9.08 Thermal Evolution of the Mantle

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9.08.1 Introduction

The thermal evolution of Earth's interior has featured fundamentally in the geological sciences, and it continues to do so. This is because Earth's internal heat fuels the driving mechanism of tectonics, and hence of all geological processes except those driven at the surface by the heat of the Sun. The thermal evolution also featured famously in a nineteenth-century debate about the age of Earth.

9.08.1.1 Cooling and the Age of Earth

According to Hallam (1989), from the time when the origin of Earth was considered by scientists of the European Enlightenment it was conceived as having started hot, notably in the nebular theories of Kant and Laplace. This led to conjectures and very rough estimates of how long it may have taken to cool down, and thus to estimates of the age of Earth. However, it was not until Fourier had formulated his 'law' of conduction and the science of thermodynamics was established that a quantitatively reliable, though physically inappropriate, estimate of Earth's age emerged. This was done by William Thomson, better known as Lord Kelvin, who played a key role in the development of thermodynamics, notably as the formulator of the 'second law'.

Kelvin actually calculated an age for the Sun first, using estimates of the gravitational energy released by the Sun's accretion from a nebular cloud, the heat content of the present Sun, and the rate of radiation of heat, concluding that the Sun was likely to be younger than 100 Ma and most unlikely to be older than 500 Ma (Kelvin, 1862). The following year Kelvin calculated an age of Earth using different physics (Kelvin, 1863). He reasoned that the known geothermal gradient of about 20°C km^{-1} near Earth's surface would have declined from an essentially infinite gradient at the time of Earth's formation. Making use of Fourier's law to solve the resulting thermal diffusion problem (Davies, 1999) he obtained an age between 20 and 400 Ma with his most probable value being 98 Ma. There ensued several decades of heated debate between and among geologists and physicists, with Kelvin's last estimate (Kelvin, 1899) being only 24 Ma.

A number of Kelvin's assumptions were challenged during the course of this long controversy, the most telling challenge in retrospect coming from The Reverend Osmond Fisher (Fisher, 1881) who pointed out that if Earth's interior were 'plastic' (meaning in some degree fluid), then a much greater reservoir of heat would be tapped and the resulting age could be much greater. Fisher's point was that conduction alone would have cooled only the outer few hundred kilometers of Earth, whereas a circulating fluid could bring heat from much deeper as well, so it would take longer for the larger amount of heat to be removed at the currently observed rate. Fisher's point did little to settle the nineteenth-century controversy, but it returned in a slightly different guise a century later.

The controversy with Kelvin was finally settled through the discovery of radioactivity, which provided a previously unknown source of heat within Earth, so heat could be replenished as it is lost through the surface, and the flow of heat thus maintained for much longer than Kelvin's estimates. The actual determination of Earth's age relied on exploiting a different aspect of radioactivity, the production of daughter isotopes. This task was pursued most notably by Holmes from early in the twentieth century (Holmes, 1911, 1913) but not completed to general satisfaction until meteorites were dated by the lead–lead method and Earth was shown to fit within the meteorite trend (Patterson, 1956). The age of Earth is still determined only indirectly.

9.08.1.2 A Static Conductive Mantle?

The discovery of the heat generated by radioactivity provided Earth with an enduring source of heat, but it created another problem by breaking the consistency in Kelvin's argument between age and the removal of heat by conduction. If heat escapes from Earth's interior by conduction, then Kelvin's argument would still seem to require Earth to be less than 100 My old, otherwise the surface heat flow would be much less than is observed. Put another way, if Earth is 4.5 Gy old, roughly 100 times Kelvin's estimate, then the geothermal gradient ought to be 10 times less than the 15–20 °C km^{-1} commonly observed in deep mines and boreholes. This is because in the relevant thermal diffusion process the heat flow (and the temperature gradient) declines in proportion to the square root of time (Davies, 1999).

Until the 1960s, it was most generally held that Earth's mantle is solid and unyielding (Jeffreys, 1976). This conclusion was based on several lines of evidence, the most obvious being that the mantle transmits seismic shear waves, and shear waves cannot propagate through a liquid because a liquid does not have any shear strength. A second argument is that Earth's response to tidal forces implies quite high rigidity. Finally, Earth has, on average, a slightly greater equatorial bulge than is accounted for by its rotation, and this was intepreted to imply that the mantle has a viscosity (a fluid's resistance to flow) sufficiently high to preclude significant internal motion. Against this, the concept of a deformable 'asthenosphere' extending down for several hundred kilometers was widely held among geologists as necessary to accommodate the considerable movements they inferred to have occurred in the crust (Barrell, 1914).

The initial presumption used to reconcile the observed heat flow with a static mantle was that the radioactivity must be confined to within about 20 km of the surface, and observations in continental areas support this (Jeffreys, 1976). The implication is that most of the heat emerging from continental crust is generated there and can readily be conducted to the surface. The amount of heat coming from deeper was presumed or implied to be very small. However, this encounters the problem that heat flow from ocean basins is greater than from continents, but the thin, less radioactive oceanic crust can account for only a small fraction of the total (Sclater *et al.*, 1980).

A later presumption was that thermal conductivity must increase substantially with depth, so that a sufficient heat flow could be sustained for a much greater time. This possibility was bolstered by the proposal that radiative transfer of heat could be important in the mantle (Clark, 1957). The idea is that if mantle minerals are sufficiently transparent, then the blackbody radiation of the hot materials could transmit significant distances. Since the transmission would be like a random walk it constitutes a diffusion process and so can be described by an enhanced effective conductivity. However, the presence of iron in minerals tends to reduce the transparency of the minerals at the relevant wavelengths (Shankland *et al.*, 1979). Thus, it was not at all obvious that radiative transfer could resolve the issue. There have been recent claims for a significant effect (Hofmeister, 2005), though if they are significant it would probably only be for the deepest mantle.

9.08.2 The Convecting Mantle

The emergence of plate tectonics in the 1960s directly implied that at least the upper part of the mantle is mobile. In 1969 Goldreich and Toomre

(1969) pointed out that the excess bulge at the equator varied about as much with longitude as with latitude, so the old explanation that it was a fossil from past faster rotation could not be correct. In their reinterpretation, the Earth adjusts its rotation to bring its largest bulges to the equator, and this implies an 'upper' bound on viscosity that is quite compatible with a mobile mantle. With later interpretations of postglacial rebound constraints and positive geoids over subduction zones, the current picture is that the upper mantle has a viscosity of $(1\text{–}3) \times 10^{20}$ Pa s while the lower mantle is perhaps 30 times more viscous than this (Hager, 1984; Mitrovica, 1996; Mitrovica and Forte, 1997).

A robust argument in favor of mantle convection was put by Tozer, starting in the 1960s at a time when the possibility was still hotly disputed (Tozer, 1965, 1972). He noted that the viscosity of mantle material is strongly dependent on temperature, as we will be considering shortly, which means its resistance to convection decreases rapidly as its temperature increases. He argued that either the mantle started hot and would therefore be soft enough to convect, or radioactivity would heat it until it reached a temperature at which it would become soft and mobile, and convect. This is essentially our present understanding, as we will shortly see.

The idea of some kind of mantle convection quickly became widely accepted, though for a time there was some reluctance to admit the lower mantle into the convection mileu (Isacks *et al.*, 1968; McKenzie *et al.*, 1974). Eventually, the lower mantle was also conceived as mobile (McKenzie and Weiss, 1975; Davies, 1977; O'Connell, 1977) and it was then realized that convection would be efficient enough at transporting heat that the surface heat flux would fairly closely approximate the radioactive heat generation in the interior. However, it was difficult to quantify the heat transported by convection because the required numerical models challenged the computers available at that time and because the temperature dependence of mantle viscosity creates technical challenges. Nevertheless, there is a simple and fairly general relationship that relates the heat transported by a convecting fluid to the temperature difference across the fluid layer (e.g., Turcotte and Oxburgh, 1967; Rossby, 1969; McKenzie *et al.*, 1974). McKenzie and Weiss (1975) and Davies (1979) used this relationship to estimate temperature changes in the mantle on the assumption that the heat loss at Earth's surface reflects the heat generation in the interior.

Subsequently several groups realized, more-or-less independently, that the relationship could be used to calculate the thermal evolution of the Earth without assuming that heat loss is equal to heat production. Sharpe and Peltier (1978, 1979) were the first, although they implemented it in terms of an enhanced thermal diffusion, which does not accurately represent the internal temperature profile. Others soon used it to characterize the total heat transport in terms of a representative internal temperature (Cassen *et al.*, 1979; Schubert, 1979; Schubert *et al.*, 1979a, 1979b, 1980; Stevenson and Turner, 1979; Davies, 1980; Stacey, 1980). The treatments in these papers were closely equivalent, although differing in minor detail. Systematic presentations are given by Davies (1999), Schubert *et al.* (2001), and Turcotte and Schubert (2001). The presumption in the simple theory is that the essential relationship between temperature difference and heat flow is not sensitive to the local details, like geometry, whether the flow is turbulent or laminar, and so on. Experience in other fields had already shown that it can be a good first approximation (Rossby, 1969). The influence of some factors will be illustrated after a reference case is presented.

The relationship between heat flow and internal temperature is expressed most generally in terms of the Nusselt number, *Nu*, and the Rayleigh number, *Ra*:

$$Nu = a(Ra/Ra_c)^p \tag{1}$$

where a and p are constants and Ra_c is the critical Rayleigh number, that is, the Rayleigh number at which convection just begins. Typically $a \sim 1$, $p \sim 1/3$ and $Ra_c \sim 1000$. Here

$$Nu = qD/K\Delta T \tag{2}$$

and

$$Ra = \frac{g\rho\alpha\Delta T D^3}{\kappa\mu} \tag{3}$$

where q is the surface heat flux, D is the depth of the fluid layer, K is thermal conductivity, ΔT is the temperature difference across the fluid layer, g is gravity, ρ is density, α is thermal expansion, $\kappa = K/\rho C_P$ is thermal diffusivity, μ is viscosity, and C_P is specific heat at constant pressure. Equation [1] can be rearranged to give

$$q = q_r\left(\frac{\Delta T}{\Delta T_r}\right)^{1+p}\left(\frac{\mu}{\mu_r}\right)^{-p} \tag{4}$$

where subscript r refers to a reference state, which could be the present Earth, for example. This gives the heat flux in terms of the temperature difference driving the convection.

