well that plume heads and strong plume tails can probably penetrate a phase barrier [19,20]. Such plate or plume penetration would tend to break down or prevent layering and its associated pulsing.

There is good evidence that at least some subducted plates do penetrate into the lower mantle at present [21–23], the tomographic results of Grand [21] being perhaps the most convincing. There is also strong evidence that the present mantle is not thermally layered, otherwise there would be strong, buoyant upwellings from the transition zone that would produce surface swells rivalling the mid-ocean ridge system in extent and amplitude [24,25]. Neither the observed hotspot swells nor the low-amplitude deviations from the square-root-of-age seafloor subsidence, predicted by the cooling boundary layer model of plates, are large enough to represent such upwellings [24,26].

These observations and results indicate on the one hand that the present mantle is not significantly layered, with substantial fluxes of mass and advected heat passing through the transition zone, and on the other hand that layering would have been more likely in the past, when the mantle was hotter and plates as a consequence were younger, thinner, and less likely to penetrate the transition zone.

This paper reports results of a quantitative exploration of these possibilities using a thermal evolution model incorporating parameterised (boundary layer theory) convection and two criteria for when layering may break down. The resulting models yield not just the pulses that would be expected, but longer phases of behavior as well. More generally, these models demonstrate a repertoire of behavior that is implicit in our present understanding of mantle dynamics, and as such they represent physically-based, quantified, testable conjectures about the tectonic evolution of the earth.

2. Modelling approach

In the present simple treatment, each mantle layer is represented by one characteristic temperature. If neither criterion for the breakdown of layering is satisfied, the upper mantle is assumed to convect separately from the lower mantle, with some conductive heat transfer across their common boundary, as illustrated in Fig. 1a by a numerical model. If the layering breaks down, the cooler upper mantle material is assumed to sink into the lower mantle and to be replaced by hotter lower mantle material, while the upper mantle material is assumed to mix with the remaining lower mantle material. Fig. 1b shows a numerical illustration of this type of overturn.

Radioactive heating is assumed to occur in both layers. The upper mantle is cooled strongly by heat loss through the earth's surface and heated weakly

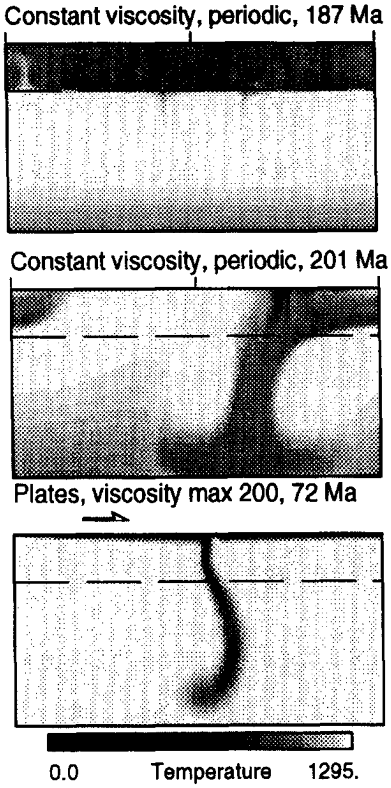

Fig. 1. Numerical models illustrating possible modes of mantle convection in the presence of a phase transformation barrier. (a) A constant viscosity model that has become layered. (b) An overturn event from the same model, in which material has breached the phase barrier. Virtually all of the cooled upper layer material falls through and is replaced by hotter material from the lower layer. (c) Convection with a high viscosity, mobile plate. The greater thickness and strength of the subducted plate has allowed it to break through the phase barrier immediately. This model does not become layered.

by contact with the lower mantle. The lower mantle is cooled weakly by this contact. The upper mantle convection is assumed to involve plates as an integral, driving component, at least at cooler temperatures, consistent with the plates at present comprising the upper thermal boundary layer of the mantle [25].

Layering is assumed to break down if a critical temperature difference between the layers is reached (based on a local internal Rayleigh number), or if the upper mantle is cold enough and consequently plates old and thick enough that they can penetrate through the phase barrier, in the manner illustrated in Fig. 1c.

Restriction of the mass exchange between layers to complete overturns is almost certainly too simple. Three-dimensional, constant-viscosity models [16] show more localised exchange between layers, but these models probably exaggerate the effect by not including plates. The mantle may be intermediate between those models and that shown in Fig. 1b, with breakthroughs that are more limited in time and space as well as occasional complete overturns. Thus the occurrence of some regional tectonic events that do not fit the patterns of these models is not necessarily inconsistent with the main point being made here.

Plumes are not included in the formulation, on the grounds that at present they transport only about 10% of the mantle heat budget [24,27], and so they will not exert the primary control on the layer temperatures. It turns out in these calculations that plumes were probably less important in the past, and so neither are they considered here as potential triggers of overturns.

3. Model formulation

3.1. Thermal evolution equations

The formulation is adapted from that used by Davies [28] to model the coupled thermal evolution of the mantle and core. It is only necessary to modify it to the case of the upper mantle and lower mantle with different heat sources and sinks. The boundary layer theory of convection (sometimes called parameterised convection) replaces full convection calculations with an approximate expression relating the convective heat transport to the temperature difference between the surface and the interior of the fluid and to the viscosity of the interior fluid. The rate of change of temperature of the fluid layer is simply related to the heat lost or gained by convection and the heat gained from radioactivity. The equations governing the evolution of the temperature of each layer between overturns are:

$$\frac{\partial T_u}{\partial t} = \frac{H}{C_m} + \frac{Q_{tz} - Q_s}{M_u C_m} \tag{1}$$

$$\frac{\partial T_l}{\partial t} = \frac{H}{C_m} - \frac{Q_{tz}}{M_l C_m} \tag{2}$$

where T is temperature, subscripts u and l denote upper and lower mantle, respectively, t is time, H is the rate of radioactive heating per unit mass, C_m is the specific heat of the mantle material, Q_{tz} is the rate of heat transfer across the transition zone from the lower to the upper mantle, Q_s is the rate of heat

Table 1
Input values for thermal evolution calculations

Symbol	Quantity or expression	Value
H_0	present radioactive heat generation rate	30 TW
C_m	specific heat of mantle material	1000 J/kgK
M_u	mass of upper mantle	1.3×10^{24} kg
M_l	mass of lower mantle	2.7×10^{24} kg
$\eta_u{}^0$	present upper mantle viscosity	10^{21} Pa s
$\eta_l{}^0$	present lower mantle viscosity	10^{22} Pa s
a_s	$Q_0 \eta_u^{0^{n_s}} / T_r^{1+n_s}$ (equation 3)	6.7×10^{15} (S.I.)
Q_0	present heat flow	36 TW
T_r	reference temperature	1280°C
T_s	surface temperature	300°K
a_z	$K_l(\rho_l \alpha_l g/\kappa_l)^{n_z} 4\pi R^2$ (equation 4)	1.4×10^{17} (S.I.)
K_l	thermal conductivity ($\rho_l C_m \kappa_l$)	4.8 W/m°K
ρ_u	upper mantle density	3500 kg/m^3
ρ_l	lower mantle density	4800 kg/m^3
α_u	thermal expansion (upper mantle)	2.5×10^{-5}K^{-1}
α_l	thermal expansion (lower mantle)	1.5×10^{-5}K^{-1}
g	acceleration due to gravity	10 m/s^2
κ_l	thermal diffusivity	10^{-6} m^2/s
R	radius of the earth	6371 km
E	activation energy	300 kJ/mol
R_g	universal gas constant	8.3 J/mol°K
b_{up}	rheology constant, upper mantle (equation 6)	5.02×10^7 (S.I.)
b_{lo}	rheology constant, lower mantle (equation 6)	5.02×10^8 (S.I.)
$w_s{}^0$	present maximum plate thickness	100 km
$\tau_m{}^0$	present maximum plate age	100 Ma
$\Delta\rho_s$	slab thermal density excess	70 kg/m^3
ΔT_s	slab mean temperature deficit	800°C
$\Delta\rho_p$	density increase through phase transformation	280 kg/m^3
β	Clapeyron slope	-3 MPa/°K
Ra_{loc}	critical local Rayleigh number (equation 11)	20

loss at the earth's surface, and M is the mass of the layer. An expression for H as a function of time is given by Davies [28].

As discussed by Davies [28], the mantle is cooled at present mainly by the formation and subduction of plates, which comprise the upper thermal boundary layer of the convecting mantle. An expression is derived there for this heat loss, and it is put in the form:

$$Q_s = a_s \frac{(T_u - T_s)^{1+n_s}}{\eta_u^{n_s}} \tag{3}$$

where a_s represents a collection of material constants, T_s is the temperature at the surface of the earth, η_u is the viscosity of the upper mantle, and n_s is a constant taken here to be 0.3. The value of a_s is calibrated here from the present earth, such that Eq. 3 yields the present earth heat flow at a mantle reference temperature, T_r, of 1280°C (see Table 1).

The heat transfer between the layers plays an important role in the present calculations, and it requires some discussion. This heat transfer involves two adjacent thermal boundary layers, one at the base of the upper mantle and one at the top of the lower mantle. In general, one should consider both of these boundary layers, and determine the temperature at their interface such that they transport equal amounts of heat. However, the upper one will have a lower viscosity and so may be relatively thin, with a relatively small temperature jump across it. Conversely it is likely that the lower one will have a higher viscosity than either the lower mantle below it (because that is warmer) or the upper mantle above it (because the upper mantle is in lower-pressure phases and at present seems to have a lower viscosity). Consequently it is assumed here that this stiffer boundary layer at the top of the lower mantle will limit the rate of heat transfer, and that the lower-viscosity boundary layer at the base of the upper mantle will be able to adjust rapidly and accordingly. This assumption is not critical, although it affects some details of the results.

This boundary layer at the top of the lower mantle is assumed to behave viscously, and in this respect it has a critical difference from the boundary layer at the surface of the earth, which is brittle, and broken into mobile plates. Convection in a fluid with a temperature-dependent viscosity and cooled from above has been studied experimentally by Davaille and Jaupart [29], who show that the following formula describes the resulting heat transport very well:

$$Q_{lz} = \beta_z a_z \frac{\Delta T_d^{1+n_z}}{\eta_l^{n_z}} \tag{4}$$

where βz is an adjustable parameter with a value near 1, a_z is a collection of constants (Table 1), η_l is the viscosity of the interior of the lower mantle, $n_z = 0.3$ is a constant, and ΔT_d is the temperature difference across that part of the boundary layer that is warm and mobile enough to be involved in driving the convection (the cooler part can become so viscous it is relatively immobile). The measure of the latter used by Davaille and Jaupart is:

$$\Delta T_d = -\frac{\eta}{\partial\eta/\partial T} \tag{5}$$

evaluated at the internal temperature of the layer. With this measure, they found $\beta z = 0.47$ in their experiments.

The temperature-dependence of viscosity is taken here to have the standard thermal activation form:

$$\eta = bT\exp E/R_g T \tag{6}$$

where b is a constant, E is the activation energy and R_g is the gas constant. Substitution of (6) into (5) yields:

$$\Delta T_d = \frac{T}{E/R_g T - 1} \tag{7}$$

and for $T = 1873$ K and $E = 300$ kJ/mol this comes to 102 K. For the calculations presented here, ΔT_d was taken simply to be the smaller of 100°C and $(T_l - T_u)$, with a minimum value of 0.

The viscosities that appear explicitly in Eqs. 3 and 4 are assumed to have the form of Eq. 6. In the absence of accurate data for high-pressure phases, the same activation energy is assumed for the lower mantle as for the upper mantle, and the value $E = 300$ kJ/mol is used here. The types of behavior presented below are not sensitive to this value, although some modest compensating adjusments in other parameters are required to get similar results. For the upper mantle, the value of b is determined by requiring the viscosity to be 10^{21} Pa s at a temperature of

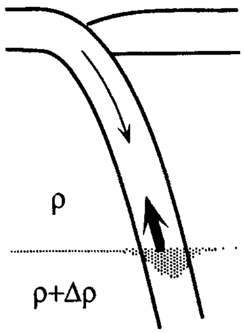

Fig. 2. Illustration of the resistance due to a phase transformation with a negative Clapeyron slope. The transformation occurs at greater pressure and depth in the cool, subducting lithosphere, so that there is a region (shaded) where the less-dense upper layer material persists to greater depth. The buoyancy of this material (broad arrow) opposes the descending motion. Rising hot flows are also resisted.

1280°C (Table 1). The lower mantle is taken to have a viscosity 10 times greater at the same temperature.

3.2. Layering mechanism and breakdown criteria

Layering may be induced by a phase transformation with a negative Clapeyron slope because, for example, the transformation occurs at a greater depth in cool, descending material, thus allowing less dense upper phase material to persist to greater depth, where it is buoyant in comparison with adjacent, warmer, material that has transformed to the denser lower phase (Fig. 2). This buoyancy opposes the descent of the cool material, and may prevent it. Warm upwelling material is also impeded. If the phase transformation buoyancies are strong enough, the convection may be separated into two layers.

It is also likely that some resistance to penetration will derive from the different composition of the oceanic crustal component of subducted lithosphere, within which the transformation to perovskite structure is substantially delayed [30,31]. This mechanism will have been more important in the past, when the hotter mantle would have produced thicker oceanic crust. It may substitute to some degree for the endothermic transition mechanism here, and it might yield other kinds of behavior. Some of these possibilities will be explored elsewhere.

Numerical models suggest two ways in which layering induced by such a phase transformation barrier can break down. The first is if cold material from the surface thermal boundary layer (i.e., the lithosphere) is sufficiently heavy to push through. The second is if material from one of the internal boundary layers at the interface detaches from the interface with sufficient vigour to pull material across the interface. These will be discussed in turn.

Davies [19] and Zhong and Gurnis [18] have shown that a stiff subducted lithospheric plate can penetrate a phase transformation if its Clapeyron slope is more positive than about -4 MPa/K. A key difference from a constant-viscosity downwelling is that more of the stiff slab's weight is brought to bear against the phase transformation resistance because of its stiffness. This will be limited by buckling: Griffiths and Turner [32] have shown that a viscous sheet will buckle if its height is roughly 5 times its thickness (Fig. 3).