The viscosity is kept explicit in eqn [4] because it is a strong function of temperature in the mantle, and this has an important effect. The viscosity can be written as

$$\mu = \mu_{\mathrm{r}} \exp\left[T_{\mathrm{A}}\left(\frac{1}{T} - \frac{1}{T_{\mathrm{r}}}\right)\right] \quad [5]$$

where

$$T_{\mathrm{A}} = (E^* + PV^*)/R_{\mathrm{G}} = H^*/R_{\mathrm{G}} \quad [6]$$

and E^*, V^*, and H^* are the activation energy, volume, and enthalpy, respectively, P is pressure, R_{G} is the gas constant and μ_{r} is the viscosity at a reference temperature T_{r}. T_{A} could then be called an activation temperature. With $E^* = 400\ \mathrm{kJ\ mol^{-1}}$ and $R_{\mathrm{G}} = 8.31\ \mathrm{J\ mol^{-1}\ K^{-1}}$, $T_{\mathrm{A}} = 48\,100\ \mathrm{K}$, which is the source of the strong dependence of μ on T.

To compute the thermal evolution of the mantle we need the energy equation for the mantle, considered as a covecting layer. With the situation depicted in Figure 1, the energy equation can be written as

$$\frac{\mathrm{d}T_{\mathrm{u}}}{\mathrm{d}t} = \frac{M_{\mathrm{m}}H_{\mathrm{m}} + Q_{\mathrm{c}} - Q_{\mathrm{m}}}{\chi_{\mathrm{m}} M_{\mathrm{m}} C_{\mathrm{m}}} \quad [7]$$

where T_{u} is the upper-mantle temperature, H_{m} is the rate of heat generation per unit mass of the mantle,

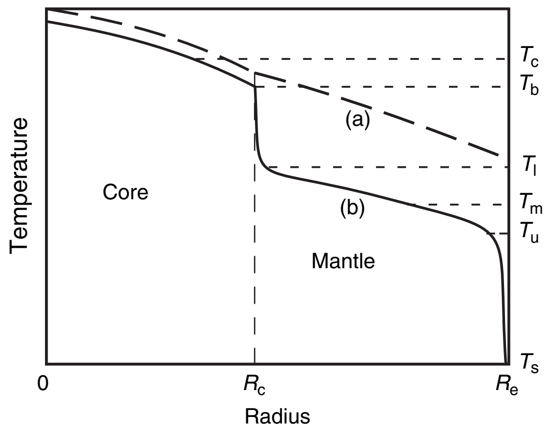

Figure 1 Sketch of temperature profiles through Earth (geotherms). (a) Initial state. (b) Present state. T_{c}, mean core temperature; T_{b}, core–mantle boundary temperature; T_{l}, lower-mantle temperature; T_{m}, mean mantle temperature; T_{u}, upper-mantle temperature; T_{s}, surface temperature.

M_{m} and C_{m} are the mass and specific heat of the mantle, respectively, and $\chi_{\mathrm{m}} = T_{\mathrm{m}}/T_{\mathrm{u}}$, where T_{m} is the mean temperature of the mantle. Q_{c} and Q_{m} are the heat flows out of the core and the mantle, respectively. For the moment we will take Q_{c} to be zero. Q_{m} is just given as

$$Q_{\mathrm{m}} = 4\pi R_{\mathrm{e}}^2 q \quad [8]$$

where R_{e} is the radius of Earth and q is given by eqn [4]. In this context it is appropriate to take $\Delta T = (T_{\mathrm{u}} - T_{\mathrm{s}})$, where T_{s} is the temperature at Earth's surface (Figure 1).

The heat generation is taken to be due to radioactive heating and is given by

$$H_{\mathrm{m}} = \frac{U_{\mathrm{e}}}{Ur}\sum_{i=1}^{4} h_i \exp[\lambda_i(t_{\mathrm{E}} - t)] \quad [9]$$

where U_{e} is the equivalent uranium concentration that would yield the observed heat loss, the index i refers to the isotopes $^{238}\mathrm{U}$, $^{235}\mathrm{U}$, $^{232}\mathrm{Th}$, and $^{40}\mathrm{K}$, h_i is the heat production per unit mass of uranium, λ_i is the decay constant, and t_{E} is the age of the Earth. Ur is the Urey ratio, defined as the ratio of present heat loss to present heat generation in the Earth, that is,

$$Ur = Q_{\mathrm{m}}/H_{\mathrm{m}}M_{\mathrm{m}} \quad [10]$$

This is the reciprocal of the way Ur is sometimes defined. Values for the required parameters are given in Table 1.

9.08.2.1 A Reference Thermal History

A solution for eqn [7] that reasonably satisfies observational constraints is shown in Figure 2 (heavy curves). Parameters used generally in calculations here are given in Table 2, while observational constraints are given in Table 3. The level of radioactive heating has been adjusted, through the Urey ratio, to yield a present heat loss of about 36 TW (36×10^{12} W), which is the portion of Earth's total heat loss (41 TW) emerging from the mantle, the balance being generated in the continental crust and lost directly to the surface (Davies, 1999).

The initial temperature is taken rather arbitrarily to be 1800°C. The peak temperature during the formation of Earth is difficult to estimate, because it was presumably determined by a competition between heat deposition by large impacts and heat removal by the 'gardening' effect of further impacts, by conduction near the surface, by mantle convection, and

Table 1 Parameters of heat-producing isotopes

Isotope (i)	Half life (Ga)	Decay const.[a] (λ_i) Ga^{-1}	Power[a] (μW) (kg Element)$^{-1}$	Element/U[b] (g/g)	Power (hi) (μW (kg U)$^{-1}$)
^{238}U	4.468	0.155	94.35	1	94.35
^{235}U	0.7038	0.985	4.05	1	4.05
^{232}Th	14.01	0.049	26.6	3.8	101.1
^{40}K	1.250	0.554	0.0035	1.3 × 104	45.5
				Total	245

[a]Stacey (1992).
[b]Galer *et al.* (1989).

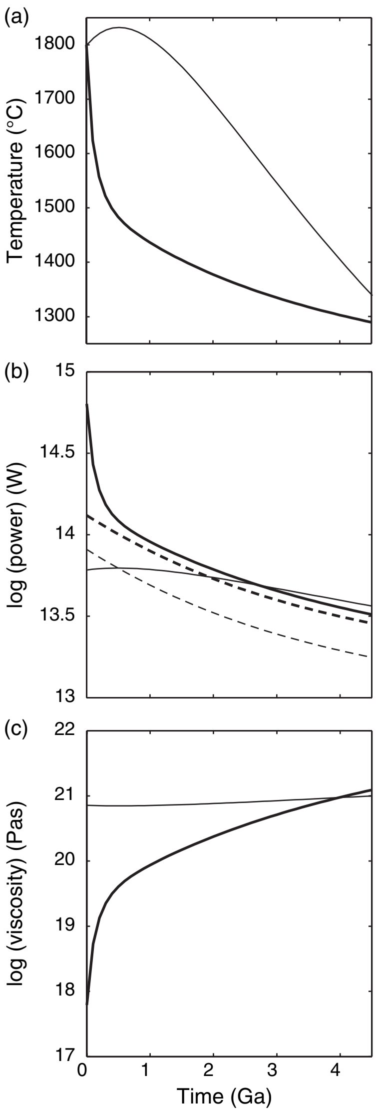

Figure 2 A typical thermal evolution of the mantle (heavy curves) and a thermal evolution in which the viscosity is almost independent of temperature (light curves). (a) Temperature. (b) Heat generation (dashed) and heat loss (solid). (c) Viscosity. The parameters of the calculations are adjusted to fit presently observed conditions given in **Table 3**.

by rapid cooling of surface melt, including the possibility of a transient magma ocean (Davies, 1990). Heat transport by magma is generally so much more efficient than the other mechanisms that it would presumably have prevented the temperature from rising too high, but all of the processes are complex and poorly constrained enough to make only rough estimates. Anyway the point here is just to illustrate the kind of behavior during later thermal evolution and the reasons for it.

Figure 2 illustrates typical features of this kind of solution – there is an early transient phase of rapid cooling, lasting about 0.5 Ga, and thereafter the heat loss tracks the heat generation, which slowly declines due to radiaoactive decay. As the heat generation declines, the mantle adjusts by slowly cooling so that its heat loss also declines. However, the slow cooling releases some internal heat that must also be removed, so the heat loss is larger than the heat generation. Another way to look at this is that the cooling occurs because the heat loss is a little larger than the heat generation, as described by eqn [7]. If the heat generation were constant, the heat loss would asymptotically approach the heat generation, and the ultimate state would be a steady state in which heat loss balanced heat generation.