Here it is assumed that the height of the slab, h, that couples to the resistance is a multiple, b_u, of its thickness, w_s: $h = b_u w_s$. The maximum thickness of

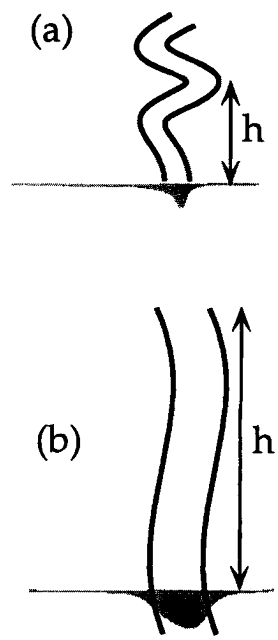

Fig. 3. Sketch of the effect of the slab stiffness on the effective height of the dense slab that is coupled to the phase boundary resistance. The wavelength of slab buckling is approximately proportional to its thickness. Once substantial buckling occurs, the weight of the higher parts of the slab is no longer transmitted through the slab to the phase boundary.

plates is a function of the maximum plate age, τ_m: $w_s = b_t \tau m 1/2$; this in turn is a function of the surface heat flow, Q_s, since $Q_s = a_Q \tau m - 1/2$. (These formulas assume that at any time the age of plates at subduction is distributed uniformly between zero and the maximum age.) The constants b_τ and a_Q in the last two relationships are evaluated from present mantle values listed in Table 1. Combining these yields:

$$h = b_u b_\tau a_Q / Q_s \quad (8)$$

The stress exerted by the heavy slab on the phase boundary is:

$$P_s = g h \Delta\rho_s = g h \rho \alpha \Delta T_s \quad (9)$$

and the resistance from the phase boundary deflection is:

$$P_p = g \Delta\rho_p \beta \Delta T_s / \rho g \quad (10)$$

where g is the acceleration due to gravity, $\Delta\rho_s$ is the thermal density difference between the slab and adjacent mantle, ρ is the density of adjacent mantle, α is the thermal expansion coefficient, ΔT_s is the equivalent mean temperature difference between the slab and adjacent mantle, $\Delta\rho_p$ is the density increase through the phase transformation, and β is the Clapeyron slope of the phase transformation. Required values are listed in Table 1, $\Delta\rho_s$ being calculated from the observed 3 km subsidence of old sea floor relative to zero-age sea floor. The condition for slab penetration and consequent mantle overturn is that $P_s = P_p$, which turns out to define a critical upper mantle temperature below which slab penetration is possible.

The second assumed condition for triggering overturn is that the stiff boundary layer at the top of the lower mantle becomes unstable, drips down and pulls upper mantle material after it (Fig. 4). Solheim and Peltier [17,33] have found that overturns in constant-viscosity models are predicted by a local internal Rayleigh number based on the internal thermal boundary layers, and this is a measure of those boundary layers becoming unstable. In the temperature-dependent rheology assumed here, it will presumably be the stiffer boundary layer at the top of the lower mantle that exerts the greatest stress on the interface (Fig. 4b), as discussed above, and so the

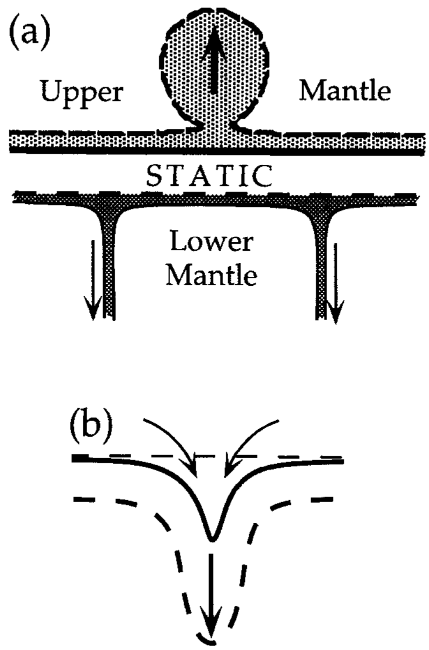

Fig. 4. Sketch of internal thermal boundary layers when flow does not penetrate the phase boundary. The interface (heavy line) has a temperature intermediate between the upper mantle and lower mantle. (a) Above the interface is a hot boundary layer that generates plumes, because of its lower viscosity. Below the interface is a boundary layer that is cooler than the lower mantle. For temperature contrasts of 200°C or more, this will have a viscosity two or more orders of magnitude greater than the lower mantle. As a result, the cooler part of it may be static or very sluggish. The lowest, warmest part will drip off in the manner sketched. (b) The stiff part of the lower boundary layer will eventually become unstable, and it may then pull material through from the upper layer and trigger an overturn.

Rayleigh number criterion is applied here only to this layer. The local Rayleigh number used here is:

$$Ra_{loc} = \frac{g \rho_l \alpha_l D_{tz}^3 (T_l - T_u)}{\kappa_l \eta_u} \quad (11)$$

where D_{tz} is the thickness of the thermal boundary layer, estimated here as being such that the heat flux Q_{tz} will conduct through the interface:

$$D_{tz} = \frac{4\pi R^2 K_l (T_l - T_u)}{Q_{tz}} \quad (12)$$

The criterion for an overturn is that Ra_{loc} reaches a critical value. This critical value is left as an adjustable parameter here.

4. Thermal evolution models

A thermal evolution of the two mantle layers can be calculated from Eqs. 1 and 2 by specifying starting temperatures, integrating forward in time, and monitoring the two criteria for overturns. If an overturn is indicated, the integration is stopped, the layer temperatures are redefined as specified above, and the integration is resumed.

A particular model is illustrated in Fig. 5a, starting with both layers at a temperature of 1940 K (1667°C; these are representative decompressed or potential temperatures). A continuously evolving whole-mantle model is included for reference. Initially the upper mantle cools rapidly and the lower mantle warms. After about 260 Ma, the temperature difference between them (about 240°C) is such that the assumed critical local Rayleigh number of the internal boundary layer is exceeded, and there is an overturn. After the overturn, the upper mantle has the former temperature of the lower mantle, and the lower mantle has the mean temperature of the former upper layer and the balance of the lower layer.

This sequence repeats four times in this model, with increasing intervals between overturns, the last of these overturns being 2.71 Ga before present. There is an overall cooling through this sequence. If there were only the one criterion for overturns, there would be no more in this model: because the radioactive heating is decaying, the lower mantle would never again become hot enough to trigger an overturn in this way.

However, at about 2 Ga before present the upper mantle becomes cool enough so that subducted plates would be old and thick enough to punch through into the lower mantle, and another overturn is triggered in this way. There follows a second sequence of overturns triggered by plate penetration. These occur after successively shorter intervals, because after each one the upper mantle cools to the critical temperature sooner, mainly because now the peak temperature of the lower mantle is substantially reduced with each overturn. This sequence is terminated at about 700 Ma in this model, the lower mantle temperature having been reduced to within a few degrees of the upper mantle temperature.

At this stage, plates being able to penetrate readily into the lower mantle, the convection will become effectively unlayered, and the thermal evolution is continued to the present as whole mantle convection.