The character of this solution is determined by the strong temperature dependence of the mantle viscosity, which changes by three orders of magnitude during the evolution (**Figure 2(c)**). At the beginning the viscosity is much lower than at present, and this reduces the resistance to mantle convection (see eqn [4]) which can therefore remove heat very rapidly (**Figure 2(b)**). This high heat loss causes the temperature to drop rapidly (**Figure 2(a)**), but then the viscosity rises rapidly (**Figure 2(c)**) and reduces the heat loss (**Figure 2(b)**). This early transient stage continues until the heat loss approaches the heat generation, at which point the initially large

Table 2 Quantities used in calculations

Symbol	Quantity	Value
M_c	Mass of the core	2.10^{24} kg
M_m	Mass of the mantle	4.10^{24} kg
R_e	Radius of the Earth	6371 km
R_c	Radius of the core	3485 km
C_c	Specific heat of the core	$1000\,\mathrm{J\,kg^{-1}\,K^{-1}}$
C_m	Specific heat of the mantle	$1000\,\mathrm{J\,kg^{-1}\,K^{-1}}$
ρ_m	Density of the mantle	$3500\,\mathrm{kg\,m^{-3}}$
α_m	Thermal expansion of the mantle	$2\times10^{-5}\,\mathrm{K^{-1}}$
κ_m	Thermal diffusivity of the mantle	$0.86\times10^{-6}\,\mathrm{m^2\,s^{-1}}$
η_u	Viscosity of the mantle	10^{21} Pa s
E^*	Activation energy of the mantle	$400\,\mathrm{kJ\,mol^{-1}}$
V^*	Activation volume	$2.5\times10^{-6}\,\mathrm{m^3\,mol^{-1}}$
T_s	Surface temperature	0°C
T_u	Present upper-mantle temperature	1300°C
n_m	Heat flow – temperature exponent (mantle)	1/3
U_e	Uranium abundance equivalent to present mantle heat loss	$38\,\mathrm{ng\,g^{-1}}$

Table 3 Empirical constraints on the present thermal regime

Quantity	Value	Reference
Upper-mantle temperature	1300°C	McKenzie and Bickle (1988)
Mantle heat loss	36 TW	Davies (1999)
Plume heat transport	3.5–7 TW	Davies (1999), Bunge (2005)

imbalance between them becomes small and therefore the rate of decline of temperature slows (**Figure 2(a)**).

The behavior is rather different if the viscosity is only a weak function of temperature, as is also illustrated in **Figure 2** (light curves) for a case in which the activation energy is reduced from 400 to 20 kJ mol^{-1}. In this case the initial heat loss is only a little greater than the present heat loss (61 vs 37 TW), because only the temperature term in eqn [4] is significantly different, the small variation in viscosity hardly affecting the heat loss. In fact, the heat loss in this example starts out lower than the heat generation, which means the mantle actually heats up initially. Only after the heat generation decays to lower values does the mantle temperature start to decrease. Thereafter the temperature declines, but never as fast as in the early transient phase of the black curve in **Figure 2(a)**. In fact, the transient phase is still in progress in the constant viscosity case. This is also reflected in the large difference between the heat loss and the heat generation (**Figure 2(b)**) – the Urey ratio in this case is 2.0.

The difference between these two solutions arises because in the former case a relatively small temperature change, about 200°C, is sufficient to reduce the heat loss as the rate of heat generation declines from about 130 to 29 TW, a decline by a factor of 4.5. The change in heat loss comes mainly through the large increase in viscosity (**Figure 2(c)** and eqns [4] and [5]) accompanying this modest change of temperature. In the latter case, the change in heat loss must be accomplished solely through ΔT, which must decrease by $(4.5)^{3/4}$, or a factor of 3. Thus, the mantle would be required to cool by 1200°C (from 1800 to 600°C) to reduce the heat loss to its presently observed value. This requires the removal of much more internal heat, which is why heat production is only half of the heat loss rate in this case, the other half coming from internal heat.

Thus, the strong temperature dependence of mantle viscosity results in the early transient cooling phase being relatively brief, about 500 Ma. Thereafter, the thermal regime tracks the slow decay of radiogenic heat. Incidentally, it is not very useful in light of this to think of contributions from primordial heat versus radiogenic heat, since there is no clear way to separate them in the present Earth. It is more useful to separate the early transient phase (which 'can' usefully be thought of as removing 'excess' primordial heat) from the later phase regulated by radioactivity. It is also useful to separate current radiogenic heat production and the heat released by the decreasing internal heat of the mantle, as its temperature declines.

9.08.3 Internal Radioactivity and the Present Cooling Rate

The approach taken above of adjusting the Urey ratio amounts to inferring the internal heat production rate that would account for the present rate of heat loss, assuming the present theory is accurate. The early conclusion from models like those in **Figure 1** was that the Urey ratio is significantly greater than 1, meaning the heat loss rate is greater than the heat generation rate. For example, Schubert *et al.* (1980) found $Ur = 1.2–1.5$, while Davies found Ur could be as high as 2 (Davies, 1980), though subsequent work yielded values in the range 1.25–1.35 (Davies, 1993).

Another early conclusion was that these results are marginally consistent with Earth having a complement of uranium and thorium like that of chondritic meteorites, with a potassium complement such that $K/U = (1–2) \times 10^4$. The latter ratio is lower than for chondrites, but consistent with the ratio found in surface rocks. The depletion of potassium relative to chondrites fits a trend of stronger depletion for more volatile elements (McDonough and Sun, 1995; Palme and O'Neill, 2004).

Subsequently, it has emerged that there seems to be a significant discrepancy. The U content of the primitive mantle (precrust extraction) is estimated to be $22\,\mathrm{ng\,g^{-1}}$ (Palme and O'Neill, 2004). With a current heat production from U, Th, and K of 245 $\mu\mathrm{W\,kg^{-1}}$ of U (**Table 1**), this implies a heat production of 5.4 $\mathrm{pW\,kg^{-1}}$ of mantle and a total heat production of 22 TW. On the other hand, the observed mantle heat loss is about 36 TW, which implies $Ur = 1.6$, higher than the estimates from thermal evolution modelling. $Ur = 1.3$ would require the total heat production to be about 28 TW.

The discrepancy is actually worse than that, because about half of Earth's complement of heat-producing elements is estimated to be in the continental crust (Rudnick and Fountain, 1995), which implies that the average mantle U content is only about $10\,\mathrm{ng\,g^{-1}}$. The actual U content of the source of mid-ocean ridge basalts is even smaller, less than $5\,\mathrm{ng\,g^{-1}}$ (Jochum *et al.*, 1983), which implies that there may be some deeper region of the mantle that has higher concentrations. If we take only the inferred mean mantle U content of $10\,\mathrm{ng\,g^{-1}}$ then the implied heat production is only about 10 TW and $Ur = 3.6$.

Such a large imbalance would imply that the mantle is cooling rapidly. **Table 4** summarizes several examples of heat input, Urey ratio, and the resulting cooling rates. Chondritic mantle heating implies that the mantle is cooling by about $110\,°\mathrm{C\,Ga^{-1}}$, whereas if only half of the chondritic heat sources are in the mantle, the cooling rate would be over $200\,°\mathrm{C\,Ga^{-1}}$. Thermal evolution models that match the present heat flow yield cooling rates of 50–$70\,°\mathrm{C\,Ga^{-1}}$, although the example given in **Figure 2**, which falls a little short of the observed heat loss rate, has a cooling rate of only $30\,°\mathrm{C\,Ga^{-1}}$. The case in

Table 4 Implied cooling rates of the mantle

Case	Thermal	Chondr.	.5 Chond.	Fig. 2	Fig. 2
Q_m(TW)	36	36	36	32.4	36.9
H_m(TW)	28	22	10	28.6	17.9
Q_{net}(TW)	8	14	26	3.8	19.0
Ur	1.25	1.6	3.6	1.13	2.1
dT/dt ($°\mathrm{C\,Ga^{-1}}$)	−63	−110	−205	−30	−150

Figure 2 with little variation in viscosity has a present cooling rate of $150\,°\mathrm{C\,Ga^{-1}}$.

It is difficult to see how the larger cooling rates just quoted can be consistent with the geological record, since they imply that the late Archean mantle would have been 300–600 °C hotter than at present. As pointed out by Campbell and Griffiths (1992), although there are Archean komatiites with source potential temperatures of 1800–1900 °C, the komatiites comprise only a small fraction of Archean mafic rocks, the great majority of which are basalts with source potential temperatures no more than 100–200 °C above the present mantle temperature. The komatiites are interpreted to come from the core of plumes or analogous high-temperature mantle upwellings, and would therefore not be representative of mean mantle temperatures. The lower cooling rates of $30–60\,°\mathrm{C\,Ga^{-1}}$ in **Table 4** would be consistent with this interpretation of the dominant basaltic Archean rocks.

9.08.4 Variations on Standard Cooling Models

A number of variations on this kind of thermal history model have been explored. These have the same general character as the reference model in **Figure 2**, but the details differ.

9.08.4.1 Effect of Volatiles on Mantle Rheology

The viscosity of mantle materials is affected by water content, as well as by temperature and pressure (Hirth and Kohlstedt, 1996). Reducing the water content from typical values of a few hundred micrograms per gram to zero can increase the viscosity by about an order of magnitude. The effect of this on thermal evolution was explored by McGovern and

Schubert (1989) and Schubert *et al.* (2001). Typical results are illustrated in **Figure 3**, in which 'degassing' and 'regassing' models are compared with a model in which the viscosity is not affected by water (Schubert *et al.*, 2001). Their parametrization of the degassing and regassing processes was rather simplified, but it suffices to illustrate the effect here. The main effect of the removal by degassing of up to 1.5 ocean masses from the mantle is to increase the mantle temperature by about 100°C. Returning water to the mantle ('regassing') has the reverse effect.

This result can be simply understood as the mantle self-regulating to compensate for the stiffening of its material. If all the water were to be removed suddenly, the mantle viscosity would increase and convection would slow. This would reduce the rate of heat loss, so the mantle would begin to heat up (or to cool more slowly) through its internal radioactivity. As it heated the viscosity would drop, and when the viscosity reached the value it had before the removal of the water the convection would have returned to its previous vigor and the previous rate of heat loss would be restored. The authors confirm that the heat flow and the viscosity are similar in the 'dry' and 'wet' models, the only significant difference being in their temperature.