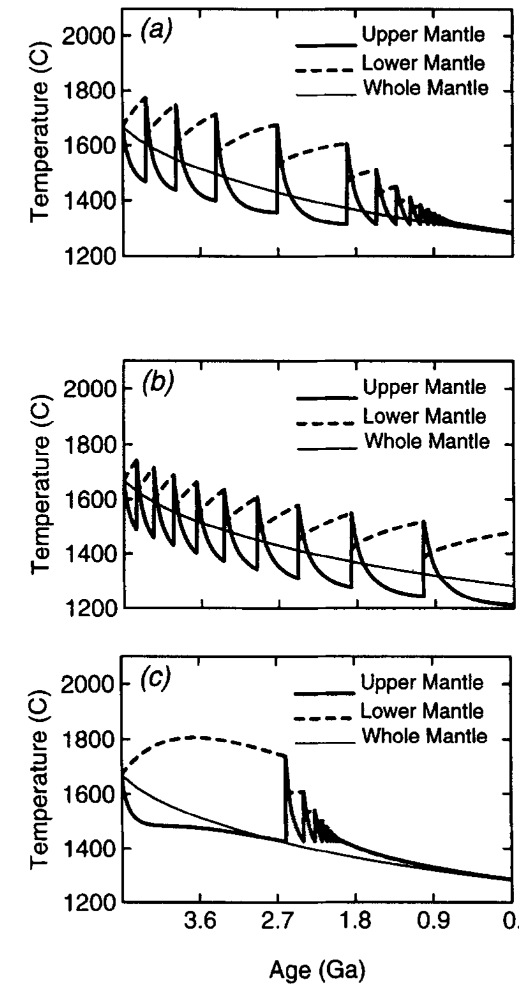

Fig. 5. Illustrative thermal evolution models. (a) Preferred model, with early episodic layering due to a phase change barrier. The first four overturns are triggered by the instability of the internal boundary layer. The subsequent "Proterozoic" overturns are caused by penetration of subducted plates through the mantle phase barrier. The layering is broken down by 0.7 Ga, after which whole mantle convection applies. (b) Heat transfer between the layers is less efficient in this model ($\beta_z = 0.47$), so the early overturns are more frequent. They continue to the present because plate penetration has been artificially inhibited. (c) With more efficient heat transfer between the layers, no overturns occur until plates begin to penetrate, in this case at a relatively early time. Parameters are given in Tables 1 and 2.

Fig. 5b shows a model in which there have been two changes relative to Fig. 5a. In the first model, the value of β_z, which governs the efficiency of heat transfer between the layers (Eq. 4), was 0.60, slightly higher than the value of 0.47 found by Davaille and Jaupart [29] in their experiments. If their lower value is used, the early overturns occur more frequently, because with less efficient heat transfer the layer temperatures diverge more rapidly. The second change in Fig. 5b is that ability of subducted plates to penetrate the transition zone has been greatly reduced, by the artifice of reducing the value the buckling parameter to $b_u = 0.5$ Table 2. As a result, the second sequence of overturns caused by plate penetration never occurs. We then see that with $\beta_z = 0.47$, the first series of overturns continues, with the most recent one occurring at about 1 Ga. In this model the present mantle is layered.

A third model is shown in Fig. 5c in which the efficiency of heat transfer is larger than in Fig. 5a ($\beta_z = 0.80$). This prevents the temperature difference between the layers (or rather, the local Rayleigh number, Eq. 11) from reaching the critical value at which overturn occurs. In this case there are no early overturns. However, plate penetration begins fairly early, at 2.6 Ga (from choosing $b_u = 5.5$, which makes the plates more able to penetrate), and the resulting sequence of overturns evolves into whole mantle convection at 1.9 Ga.

These examples illustrate some general features of solutions to the governing equations. The occurrence and frequency of the early overturn sequence depends on the efficiency of heat transfer between the layers. With higher efficiency, the overturns are less frequent and they may terminate early or not occur at all (Fig. 5c). With lower efficiency they occur more frequently and may continue to the present (Fig. 5b).

Table 2
Parameters for the cases shown in Fig. 5

Case	β_z (eq. 4)	b_u (eq. 8)
(a)	0.60	3.25
(b)	0.47	0.5
(c)	0.80	5.5

The occurrence and timing of the second overturn sequence depends on the ease with which subducted plates can penetrate the phase transformation barrier. If they can penetrate more easily, then penetration begins earlier while plates are younger and thinner. The duration of the resulting second sequence depends on the upper mantle temperature at which penetration occurs, among other things, but is of the order of 1 Ga. The ease of penetration depends on a combination of parameters (Eqs. 8–10) of which the Clapeyron slope, β and the buckling parameter, b_u, are the most important. These two parameters can be traded against each other. It is not possible to defer plate penetration later than about 1.5 Ga without these parameter values becoming implausible (Fig. 5b is not meant to be realistic, but to illustrate the model behavior).

Another important parameter is the critical value of the local Rayleigh number that determines when a phase one overturn occurs. If this is smaller, they occur more often. The frequency of these overturns is not well determined in the model.

The earth would not be expected to be as regular as this model. The overturn triggers would be somewhat variable, so that less regular pulsing is likely. Thus the exact timings of particular pulses in this model are not of great significance. In this model the episodes are intrinsically global, but in the earth the later, weaker episodes could be expected to be more localised.

4.1. Notable and robust features of the models

The modelling has shown that some potentially important modes of behavior are implicit in our current understanding of mantle dynamics. A novel feature is the long-term (billion-year) phases that show some encouraging resemblance to the major tectonic eras of the earth. The models also show that episodic behavior with a timescale of a few hundred million years is plausible in the early part of earth history. They show that layering, episodic or not, induced by mantle phase changes is unlikely to persist to the present.

The details and detailed timing of these behaviors are not robust, as they are sensitive to parameters that are not yet well constrained. Thus the occurrence, frequency and duration of episodic layering is sensitive to the efficiency of heat transfer between the mantle layers, and the timing of initial plate penetration leading into whole-mantle convection is sensitive to the thermodynamic parameters of mantle phase changes and to details of the dynamics of subducted plates.

Nevertheless it is notable that a major pulse plausibly can occur at about 2.7 Ga, corresponding to the terminal Archean event, and that another major pulse can occur at about 1.9 Ga, with declining activity thereafter (Fig. 5a). It would have been more gratifying if the 2.7 Ga pulse were the strongest in the sequence, since the terminal Archean event is the most strongly recorded in the crust, but exploration of this type of model is only just beginning and the connection between the putative mantle events in this model and the creation, consolidation and preservation of continental crust is not simple.

Although the details of the early episodes are not robust, the model of Fig. 5a shows a general consistency with observed clustering of Archean ages, particularly around 2.7, 3.5 and 3.8 Ga. There is also some general consistency between the 200–300 Ma spacing of the later overturns and observed Proterozoic age clustering.

The cooler upper mantle of the episodically layered models (Fig. 5a and b) would permit modern-style plate tectonics to develop much earlier than in the reference whole mantle model. (Davies [12] argued that it was not clear that plate tectonics could operate if the mantle were as little as 50°C hotter than at present, since plates would be faster and thinner, while the oceanic crust would be thicker, and both factors would tend to make plates more buoyant and less subductable.) Also a cooler upper mantle would yield basalts, rather than higher-temperature picrites and komatiites, as the dominant mantle melt product, as is observed in greenstone belts. On the other hand, neither of these things would be true during overturns, and the overturns might have left the dominant imprint in the continental crust.

5. Potential implications

The modelling has demonstrated several possible tectonic modes that should be considered for pre-Phanerozoic earth history. This discussion is conjectural, and both the apparent strengths and apparent weaknesses of the modelling are subject to revision.