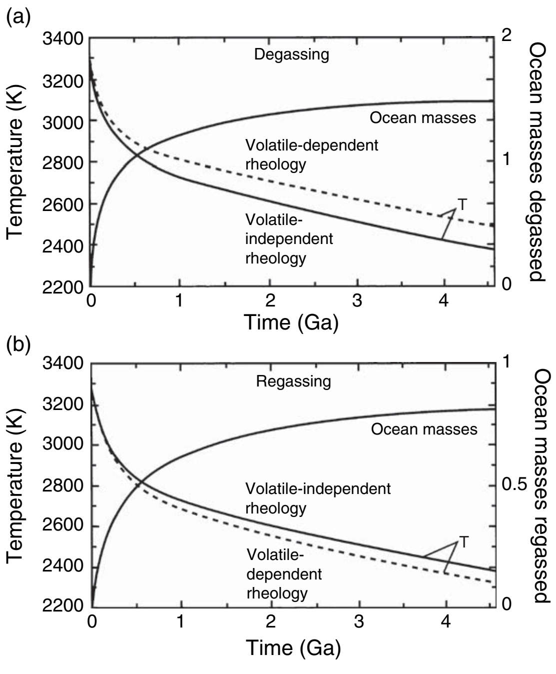

Figure 3 The effect of volatile-dependent viscosity on mantle cooling history (dashed curves). Volatile-independent case is included for comparison (solid curves). The main volatile component is assumed to be water, which reduces the viscosity of mantle materials. (a) Degassing: 1.5 ocean masses of water are progressively removed. Mantle temperature increases to compensate for the stiffening resulting from degassing. (b) Regassing: about 0.8 ocean masses of water are progressively added. Mantle temperature decreases as regassing reduces viscosity. From Schubert G, Turcotte DL, and Olson P (2001) *Mantle Convection in the Earth and Planets.* Cambridge: Cambridge University Press.

9.08.4.2 Two-Layer Mantle Convection

The hypothesis that the mantle convected in two layers separated at the transition zone at 660 km depth has fallen from favor since seismic tomography yielded images of subducted lithosphere penetrating deep into the lower mantle (Grand *et al.*, 1997). However, the possibility of layering deeper in the mantle is still debated (Kellogg *et al.*, 1999). Although the example given here is of the earlier form of two-layer convection, it adequately illustrates the effect of layering.

Spohn and Schubert (1982) and Schubert *et al.* (2001) formulated the viscosity in terms of the so-called homologous temperature, which is the ratio of temperature to melting temperature (or more accurately for the mantle, the solidus temperature, at which melting begins). They also assumed that the lower-mantle material is 60 times more viscous than the upper mantle at the same temperature and pressure, due to it being in high-pressure phases. The resulting temperature structure is shown in **Figure 4**(a), with a whole-mantle convection model for comparison. There is a steep increase in temperature by nearly 500°C near 660 km depth. This increase actually comprises two thermal boundary layers on opposite sides of the interface – a lower thermal boundary layer of the upper mantle and an upper thermal boundary layer of the lower mantle. There is also a small thermal boundary layer at the base of the lower mantle and a strong one at the top of the upper mantle, for a total of four thermal boundary layers.

In spite of the higher intrinsic viscosity of the lower mantle in this model, the viscosity ends up being almost the same as in the upper mantle because of the higher temperature in the lower mantle. As in the previous example, the mantle has self-regulated by adjusting its temperature until the viscosity is such as to allow the required heat transport to occur. (In this particular model, the rate of heat generation in the lower mantle had to be kept quite low so the

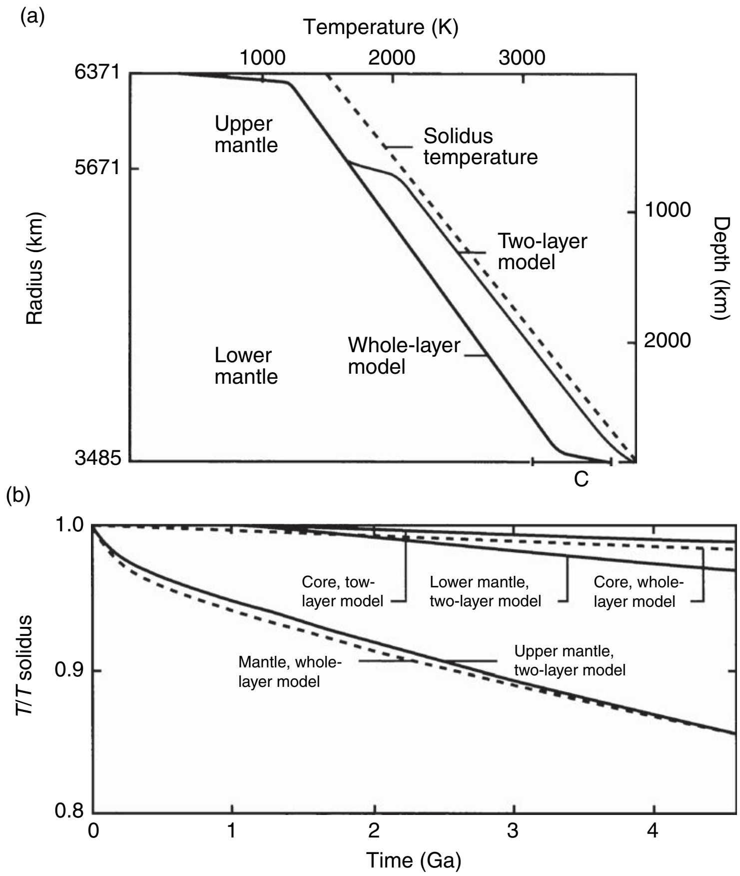

Figure 4 The effect of mantle convection operating in two separate layers. (a) Temperature profiles. (b) Temperature histories. A single-layer ('whole-layer') model is included for comparison. The lower-mantle layer is about 500°C hotter and has a similar viscosity to the upper mantle, in spite of being intrinsically more viscous. See text for details. From Schubert G, Turcotte DL, and Olson P (2001) *Mantle Convection in the Earth and Planets.* Cambridge: Cambridge University Press.

lower mantle did not melt. However, the mantle solidus is not very well determined at very high pressures so the real mantle might be more tolerant of higher heating rates.)

The evolution of homologous temperature is shown in Figure 4(b). The upper-mantle cooling is similar to the reference example given earlier – there is a brief, rapid transient cooling followed by a slow decline as radioactivity declines. The lower mantle cools much less. This is because it can only begin to cool as the upper-mantle temperature falls, and because the temperature drop across the top thermal boundary layer of the lower mantle is relatively small, so it drives weaker convection and so transports less heat. (The authors also included a separate core layer in both models. An example in which the core is treated explicitly is given below.)

9.08.4.3 Static Lithosphere Convection, and Venus

Earth's lithosphere turns out to be unusual within the solar system, in that it is mobile in spite of being cold and strong. Its mobility comes about because it is also, loosely speaking, brittle. Evidently internal stresses are enough to fracture it into pieces, and the pieces comprise the tectonic plates. Other solid planets and satellites in the solar system seem to have strong but unfractured lithospheres, with the possible exception of Venus. These are sometimes called one-plate planets. The big difference, from the point of view of mantle convection, is that most of the lithospere does not partake in the convection. It is expected that only the lower part of the lithosphere will be warm enough to be mobile and hence 'drip' away and drive convection. Since the lithosphere is also the location of most of the (negative) buoyancy that might drive convection, it means that less buoyancy is available to drive convection than is available in Earth's mantle, in which the whole lithosphere founders into the mantle.

The effect of a static lithosphere can be illustrated by using a higher surface temperature, T_s, in calculating the temperature difference ΔT across the top thermal boundary layer in eqn [4]. In effect T_s is taken to define the boundary between the static and mobile parts of the lithosphere. An example with $T_s = 900°C$ is shown in **Figure 5**. The main effect is to raise the temperature of the mantle by nearly 200°C. As in previous examples, we see that the mantle self-regulates by adjusting its internal viscosity, through its temperature, until the convection is vigorous enough to transport the required amount of heat. The viscosity is reduced by about a factor of 25 (**Figure 5(c)**). This type of model might also apply to Venus, whose surface temperature really is high, about 475°C.

The more difficult question in considering a static lithosphere is in estimating how much of the lower lithosphere will take part in convection, especially in different circumstances or as a mantle slowly cools. This has been considered in a series of studies, and various formulations have been proposed. One approach is to use the temperature dependence of viscosity to define a length scale over which the viscosity varies by a characteristic amount, say a factor of 10. Several studies have found scaling laws using a mobile layer thickness defined in terms of this length scale (Davaille and Jaupart, 1993; Solomatov, 1995; Ratcliffe *et al.*, 1997; Grasset and Parmentier, 1998) and using the viscosity of the actively convecting fluid below the lithosphere. On the other hand, Manga *et al.* (2001) used the viscosity at the mean temperature across the lithosphere and concluded this more usefully characterized their results.

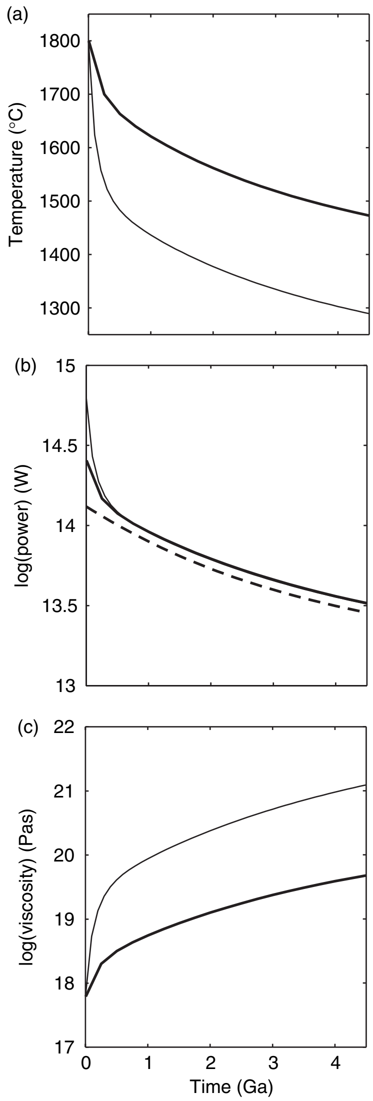

Figure 5 Effect of a static lithosphere, simulated through a surface temperature of 900°C (heavy curves). A mobile lithosphere case is included for comparison (light curves). (a) Temperature. (b) Rate of heat loss. Dashed curve, heat generation. (c) Viscosity. The high surface temperature effectively defines the boundary between the lower, warmer part of the lithosphere that is soft enough to partake in convection and the upper, cooler part that is too strong to move. Only the lower, mobile part of the lithosphere is available to drive convection, and the mantle compensates for the reduced driving buoyancy by running hotter, by nearly 200°C, so the viscous resistance is reduced in proportion. The heat loss (panel (b)) is the same as for the mobile-lithosphere model. This type of model could also apply to Venus, whose surface tempartyure is high (475°C).

9.08.5 Coupled Core–Mantle Evolution and the Geodynamo

Stevenson and Schubert (1983) and Stacey and Loper (1984) first treated the core as a separate layer with its own thermal history, and the latter authors made the fundamental point that plumes or other upwellings from the base of the mantle are driven by heat emerging from the core. Subsequently, others have further explored such models (Davies, 1993, 2007b; Schubert *et al.*, 2001; Nimmo, 2007). **Figure 6** shows an example, taken from Davies (2007b). The evolution of mantle temperature and surface heat flow is very similar to that of the reference model (**Figure 1**). The model is adjusted so the present heat flow from the core,

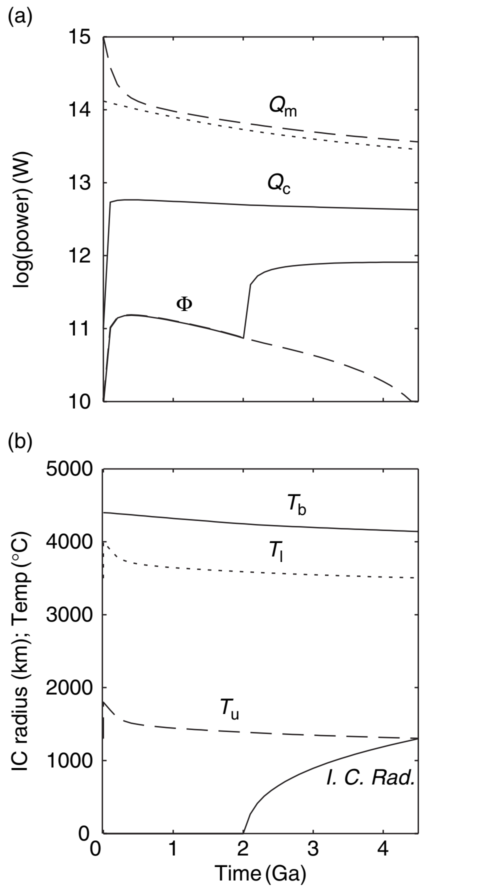

Figure 6 Coupled thermal history of the mantle and core, including dissipation in the core and growth of the inner core. (a) Heat flow. Q_m, mantle heat loss; Q_c, core heat loss; Φ, dissipation in the core with (solid) and without (dashed) the effects of the inner core; dotted curve, heat generation. (b) Temperature histories and inner core radius. Symbols are defined in **Figure 1**. From Davies GF (2007b) Mantle regulation of core cooling: A geodynamo without core radioactivity? *Physics of the Earth and Planetary Interiors* 160: 215–29.

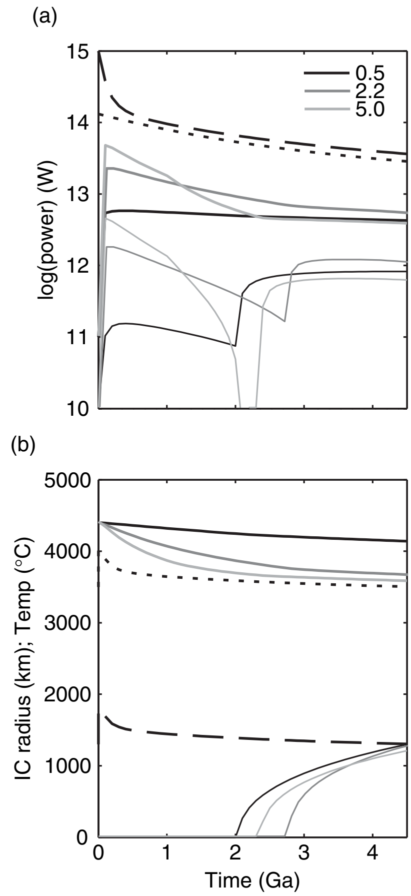

Figure 7 Core thermal histories with different efficiency factors governing removal of heat by the mantle. Curve identifications are the same as in **Figure 6**. From Davies GF (2007b) Mantle regulation of core cooling: A geodynamo without core radioactivity? *Physics of the Earth and Planetary Interiors* 160: 215–229.

4.3 TW, is within the range inferred from hot spot swells (**Table 3**), on the assumption that hot spots are caused by plumes rising from the base of the mantle (Davies, 1988; Sleep, 1990). In this model the core heat flow is remarkably steady through Earth history, with a maximum of only 5.8 TW. This low and steady heat flow causes only moderate core cooling, by 260°C. The temperature difference between the mantle and the core also remains fairly steady.

It turns out that there are other possible thermal histories that yield similar present heat flows. This is illustrated in **Figure 7**, which compares three cases in which the efficiency with which mantle plumes transport heat away from the core is varied. There is some uncertainty in a numerical factor controlling this efficiency, with which the cases are labeled (Davies, 2007b). With higher plume efficiency, the early core heat flow is higher (**Figure 7(a)**) and the core cools more (**Figure 7(b)**). The mantle cooling history is not greatly affected and only one case is shown, for clarity. Rapid core cooling reduces the temperature difference between the core and the mantle and, since this also determines the plume heat flow, the core heat loss declines (**Figure 7(a)**) and ends up close to the first case. The value of the present core heat loss actually passes through a maximum in this kind of model as the efficiency of plume transport is varied (final heat losses are, respectively, 4.3, 5.3, 3.8 TW). To remove this ambiguity of past core heat loss we will need either better characterization of the efficiency of plumes (or other upwelling), or perhaps contraints from paleomagnetism, or both.

The thermal history of the core is closely tied to the generation of the Earth's magnetic field, which is believed to occur by the dynamo action of convection in the core (e.g., Labrosse and Macouin, 2003; Roberts *et al.*, 2003). A puzzle has arisen recently concerning how to reconcile the calculated core cooling rate with the energy required to maintain the dynamo and also with the inferred age of the inner core. The inner core is involved because compositional convection arising from the solidification of the inner core couples to the dynamo more efficiently, which is therefore readily maintained. Before the inner core began to solidify fairly strong thermal convection might have been required to maintain the dynamo, but this would imply faster core cooling and, eventually, rapid growth of the inner core. Unless the inner core began to grow relatively recently (only about 1 Ga), it would have grown bigger than is observed. This only intensifies the difficulty of driving the dynamo before the inner core started to form, and yet the paleomagnetic evidence is that a magnetic field of comparable strength to the present field has existed at least since 2.5 Ga, and possibly since 3.5 Ga. To reconcile these difficulties, it has been proposed that the core contains a radioactive heat source, in the form of dissolved potassium (Buffett, 2002; Nimmo *et al.*, 2004). None of the models in **Figures 6** and **7** include core radioactivity.

Included in **Figures 6**(**a**) and **7**(**a**) is the rate of energy dissipation in the core, Φ, which is the maximum energy available to the dynamo. Also included (**Figures 6**(**b**) and **7**(**b**)) is the radius of the inner core, with the initiation temperature adjusted to yield the present inner core radius of 1220 km. There is some uncertainty in the energy required to maintain the dynamo, with estimates ranging from 0.1 to 2 TW (Buffett, 2002; Roberts *et al.*, 2003). Two of the models have dissipation around 1 TW for much of Earth history, so they would be viable even with a fairly high dynamo requirement. The low core heat model of **Figure 6** would only be viable before the inner core formed if a low-energy requirement applies. This argument is explored in more detail elsewhere (Nimmo *et al.*, 2004; Davies, 2007b; Nimmo, 2007). The main point here is to illustrate how the core thermal history is regulated by mantle dynamics and therefore quite sensitive to the details of mantle behavior.

9.08.6 Alternative Models of Thermal History

The models considered so far are characterized by assuming mantle convection to be basically similar to convection in more familiar fluids and by their smoothly decreasing temperature and heat flow. On the other hand, Earth's tectonic history appears to have been quite episodic, since the distribution of ages in the continental crust is strongly peaked, as is illustrated in **Figure 8**. Since plate tectonics is a manifestation of mantle convection, it is not obvious that such a smoothly varying history could give rise to such a peaked age distribution. Therefore, mechanisms that might produce an episodic thermal evolution have been considered. We have also seen that estimates of Earth's internal radioactivity based on conventional thermal evolution models are hard to reconcile with estimates from cosmochemistry. This has given rise to one quite novel kind of model, which is considered below.

9.08.6.1 Episodic Histories

It was suggested fairly soon after plate tectonics became accepted that pressure-induced phase transformations in the mantle transition zone (400–660 km depth) could affect convection by either hindering or enhancing the rising and sinking of buoyant and negatively buoyant fluid (Schubert *et al.*, 1975). Machatel and Weber (1991) demonstrated that, depending on the thermodynamic parameters involved, phase

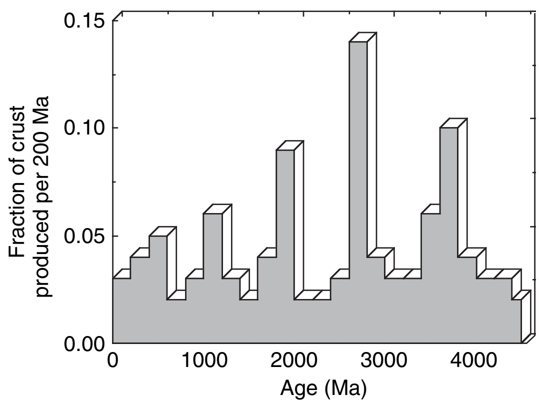

Figure 8 Distribution of the age of continental crust. From McCulloch MT and Bennett VC (1994) Progressive growth of the Earth's continental crust and depleted mantle: Geochemical constraints. *Geochimica et Cosmochimica Acta* 58: 4717–4738.

transformations could so hinder convection that it separates into two layers. This was confirmed in more detail in subsequent studies (e.g., Honda *et al.*, 1993; Tackley *et al.*, 1993; Nakakuki *et al.*, 1994; Solheim and Peltier, 1994b, 1994a; Weinstein, 1995). However, studies that incorporated strong temperature dependence of viscosity, thus allowing plates and plumes to be incorporate into the models, showed less propensity for layering (Nakakuki *et al.*, 1994; Davies, 1995a). Also it seems that the thermodynamic parameters of the phase transformations are only marginally adequate to cause layering, at least in the present mantle (Bina and Helffrich, 1994; Tackley *et al.*, 2005).