5.1. Archean mantle overturns—geological effects

The relacement of the upper mantle with lower mantle material some 200–300°C hotter on a time scale probably of the order of 10 Ma would undoubtedly have dramatic magmatic and tectonic consequences. Some idea of the possibilities can be obtained from a comparison with flood basalts. A flood basalt covers an area about 2000 km across, or about 3.10^6 km^2, while the area of the earth's surface is about 5.10^8 km^2, a ratio of about 170. A flood basalt produces 10^6–10^7 km^3 of erupted basalt [34,35]. If rising hot lower mantle material melted by an average of 3% in a main melt zone about 100 km thick, it would cover the earth with a 3 km layer of basalt, with a volume of $1.5.10^9$ km^3, over 150 times the volume of a flood basalt. The precise numbers are not important here: it is clear that a mantle overturn event could involve volcanic episodes two or more orders of magnitude greater than flood basalts.

The oceans would probably have had plate tectonics operating, at 2–3 times its present rate, prior to an overturn. There would be large topographic changes during an overturn, in the range of 3–5 km. Such elevation differences would occur over distances of less than a thousand kilometers, and would propagate across the earth's surface at rates possibly of the order of 1 m/a. These alone would cause large changes in motions of pre-existing plates. Also the plates may have speeded up if lower-viscosity hot mantle arrived under them, or if it laterally displaced them together with the cooler, stiffer upper mantle material already under them.

There would be a transitional phase during which pre-existing spreading centers would have produced greater thicknesses of oceanic crust (20–40 km thick, by analogy with oceanic flood basalt provinces) as hotter mantle arrived beneath them (Fig. 6a). This phase would end as the thicker oceanic crust arrived at subduction zones (Fig. 6b), where it would proba-

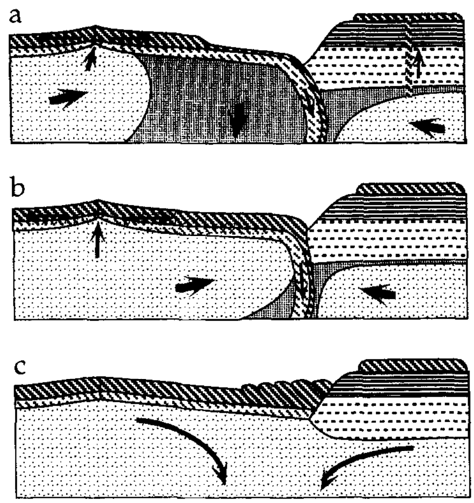

Fig. 6. Sketch of possible tectonic events accompanying a mantle overturn. (a) As hot material from the lower mantle (light shading) replaces cooler upper mantle material, it will melt. At pre-existing spreading centers, it will produce thicker oceanic crust. In continental areas, it may yield eruptions and intrusions resembling continental flood basalts. (b) Pre-existing plates will continue to subduct until the thicker oceanic crust reaches subduction zones at which time subduction may cease because of the greater buoyancy of the thick crust. (c) Subduction may be prolonged by tectonic thickening of oceanic crust, after which a different mode of mantle convection may prevail, as suggested by Davies [12].

bly be too buoyant to subduct. It might be prolonged for a while by further tectonic thickening of the thick oceanic crust (Fig. 6c).

After this transition phase, modern-style plate tectonics would probably cease or be minor, and the earth would revert to an earlier tectonic mode appropriate to a hotter mantle. This might involve the detachment and "dripping" of sub-crustal oceanic lithosphere, as suggested by Davies [12] and Campbell and Griffiths [36], or some other mode adapted to the presence of larger degrees of mantle melting and thicker, buoyant mafic crust. This phase would persist until the upper mantle cooled enough to permit plate tectonics to become important again, perhaps 100–200 Ma later according to the model shown in Fig. 5a.

This sequence of oceanic events might thus produce a large volume of mafic crust in a short time (a few tens of millions of years) and accumulate it into broad, thick piles. Such a terrain might resemble oceanic plateaus or ophiolite complexes in some respects and island arcs in other respects. Such resemblances have been noted in many Archean greenstone belts.

As with flood basalts [13], the volume of eruption would be expected to vary substantially between different tectonic regions. Continental areas presumably did not have thick lithospheric keels at this time (otherwise they would not record strong tectonic events from this time), so the expression might resemble flood basalts on relatively young continental crust (Fig. 6).

Campbell and Hill [37] have shown that in the Kalgoorlie–Norseman area of Western Australia the oldest granitoids associated with greenstone formations are 15–20 Ma younger than the greenstones, and they have proposed that the granitoids represent crustal reworking caused by the same mantle thermal event that produced the mafic greenstone rocks slightly earlier. They proposed that the mantle thermal event was the arrival of a plume head. In general terms a mantle overturn might yield a similar sequence of new mafic crust closely followed by reworking. (However, it is not clear at this stage that the stratigraphy and geochemistry can be explained as well by the overturn model as by the plume head model [13,37].) Alternatively, reworking of new mafic crust might occur predominantly during a subsequent overturn.

5.2. The early layered mantle

During periods of layering, plates would probably operate much like they do at present once the upper mantle had cooled sufficiently for them to subduct (Fig. 7a). There is no clear argument that they would necessarily be smaller just because they were part of an upper mantle system rather than a whole mantle system. At present it seems very likely that the geometry and kinematics of plates is controlled by the mechanics of the lithosphere rather than of the underlying mantle Davies [25,38]: for example the three types of plate boundary correspond with the three types of faults well-known to structural geologists. If plate strength is important at present, and accounts for the survival of large plates, then presumably plates about as large might have existed during layered intervals in the Archean.

In the model of Fig. 5a the heat flux into the base

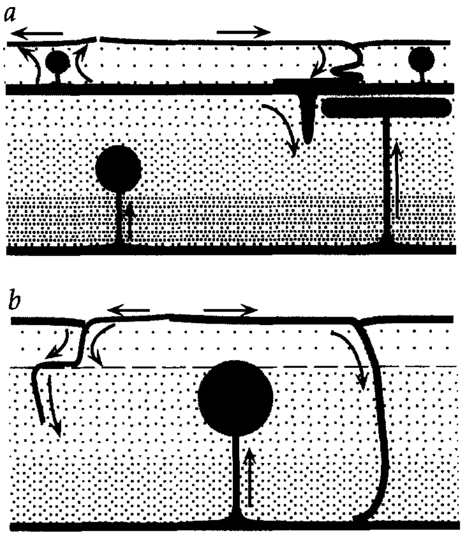

Fig. 7. (a) Sketch of mantle convection during layered periods. Plate tectonics would operate much like the present, cooling the upper mantle. Small plumes would rise from the transition zone. Cool drips would fall from the transition zone into the lower mantle, while large plumes would rise from the base of the mantle, much like present plumes except that they would be blocked at the transition zone. (b) Sketch of the likely behavior of the present mantle. Plates and plumes both seem to be able to penetrate the transition zone, though some plates, especially young, thin ones, may be temporarily delayed.

of the upper mantle is substantial (comparable to the present heat flux at the earth's surface). This would have generated a quite strong system of upwellings that would have had the head-and-tail form of modern plumes because of the lower viscosity of the hotter material (Fig. 7a). However, the heads of these plumes would have been smaller (200–300 km diameter; [39]) than those of plumes rising from the base of the mantle (800–1200 km diameter; [40]).

A separate lower mantle convection system would have existed, with large plumes similar to those inferred to exist at present, and downwellings from a cooler boundary layer at the top of the lower mantle (Fig. 7a). These downwellings would be ductile "drips" if the rheology of this material was not brittle, as seems likely.