However, it is also plausible that conditions in the past were more conducive to layering. This led Honda (1995) to develop a parametrized convection model in which the mantle is initially layered but later undergoes a transition to whole-mantle convection. Davies (1995b) developed a more elaborate model, featuring multiple episodes of layering followed by breakdown into whole-mantle convection. One such history is illustrated in **Figure 9**. There is considerable uncertainty in several of the parameters entering such models, so this should be taken as no more than a conjectural illustration of the kind of episodes that might have been induced by phase transformations.

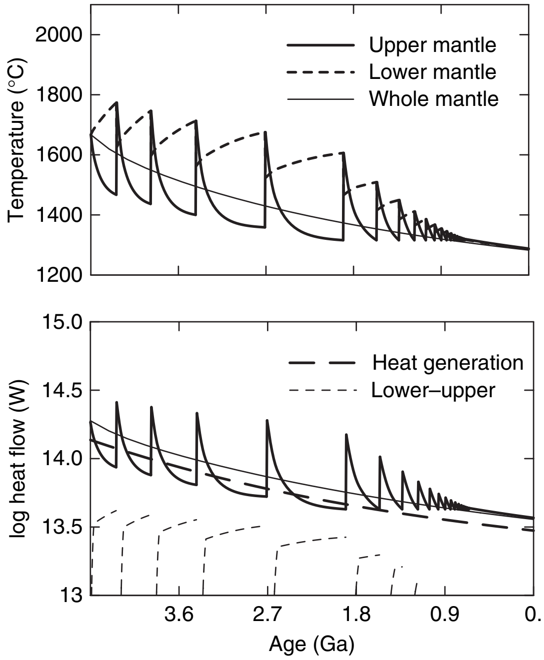

Figure 9 An episodic thermal history of the mantle, with episodes induced by parametrized effects of phase transformations. An unlayered ('whole mantle') case is included for comparison. In the lower panel the heat flow out of the mantle (heavy curve) and between the lower and upper layers (light dashed curves) are included. During the early evolution mantle convection stratifies into two layers, but eventually the temperature difference between the layers becomes large enough to cause a breakthrough and thermal homogenization, before layering is re-established. This example is conjectural and illustrative only (see text). From Davies GF (1995b) Punctuated tectonic evolution of the earth. *Earth and Planetary Science Letters* 136: 363–379.

9.08.6.2 Strong Compositional Lithosphere

Motivated in part by the discrepancy in inferred radioactive heat sources, Korenaga (2006) has devised a substantially different kind of thermal evolution model. He assumes that because of enhanced melting in a hotter mantle, the thickness of the lithosphere, defined as the strong outer region of Earth, would be controlled not by the thermal boundary layer but by the zone of strong depletion due to melting. He also assumes that the bending of plates prior to subduction is a major source of resistance to plate tectonics, and that the radius of curvature of the bending is proportional to plate thickness. The result of this combination of assumptions is that plates go slower in a hotter mantle, and that in turn means heat loss is inhibited at high mantle temperatures.

The result is illustrated in **Figure 10**. The model assumes a Urey ratio such that $\gamma = 1/Ur$ is between 0.15 and 0.3 (his definition of Urey ratio is the inverse of the one used here). The thermal history was integrated backward from the present, on the grounds that the present is better determined. The model has a number of novel features. The temperature would have been quite high in the Archean, perhaps even reaching a maximum, then declining at an increasing rate toward the present (in this respect the model is reminiscent of the viscosity-invariant model of **Figure 2**). The recent cooling rate is over $100°C\,Ga^{-1}$. On the other hand, the heat loss would have been little different in the Archean and would have peaked only about 0.5 Ga. Correspondingly, plate velocities would also have peaked recently and been lower in the Archean, while the inverse Urey ratio would have declined toward the present. A 'conventional' thermal history with such a low inverse Urey ratio would have had to be extremely hot in the late Archean and unrealistically hot in the early Archean (**Figure 10(a)**).

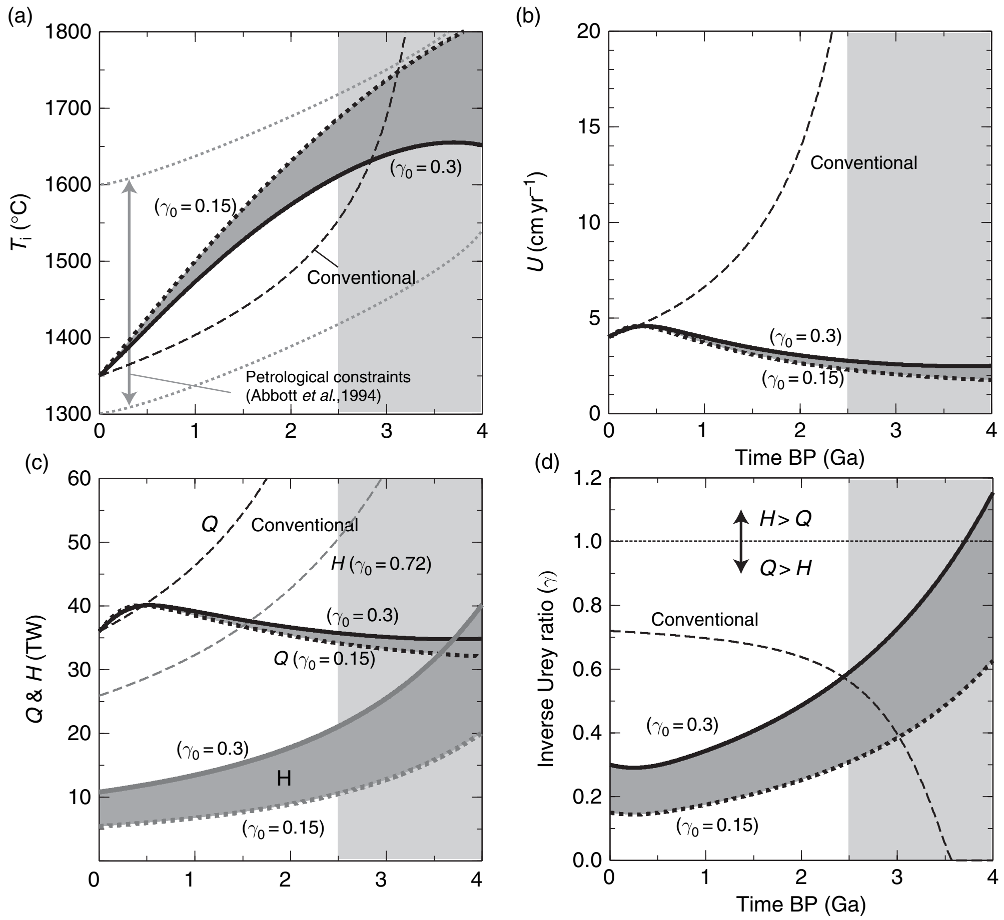

Figure 10 Mantle thermal evolution under the assumptions used by Korenaga (2006) and assuming low inverse Urey ratio $\gamma = 1/Ur = 0.15$ and 0.3 – see text. 'Conventional' cases are included for comparison (short-dashed). (a) Mantle temperature. Estimated petrological constraints from Abbott (Abbott *et al.*, 1994) are included. (b) Plate velocity. (c) Heat flow, *Q*, and heat generation, *H*. (d) Inverse Urey ratio. Gray shading on right indicates region of high uncertainty. From Korenaga J (2006) *The Archean Earth.* Washington, DC: American Geophysical Union.

Although there are aspects of this model that are not very well established, it does illustrate the possibility of quite different controls on thermal history than have usually been assumed, and it has some significant merits. For example, it reconciles the thermal history with a high Urey number, and it might help to explain why continental lithosphere surviving from the Archean is so thick. Nevertheless, its results are sensitive to some of its assumptions and parameter values. For example, assuming the bending radius of subducting lithosphere is proportional to thickness is plausible at first sight, but nonlinear rheological effects could cause the plate to bend more sharply with less dissipation, and this could considerably change the results. In addition, Korenaga argues that the high mantle temperatures in his model are compatible with the range of temperatures inferred by Abbott *et al.* (1994). On the other hand, as noted earlier, Campbell argues that komatiites yielding the higher inferred temperatures are unusual and that most of the Archean mantle was producing basalt, and therefore within 100°C or so of the present temperature (Campbell and Griffiths, 1992). This model needs to be further debated.

9.08.7 Implications for Tectonic Evolution

The two main expressions of mantle convection in the present Earth are plate tectonics and volcanic hot spots, many of which are plausibly due to mantle plumes (Davies, 2005). If mantle convection changes then its tectonic expression may also change. The main difference between the early mantle and the present mantle, according to the models presented here, is that the early mantle was hotter. The question is then whether the higher temperature would change the tectonic expression of mantle convection.

Plate tectonics can be traced back with reasonable confidence from geological evidence to about 2 Ga (Hoffman and Bowring, 1984), but whether plates operated earlier than that is debated (Kröner, 1981; Harrison *et al.*, 2005; Stern, 2005). The antiquity of plumes is also a matter requiring careful interpretation of the geological record (Campbell and Griffiths, 1992).

Korenaga's theory (Korenaga, 2006) implies that tectonics were different in the Archean, at least in terms of what regulates plate velocities. Plates could have operated in the Archean, though their dynamics would have been significantly different than at present.

The tectonic regime is determined by how the mantle gets rid of its heat (currently through plate tectonics) and how the mantle removes heat emerging from the core (currently plumes, although some debate that; Anderson (2004); Foulger (2005)). Any proposed tectonic mechanism must meet two fundamental conditions. First, there must be forces sufficient to drive it. Second, it must remove heat at a sufficient rate to cool the mantle to its present temperature (or, more generally, it must be explained how the mantle reached its present temperature). An understanding of mantle dynamics is thus an indispensable part of resolving the question of past tectonic regimes. The subject is far from being able to predict with any confidence which modes would apply. Nevertheless, it has progressed to the point of being able to examine some key questions.

9.08.7.1 Viability of Plate Tectonics

Davies (1992) proposed that plate tectonics might not have been viable when the mantle was hotter because greater melting under spreading centers would produce thicker oceanic crust, the buoyancy of which would hinder plate subduction. Even at present, plates initially have a net positive buoyancy because the oceanic crust is less dense than the mantle. They do not become negatively buoyant until conductive cooling has thickened them to the point where their negative thermal buoyancy overcomes the crust's positive compositional buoyancy. At present this occurs when they are about 15 My old. Since at present the average age of plates at subduction is about 100 Ma, this does not interfere significantly with the plates' ability to cool the mantle. However, as recently as 1.6 Ga the plates would, on average, have become negatively buoyant only as they arrived at a subduction zone. This is not only because the crust would have been thicker but also because plates would have been going faster.

Prior to this 'cross-over' time, plates would still be positively buoyant at the time they would need to subduct if the mantle was to be cooled. Plate tectonics might still have operated, but the plates would need to age for longer before they became negatively buoyant and thus able to subduct. This means plate tectonics would have been less efficient at cooling the mantle. The hotter the mantle, the less efficient plate tectonics would be. This leads to the paradox that the early hot mantle would not have been able to cool to its present temperature, in fact it might have gone into thermal runaway. The implication is that some other mode of tectonics would have been required. Another way to say this is that the top thermal boundary layer of the mantle, which drives the cooling mode of mantle convection, would have to have operated in a different dynamical mode.

Davies (1992) noted that one way out of this paradox might be provided by the fact that basaltic oceanic crust transforms to eclogite at about 60 km depth. Since eclogite is denser than the average mantle, it would then hasten plate subduction rather than hindering it. Once plate tectonics was operating it might be able to continue to operate because the deeper plate would pull the surface plate down even if its surface buoyancy were still slightly positive. However, there would remain the barrier of initially getting low-density basaltic material to sufficient depth (and temperature) to transform, and the hotter the mantle the bigger this barrier would be. This question needs to be addressed with quantitative modeling, though this is not straightforward because the sources of plate resistance are not understood in detail. A related situation in the modern Earth, the subduction of oceanic plateaus, has been examined numerically by van Hunen *et al.* (2002), and this approach can be extended to the Archean context.

In the meantime another possible way out of the paradox has emerged. Numerical models of convection in the early, hot, low-viscosity mantle have shown that subducted oceanic crust, in its denser ecologite form, would tend to settle out of the upper mantle, leaving it depleted of basaltic-composition components (Davies, 2006) This would reduce the meltable portion of the upper mantle and therefore reduce the thickness of the oceanic crust. Initial modeling indicated that the oceanic crust might be only a few kilometers thick for a mantle temperature of 1550°C, rather than about 30 km if the upper mantle were as fertile as at present (Davies, 2006). More thorough subsequent testing suggests that the crustal thickness might be 6–10 km in these conditions (Davies, 2007a). This would still be enough to make plate tectonics more viable than the earlier argument suggested. Whether the reduction is enough for mantle to plate tectonics to cool the mantle is not yet clear.

9.08.7.2 Alternatives to Plate Tectonics

If plates were too buoyant to subduct, what would happen? One possibility is that the mantle part of the lithosphere founders and leaves the crust at the surface. Two ways in which this might happen are illustrated in **Figure 11** (Davies, 1992). In both cases the buoyant crust is sheared off as the mantle part of the lithosphere founders. The difference between the cases is that in (1) the thermal boundary layer is still strong enough to behave like a plate, whereas in (2) it is assumed to be deformable. Since asymmetric subduction, as in (1), results from the formation of a fault zone, this may not happen in (2) and the foundering may be symmetric. It is possible in either case that the accumulating basaltic crust becomes thick enough for its lower reaches to transform to eclogite, in which case the whole basaltic body might become unstable and founder, progressively transforming to eclogite as it does so. This might result in tectonic activity being strongly episodic.

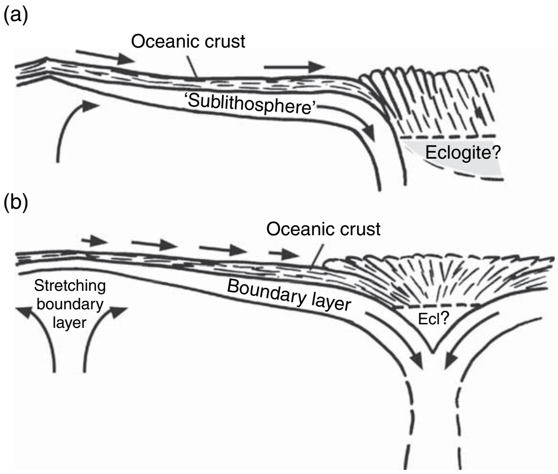

Figure 11 Sketches of possible tectonic modes if the buoyancy of oceanic crust prevents it from subducting with the rest of the lithosphere. (a) Lithosphere still thick enough to behave as a plate, giving rise to asymmetric subduction. (b) Deformable lithosphere, giving rise to symmetric foundering. From Davies GF (1992) On the emergence of plate tectonics. *Geology* 20: 963–966.

9.08.7.3 History of Plumes

The cooling model of **Figure 6** yields a relatively constant rate of heat loss from the core. Since plumes are the expected form of upwelling in the mantle, due to the strong temperature dependence of viscosity (Davies, 2005), this implies a fairly steady level of plume activity through Earth history. On the other hand, higher early core heat losses and greater early plume activity are also possible (**Figure 7**).

Campbell *et al.* (1989) and Campbell and Griffiths (1992) have argued that Archean greenstone belts can be interpreted as the melting products of plume heads, so they would be analogs of Phanerozoic flood basalts. Assuming that the highest-temperature magmas are the best available sample of the plume, Campbell notes that there are two fundamental changes around the end of the Archean. First, the highest-temperature magmas in the Archean are komatiites with inferred source potential temperatures of 1800–1900°C, whereas the hottest post-Archean magmas are picrites with source potential tempatures of 1400–1600°C (**Figure 12**). Second, the Archean komatiites are depleted in incompatible elements, or neutral, whereas the later picrites are enriched.

The higher source temperature implied by komatiites has been widely remarked upon (e.g., Abbott *et al.*, 1994), and some have taken it to indicate a much hotter mantle in the Archean. However, as noted earlier, komatiites comprise only a small fraction of Archean mafic rocks, the vast majority of which are basalts, so a more straightforward interpretation seems to be that most of the mantle was only 100–200°C hotter in the Archean, and the komatiites come from plumes. This implies that plumes were significantly hotter in the Archean than subsequently, and suggests in turn that the models with more rapid early core cooling (**Figure 7**) might be favored. On the other hand, Campbell and Griffiths note that the change from high to low

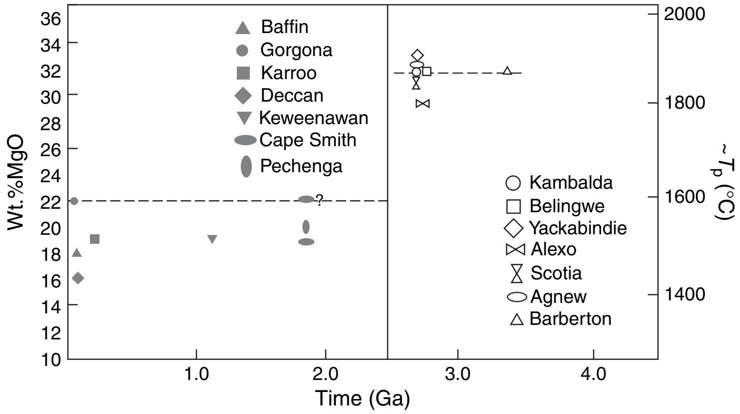

Figure 12 Maximum MgO content of komatiites and picrites versus age ('time'). The approximate mantle potential temperature is shown on the right. From Campbell IH and Griffiths RW (1992) The changing nature of mantle hotspots through time: Implications for the chemical evolution of the mantle. *Journal of Geology* 92: 497–523.

temperature seemed to occur relatively rapidly, between 2.7 and 1.9 Ga (**Figure 12**), suggesting a more discontinuous cause.

Noting the change in trace-element chemistry, from depleted to enriched, Campbell and Griffiths (1992) suggest that there was a change in material accumulating at the base of the mantle, in the D″ zone. Their scenario is sketched in **Figure 13**. The post-Archean regime is like the present, with subducted oceanic crust presumed to settle toward the base and accumulating as a trace-element enriched layer (Hofmann and White, 1982; Hofmann, 1997). On the other hand, they suppose that during the pre-Archean or Hadean (4.5–4 Ga), the mafic crust was too buoyant to founder, and only the underlying mantle part of the thermal boundary layer foundered. Since this material would have been depleted by the extraction of mafic crust and was also cool, it would sink to form a depleted layer at the base of the mantle. The post-Archean change in chemistry is then attributed to the replacement of the early depleted D″ layer by enriched subducted mafic crust. Their proposed cause of the change in temperature is less clear-cut in this scenario. It may be that in the earlier phase the D″ layer covered only part of the core, leaving hot core directly in contact with mantle elsewhere and thus generating very hot plumes. Alternatively, they speculate that a difference in chemistry may lead the later material to become positively buoyant at lower temperatures. In any case the change in character of inferred mantle plumes is attributed ultimately to a change in surface tectonics, which change the nature of the material sinking to the base of the mantle.