The lower mantle may have had significant thermal stratification, due to the episodic dumping of cooler upper mantle material. Such stratification is visible in the numerical model of Fig. 1b and is included schematically in Fig. 7a. It could have important consequences for trace element and isotope chemistry.

5.3. A tectonic minimum?

There is little record of substantial crustal formation or reworking from the period between the late Archean (roughly 2.6–2.5 Ga) and about 2.1 Ga (the Birrimian terrain of west Africa [5]). Condie [41] argues for a minimum in the formation of "oceanic greenstones" between 2.5 and 2.2 Ga. There was substantial activity at about 2.5 Ga in South Australia and formerly adjacent parts of Antarctica [42], but the global significance of this activity is not clear at present. There are also mafic dykes and complexes dating from 2.3–2.5 Ga in many areas, but these alone do not connote major crustal tectonic events. Whether the apparent minimum or hiatus is due to an actual absence of activity or to the vagaries of imperfect sampling or preservation is likely to remain uncertain. The model of Fig. 5a permits this gap in the record to be interpreted as a real hiatus in overturn activity, though plate tectonics would be expected to have occurred throughout this period according to this model.

5.4. The Proterozoic

The record of major orogenies in the Proterozoic opens with a regional episode in West Africa (the Birrimian, 2.1 Ga: [5,43]), features a major series of episodes at about 1.9 Ga, and later episodes less distinctly clustered, with a secondary peak around 1.1 Ga [6,9,44]. These episodes are of varying character. The period 1.95–1.8 Ga saw the rapid assembly of many Archean blocks to form northern Laurentia (North America and Fennoscandia; [6]). A number of episodes involved the addition of substantial amounts of young crust of oceanic affinity: the earlier Birrimian, several terrains in northern and central Laurentia around 1.9 Ga, and large areas of southern Laurentia between 1.6 and 1.8 Ga (Yavapai and Mazatlal). Some episodes were mainly intracratonic: the 1.88–1.84 Ga Barramundi of northern Australia and the "anorogenic magmatism" of central and southern U.S.A., 1.3–1.5 Ga. The Laurentian Grenville orogen is viewed by some authors as part of a global collisional series that formed a supercontinent in the period 1.3–1.1 Ga [45].

The models presented here permit alternative hypotheses about Proterozoic tectonic regimes. Hoffman [6,9] and Condie [41] have argued that peaks in crust ages are due to breakup and aggregation of supercontinents. This would be consistent with the model of Fig. 5c, in which plate penetration is presumed to start early and the main phase of overturns is complete by the early Proterozoic. However, this model is less obviously consistent with a hiatus in overturns.

An alternative hypothesis is that the Proterozoic corresponds to the period of the second phase of mantle overturns, during which plate penetration breaks down the layering, as in Fig. 5a. This model is consistent with an overturn hiatus. However, the first "Proterozoic" overturn is nearly as intense as the last Archean overturn. This tends to be a robust feature of models with a substantial hiatus, since the layer temperatures diverge substantially during the hiatus. With this caveat, this model suggests that the early Proterozoic events were mantle overturns, and that later overturns became less intense, more frequent and more regional in nature, rather than global. As the overturns died out, modern plate tectonics, moderated by continental dispersal and aggregation, would have become dominant.

Several early Proterozoic events are thought to have been quite brief, with the main tectonic and felsic magmatic events lasting less than 50 Ma [7,43,46], although in northern Australia a more extended sequence of events is becoming evident [47]. The brevity of these events may reflect faster plates at the time [46], or it may reflect the rapid jostling that might accompany a mantle overturn. The Birrimian (and others, especially the Yavapai and Mazatlal events) seems to resemble the oceanic part of the scenario in Fig. 6, in which thickened oceanic crust is rapidly produced, aggregated and reworked. The Barramundi could represent a continental analogue, involving mafic underplating, partial rifting and later compression. The Wopmay orogen of northern Canada [48] seems to be intermediate in character between the other two. The elucidation of these possibilities will be a complex task that this brief discussion cannot do more than point the way to.

5.5. The Phanerozoic

The Phanerozoic, as the name implies, is defined in terms of the fossil record, and so we should not expect it necessarily to have a distinctive tectonic style. Nevertheless the whole-mantle convection phase of the models do correspond loosely to this period. In the models, the tectonic transition from episodic overturns would be gradual, and manifest as a progressive decline of non-plate tectonic events. Plate tectonics would be operating between major overturns, and in regions unaffected by regional upwelling events. The signatures of plate tectonics would become progressively more dominant.

The so-called "Pan-African" orogeny of the late Proterozoic and early Phanerozoic may be a candidate for a regional breakthrough event. According to El-Gaby and Greiling [49] there are at least three separate episodes involved. Of these, perhaps the "volcanic arcs and associated ophiolitic melanges" of the Arabian–Nubian shield are the most plausible, with the deformation and reworking in the Mozambique belt possibly associated. Distinguishing a regional breakthrough from a plume head event may not be easy. Hoffman [9] interprets the Arabian–Nubian shield accumulation and the Mozambique belt as a continental collision event subsequent to the breakup of the Rodinia supercontinent: these interpretations are not mutually exclusive.

Stein and Hofmann [50] suggested that Mesozoic "superplume" activity in the Pacific and other regions [51] represent the latest episode of mantle overturning. However, the buoyancy and heat fluxes associated with these events have been estimated from the magnitude of hotspot swells and the Darwin Rise superswell, and they are minor, corresponding to less than 2% of the earth's heat budget [26]. For comparison, the Hawaiian plume transports about 1% of the earth's heat, while all plumes transport less than 10% [24,27]. The Mesozoic events more plausibly correspond to a regional breakthrough of some plume material temporarily blocked by the transition zone.

5.6. The present mantle

The present mantle seems to be quite close to the conditions in which layering would occur [19,25]. In the models of Fig. 5a and c this is not a coincidence, it is a consequence of the mantle having relatively recently been in the episodically layered mode, and not having evolved far from that state. The model of Fig. 5b, which is layered at present, is not consistent with other evidence noted in the Introduction. The present mantle system is represented schematically in Fig. 7b. Old, thick plates may penetrate the transition zone readily, while younger, thinner plates may penetrate after some short or long delay [19,23]. Plumes probably penetrate regularly, but some plume tails may be choked off [19]. There may also be some residual thermal stratification in the lower mantle, associated with past overturns, with associated trace element and isotopic anomalies.

5.7. Plumes from the base of the mantle

The effects of mantle plumes have not been included in this model for reasons noted earlier. It is likely that most plumes at present come from the base of the mantle [24,25,40], driven by heat conducting out of the core. In the present model, the lower mantle is quite hot for much of earth history, and so the temperature difference between it and the core would be smaller (Davies [28] calculated that the core may have cooled by only 300°C or less through earth history). Consequently less heat would conduct out of the core and plumes from there would be less important. From Fig. 5 we can infer that plumes from the base of the mantle might have been significant immediately after mantle overturns, but that otherwise they would increase in importance through the Proterozoic, approaching their present level only within the last billion years or so.

5.8. Isotopes and trace elements

The existence of long-lived (1–2 Ga) isotopic heterogeneity in the mantle has featured prominently in the debate as to whether the present mantle is layered or not, as have estimates of the fraction of the mantle depleted to form the continental crust. Galer et al. [52] have argued that a flux of less-depleted material into the upper mantle is required, so that the fraction of mantle depleted is greater than the volume of the upper mantle. Davies [53] and Christensen and Hofmann [54] have argued that the longevity of heterogeneities may be explainable in a whole-mantle convection system by invoking the higher viscosity of the lower mantle or the tendency of denser recycled oceanic crust to accumulate near the base of the mantle. Recently Stein and Hofmann [50] have suggested that episodic mantle layering would help to account for the geochemical constraints. Although they envisaged episodic layering continuing to the present, their arguments apply in general to the present model: the past episodic layering and accompanying moderate thermal stratification of the lower mantle would inhibit depletion of the deeper mantle, enhance the longevity of heterogeneities, and provide episodic replenishment of the upper mantle. It may also help to account for the moderately depleted "PREMA" mantle source type proposed by Stein and Hofmann to have been the source of large segments of new crust formed in the past three billion years. The PREMA source is inferred to reside mainly in the lower mantle. All of these things remain to be quantitatively evaluated.

McCulloch and Bennett [2] have argued on the basis of values of ϵ_{Nd} from the early Archean as high as $+4$ that only a fraction of the upper mantle was depleted of lithophile incompatible elements early in earth history, with this fraction later increasing episodically. However, this conclusion depends on estimates of the amount of continental crust (or other enriched reservoir) present at the time. The present model partially accommodates this argument, by partially isolating the upper mantle in earliest earth history.

5.9. Life and the atmosphere

Flood basalt eruptions have been proposed as causes of or contributors to mass extinctions at the end of the Permian and Cretaceous eras, and possibly at other times [55,56]. If mantle overturn events were two orders of magnitude greater than these, and were accompanied by geologically very rapid elevation changes that would have disrupted shallow seas, then it seems possible that even the relatively simple life forms present during the Archean and early Proterozoic could have been severely threatened. The decline of intense overturn events in the present model corresponds roughly with the emergence of eukaryotic life in the early Proterozoic and multicellular life during the later Proterozoic [57], suggesting that tectonic convulsions may have frustrated the development of life in earlier times.

The composition of the atmosphere may have been significantly affected by overturn events, especially if the Archean mantle was less degassed than the present mantle. For example, a conservative estimate for the concentration of CO_2 is the value for the present MORB source, about 100 $\mu g/g$ [58]. If this concentration is scavenged from a layer of mantle 100 km thick, the total mass of CO_2 released would be about 2.10^{18} kg, compared with the present total mass of the atmosphere of 5.10^{18} kg. The amount of CO_2 released is quite uncertain, and could have been several times this, since mantle concentrations are uncertain by at least a factor of two, would probably have been higher in the Archean, and a greater thickness of mantle may have been scavenged.

6. Discussion

The models of mantle evolution presented here demonstrate some of the possibilities that are implicit in our present understanding of mantle dynamics. They are not the only possiblities. It is possible, for example, to construct a model in which the mantle stratifies permanently, but the models of Fig. 5a and c are among the ones judged to be the most consistent with observational and experimental constraints, with that of Fig. 5a being preferred at this point. Some of the other possibilities and uncertainties will be demonstrated elsewhere.

The models show not only that episodic mantle overturns are possible according to our present understanding, as has been frequently speculated lately, but also that longer phases of behavior may result from the same mechanisms, and that these may correspond with tectonic eras of the earth. This has not been evident until now, and has only been revealed with the aid of quantitative modelling. Although conjectural, the models are quantitative, based on relevant physics, and testable in many respects. More generally they demostrate that our understanding of mantle dynamics has matured to a point that it can yield stimulating hypotheses about the evolution of the earth.

The ability of models such as those presented here to explain the tectonic evolution of the crust and the geochemical evolution of the mantle will need much more thorough exploration than is possible here. This will involve exploration of model behaviors, refinement of model parameters, and re-examination of geological evidence for consistency with the types of behavior suggested here.

More detailed work is desirable on many aspects of the modelling, particularly on the mechanisms by which overturns begin, the possibility that penetration does not always trigger an overturn, the role of compositional effects on phase transformations, both in the transition zone and nearer the surface during subduction, and the role of plumes in transporting heat and potentially in triggering overturns. Alternative mechanisms, such as the role of the supercontinent cycle in modulating mantle dynamics [8,59], also need more investigation.

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References

[1] G. Gastil, The distribution of mineral dates in space and time, Am. J. Sci. 258, 1–35, 1960.
[2] M.T. McCulloch and V.C. Bennett, Progressive growth of the Earth's continental crust and depleted mantle: geochemical constraints, Geochim. Cosmochim. Acta 58, 4717–4738, 1994.
[3] M.A. Etheridge, R.W.R. Rutland and L.A.I. Wyborn, Orogenesis and tectonic process in the early to middle Proterozoic of northern Australia, in: Proterozoic Lithospheric Evolution, A. Kröner, eds., Am. Geophys. Union, Washington, D.C., 131–147, 1987.
[4] L.A.I. Wyborn, Petrology, geochemisty and origin of a major Australian 1880–1840 Ma felsic volcano-plutonic suite: a model for intracontinental felsic magma generation, Precambrian Res. 40/41, 37–60, 1988.
[5] W. Abouchami, M. Boher, A. Michard and F. Albarede, A major 2.1 Ga old event of mafic magmatism in West Africa: an early stage of crustal accretion, J. Geophys. Res. 95, 1990.
[6] P.F. Hoffman, United plates of America, the birth of a craton, Ann. Rev. Earth Planet. Sci. 16, 543–603, 1988.

[7] R.W. Page, Geochronology of early to middle Proterozoic fold belts in northern Australia: a review, Precambrian Res. 40/41, 1–19, 1988.
[8] P.F. Hoffman, Speculations on Laurentia's first gigayear (2.0–1.0 Ga), Geology 17, 135–138, 1989.
[9] P.F. Hoffman, Did the breakout of Laurentia turn Gondwanaland inside-out?, Science 252, 1409–1412, 1991.
[10] R. Unrug, The supercontinent cycle and Gondwanaland assembly: component cratons and the timing of suturing events, J. Geodynamics 16, 215–240, 1992.
[11] G.F. Davies, Conjectures on the thermal and tectonic evolution of the earth, Lithos 30, 281–289, 1993.
[12] G.F. Davies, On the emergence of plate tectonics, Geology 20, 963–966, 1992.
[13] R.I. Hill, I.H. Campbell, G.F. Davies and R.W. Griffiths, Mantle plumes and continental tectonics, Science 256, 186–193, 1992.
[14] M. Gurnis and G.F. Davies, Apparent episodic crustal growth arising from a smoothly evolving mantle, Geology 14, 396–399, 1986.
[15] P. Machetel and P. Weber, Intermittent layered convection in a model mantle with an endothermic phase change at 670 km, Nature 350, 55–57, 1991.
[16] P.J. Tackley, D.J. Stevenson, G.A. Glatzmaier and G. Schubert, Effects of an endothermic phase transition at 670 km depth in a spherical model of convection in the earth's mantle, Nature 361, 699–704, 1993.
[17] L.P. Solheim and W.R. Peltier, Avalanche effects in phase transition modulated thermal convection: a model of earth's mantle, J. Geophys. Res. 99, 6997–7018, 1994.
[18] S. Zhong and M. Gurnis, Role of plates and temperature-dependent viscosity in phase change dynamics, J. Geophys. Res. in press, 1994.
[19] G.F. Davies, Penetration of plates and plumes through the mantle transition zone, Earth Planet. Sci. Lett. in press, 1995.
[20] T. Nakakuki, H. Sato and H. Fujimoto, Interaction of the upwelling plume with the phase and chemical boundary at the 670 km discontinuity: effects of temperature-dependent viscosity, Earth Planet. Sci. Lett. 121, 369–384, 1994.
[21] S.P. Grand, Mantle shear structure beneath the Americas and surrounding oceans, J. Geophys. Res. 99, 11,591–11,621, 1994.
[22] R. van der Hilst, R. Engdahl, W. Spakman and G. Nolet, Tomographic imaging of subducted lithosphere below northwest Pacific island arcs, Nature 353, 37–43, 1991.
[23] R. van der Hilst, Complex morphology of subducted lithosphere in the mantle beneath the Tonga trench, Nature 374, 154–157, 1995.
[24] G.F. Davies, Ocean bathymetry and mantle convection, 1. Large-scale flow and hotspots, J. Geophys. Res. 93, 10,467–10,480, 1988.
[25] G.F. Davies and M.A. Richards, Mantle convection, J. Geol. 100, 151–206, 1992.
[26] G.F. Davies and F. Pribac, Mesozoic seafloor subsidence and the Darwin Rise, past and present, in: The Mesozoic Pacific, M. Pringle, W. Sager, W. Sliter and S. Stein, eds., Am. Geophys. Union, Washington, D.C., 39–52, 1993.
[27] N.H. Sleep, Hotspots and mantle plumes: Some phenomenology, J. Geophys. Res. 95, 6715–6736, 1990.
[28] G.F. Davies, Cooling the core and mantle by plume and plate flows, Geophys. J. Int. 1993.
[29] A. Davaille and C. Jaupart, Transient high-Rayleigh-number thermal convection with large viscosity variations, J. Fluid Mech. 253, 141–166, 1993.
[30] A.E. Ringwood and T. Irifune, Nature of the 650-km discontinuity: implications for mantle dynamics and differentiation, Nature 331, 131–136, 1988.
[31] S.E. Kesson, J.D.F. Gerald and J.M.G. Shelley, Mineral chemistry and density of subducted basaltic crust at lower mantle pressures, Nature 372, 767–769, 1994.
[32] R.W. Griffiths and J.S. Turner, Viscous entrainment by sinking plumes, Earth Planet. Sci. Lett. 90, 467–477, 1988.
[33] L.P. Solheim and W.R. Peltier, Correction to "Avalanche effects in phase transition modulated thermal convection: a model of earth's mantle", J. Geophys. Res. 99, 18,203, 1994.
[34] M.A. Richards, R.A. Duncan and V. Courtillot, Flood basalts and hot-spot tracks: plume heads and tails, Science 246, 103–107, 1989.
[35] M.F. Coffin and O. Eldholm, Large igneous provinces: crustal structure, dimensions and external consequences, Rev. Geophys. 32, 1–36, 1994.
[36] I.H. Campbell and R.W. Griffiths, The changing nature of mantle hotspots through time: Implications for the chemical evolution of the mantle, J. Geol. 92, 497–523, 1992.
[37] I.H. Campbell and R.I. Hill, A two-stage model for the formation of the granite–greenstone terrains of the Kalgoorlie–Norseman area, Western Australia, Earth Planet. Sci. Lett. 90, 11–25, 1988.
[38] G.F. Davies, Role of the lithosphere in mantle convection, J. Geophys. Res. 93, 10,451–10,466, 1988.
[39] G.F. Davies, Generation and rise of mantle plumes, in prep., 1995.
[40] R.W. Griffiths and I.H. Campbell, Stirring and structure in mantle plumes, Earth Planet. Sci. Lett. 99, 66–78, 1990.
[41] K.C. Condie, Episodic ages of greenstones: a key to mantle dynamics?, Geophys. Res. Lett. in press, 1995.
[42] S.J. Daly and C.M. Fanning, Archaean, in: The Geology of South Australia, Vol. 1, The Precambrian, J.F. Drexel, W.V. Preiss and A.J. Parker, eds., Geol. Surv. South Australia, 33–50, 1993.
[43] M. Boher, W. Abouchami, A. Michard, F. Albarede and N.T. Arndt, Crustal growth in West Africa at 2.1 Ga, J. Geophys. Res. 97, 345–369, 1992.
[44] K.C. Condie, Growth and accretion of continental crust: inferences based on Laurentia, Chem. Geol. 83, 183–194, 1990.
[45] E.M. Moores, Southwest U.S.–East Antarctica (SWEAT) connection: a hypothesis, Geology 19, 425–428, 1991.
[46] P.F. Hoffman and S.A. Bowring, Short-lived 1.9 Ga continental margin and its destruction, Wopmay orogen, northwest Canada, Geology 12, 68–72, 1984.
[47] R. Page and S.-s. Sun, Evolution of the Kimberly region, W.A., and adjacent Proterozoic inliers—new geochronologi-

cal constraints (abstract), 12th Australian Geological Convention, Abstracts No. 37, 332–333, 1994.
[48] P.F. Hoffman, Wopmay orogen: a Wilson cycle of early Proterozoic age in the northwest of the Canadian shield, in: The Continental Crust and its Mineral Deposits, D.W. Strangway, eds., Geol. Assoc. Can. Spec. Pap., St. Johns, 523–549, 1980.
[49] S. El-Gaby and R.O. Greiling, The Pan-African event: history and evolution, in: The Pan-African Belt of Northeast Africa and Adjacent Areas, S. El-Gaby and R.O. Greiling, eds., Friedr. Vieweg, Braunschweig, 5–13, 1988.
[50] M. Stein and A.W. Hofmann, Mantle plumes and episodic crustal growth, Nature 372, 63–68, 1994.
[51] R.L. Larson, Latest pulse of the earth: evidence for a mid-Cretaceous superplume, Geology 19, 547–550, 1991.
[52] S.J.G. Galer, S.L. Goldstein and R.K. O'Nions, Limits on chemical and convective isolation in the earth's interior, Chem. Geol. 75, 257–290, 1989.
[53] G.F. Davies, Mantle plumes, mantle stirring and hotspot chemistry, Earth Planet Sci. Lett. 99, 94–109, 1990.
[54] U.R. Christensen and A.W. Hofmann, Segregation of subducted oceanic crust in the convecting mantle, J. Geophys. Res. 99, 19,867–19,884, 1994.
[55] I.H. Campbell, G.K. Czemansky, V.A. Fedorenko, R.I. Hill and V. Stepanov, Synchronism of the Siberian Traps and the Permian–Triassic boundary, Science 258, 1760–1763, 1992.
[56] M.R. Rampino and R.B. Stothers, Flood basalt volcanism during the past 250 million years, Science 241, 663–668, 1988.
[57] A.H. Knoll, The early evolution of Eukaryotes: a geological perspective, Science 256, 622–627, 1992.
[58] D. Canil, H.S.C. O'Neill, D.G. Pearson, R.L. Rudnick, W.F. McDonough and D.A. Carswell, Ferric iron in peridotites and mantle oxidation states, Earth Planet. Sci. Lett. 123, 205–220, 1994.
[59] M. Gurnis, Large-scale mantle convection and the aggregation and dispersal of supercontinents, Nature 332, 695–699, 1988.