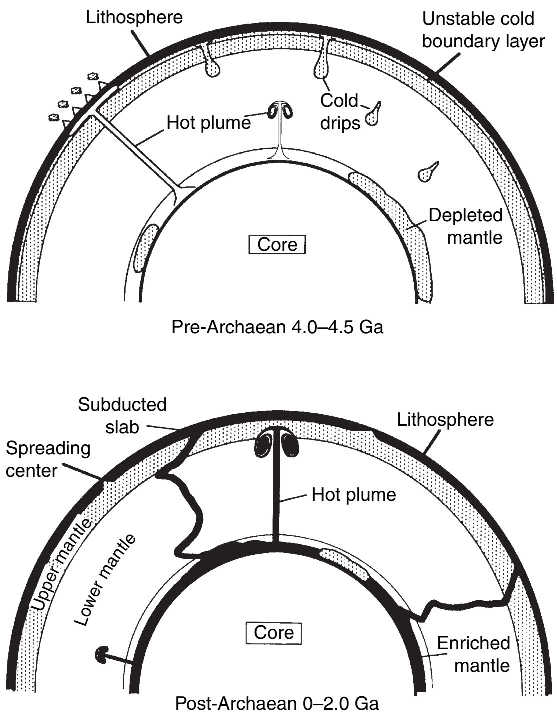

Figure 13 Sketch of proposed styles of mantle convection for pre- and post-Archean. The stippled regions are depleted of incompatible elements by melting. It is supposed that in the pre-Archean the mafic crust renders the upper lithosphere buoyant. The lower lithosphere is depleted, and cools and 'drips' away, accumulating at the base of the mantle. The post-Archean regime features modern-style plates and plumes, with subducted oceanic crust accumulating at the base of the mantle and displacing the previously accumulated depleted material. From Campbell IH and Griffiths RW (1992) The changing nature of mantle hotspots through time: Implications for the chemical evolution of the mantle. *Journal of Geology* 92: 497–523.

9.08.8 Conclusion

Our understanding of Earth's thermal history seems to have been given a firm foundation with the discovery of radioactivity and the realization that the mantle is mobile enough to transport heat by convection. We now have a quantitative and relatively simply theory of cooling since Earth's presumed hot formation by accretion of planetesimals (Safronov, 1978; Wetherill, 1985). However, there remains an important discrepancy between the heating inferred from this theory and the content of radioactive heat sources inferred from cosmochemistry. Whether this reflects an incomplete understanding of the mechanism of heat removal or of Earth's relationship with meteorites remains to be seen.

The strong temperature dependence of the viscosity of the mantle imparts a distinctive character to Earth's cooling history, as it does also to the mantle's two inferred modes of convection, involving plates and plumes (Davies, 1999). This factor allows the mantle to adjust its rate of convection, and therefore of heat transport, by relatively small adjustments in temperature. The result, in 'conventional' thermal history calculations, is a transient phase of rapid cooling, lasting perhaps 500 Ma, followed by a long phase in which the mantle heat loss follows the slow decline of heat sources due to radioactive decay.

The mantle's properties and dynamics also control the cooling of the core, whose heat must be transported away by the mantle. With present uncertainties about the efficiency of this process, there is an ambiguity in core histories, which could involve either relatively constant heat loss or steadily declining heat loss. This ambiguity affects our understanding of the history of mantle plumes and of the energetics involved with maintaining the geodynamo, apparently for at least 3.5 Ga.

One important question remains unresolved, namely whether plate tectonics could have operated in a significantly hotter mantle. There is a mechanism that might have hindered the plates, and there are suggestions of how other effects might have countered this mechanism, but more investigation is required.

The geological record seems to be plausibly, if so far only approximately, consistent with the broad picture presented here. Archean mafic rocks suggest a dominant mantle source perhaps 100–200°C hotter than the present mantle, with relatively uncommon komatiites and picrites suggesting plumes with temperatures 500–600°C hotter than the present mantle during the Archean.

The possibility of substantial deviations from this general picture must still be entertained, and some examples have been mentioned here.

References

Abbott D, Burgess L, Longhi J, and Smith WHF (1994) An empirical thermal history of the Earth's upper mantle. *Journal of Geophysical Research* 99: 13835–13850.

Anderson DL (2004) Simple scaling relations in geodynamics: The role of pressure in mantle convection and plume formation. *Chinese Science Bulletin* 49: 2017–2021.

Barrell J (1914) The strength of the Earth's crust. *Journal of Geology* 22: 655–683.

Bina CR and Helffrich G (1994) Phase transition Clapeyron slopes and transition zone seismic discontinuity topography. *Journal of Geophysical Research* 99: 15853–15860.

Buffett BA (2002) Estimates of heat flow in the deep mantle based on the power requirements for the geodynamo. *Geophysical Research Letters* 29: 1566.

Bunge H-P (2005) Low plume excess temperature and high core heat flux inferred from non-adiabatic geotherms in internally heated mantle circulation models. *Physics of the Earth and Planetary Interiors* 153: 3–10.

Campbell IH and Griffiths RW (1992) The changing nature of mantle hotspots through time: Implications for the chemical evolution of the mantle. *Journal of Geology* 92: 497–523.

Campbell IH, Griffiths R, and Hill RI (1989) Melting in an Archaean mantle plume: Heads it's basalts, tails it's komatiites. *Nature* 339: 697–699.

Cassen P, Reynolds RT, Graziani F, Summers A, McNellis J, and Blalock L (1979) Convection and lunar thermal history. *Physics of the Earth and Planetary Interiors* 19: 183–196.

Clark SP, Jr. (1957) Radiative transfer in the Earth's mantle. *Transactions of the American Geophysical Union* 38: 931–938.

Davaille A and Jaupart C (1993) Transient high-Rayleigh-number thermal convection with large viscosity variations. *Journal of Fluid Mechanics* 253: 141–166.

Davies GF (1977) Whole mantle convection and plate tectonics. *Geophysical Journal of the Royal Astronomical Society* 49: 459–486.

Davies GF (1979) Thickness and thermal history of continental crust and root zones. *Earth and Planetary Science Letters* 44: 231–238.

Davies GF (1980) Thermal histories of convective Earth models and constraints on radiogenic heat production in the Earth. *Journal of Geophysical Research* 85: 2517–2530.

Davies GF (1988) Ocean bathymetry and mantle convection, 1. Large-scale flow and hotspots. *Journal of Geophysical Research* 93: 10467–10480.

Davies GF (1990) Heat and mass transport in the early Earth. In: Newsome HE and Jones JH (eds.) *Origin of the Earth*, pp. 175–194. New York: Oxford University Press.

Davies GF (1992) On the emergence of plate tectonics. *Geology* 20: 963–966.

Davies GF (1993) Cooling the core and mantle by plume and plate flows. *Geophysical Journal International* 115: 132–146.

Davies GF (1995a) Penetration of plates and plumes through the mantle transition zone. *Earth and Planetary Science Letters* 133: 507–516.

Davies GF (1995b) Punctuated tectonic evolution of the Earth. *Earth and Planetary Science Letters* 136: 363–379.

Davies GF (1999) *Dynamic Earth: Plates, Plumes and Mantle Convection*. Cambridge: Cambridge University Press.

Davies GF (2005) A case for mantle plumes. *Chinese Science Bulletin* 50: 1541–1554.

Davies GF (2006) Gravitational depletion of the early Earth's upper mantle and the viability of early plate tectonics. *Earth and Planetary Science Letters* 243: 376–382.

Davies GF (2007a) Controls on density stratification in the early Earth. *Geochemistry*, *Geophysics*, *Geosystems* (in press).

Davies GF (2007b) Mantle regulation of core cooling: A geodynamo without core radioactivity? *Physics of the Earth and Planetary Interiors* 160: 215–229.

Fisher O (1881) *Physics of the Earth's Crust*. London: Murray.

Foulger GR (2005) Mantle plumes: Why the current skepticism? *Chinese Science Bulletin* 50: 1555–1560.

Galer SJG, Goldstein SL, and O'Nions RK (1989) Limits on chemical and convective isolation in the Earth's interior. *Chemical Geology* 75: 257–290.

Goldreich P and Toomre A (1969) Some remarks on polar wandering. *Journal of Geophysical Research* 74: 2555–2567.

Grand S, van der Hilst RD, and Widiyantoro S (1997) Global seismic tomography: A snapshot of convection in the Earth. *Geological Society of America Today* 7: 1–7.

Grasset O and Parmentier EM (1998) Thermal convection in a volumetrically heated, infinite Prandtl number fluid with strongly temperature-dependent viscosity: Implications for planetary thermal evolution. *Journal of Geophysical Research* 103: 18171–18,181.

Hager BH (1984) Subducted slabs and the geoid: Constraints on mantle rheology and flow. *Journal of Geophysical Research* 89: 6003–6015.

Hallam A (1989) *Great Geological Controversies*. Oxford: Oxford University Press.

Harrison TM, Blichert-Toft J, Müller W, Albarède F, Holden P, and Mojzsis SJ (2005) Heterogeneous Hadean hafnium: Evidence of continental crust at 4.4 to 4.5 Ga. *Science* 310: 1947–1950.

Hirth G and Kohlstedt DL (1996) Water in the oceanic upper mantle: Implications for rheology, melt extraction and the evolution of the lithosphere. *Earth and Planetary Science Letters* 144: 93–108.

Hofmann AW (1997) Mantle chemistry: The message from oceanic volcanism. *Nature* 385: 219–229.

Hoffman PF and Bowring SA (1984) Short-lived 1.9 Ga continental margin and its destruction, Wopmay orogen, northwest Canada. *Geology* 12: 68–72.

Hofmann AW and White WM (1982) Mantle plumes from ancient oceanic crust. *Earth and Planetary Science Letters* 57: 421–436.

Hofmeister AM (2005) Dependence of diffusive radiative transfer on grain size, temperature, and Fe-content: Implications. *Journal of Geodynamics* 40: 51–72.

Holmes A (1911) *Proceedings of the Royal Society of London A* 85: 248.

Holmes A (1913) *The Age of the Earth*. London: Harper & Brothers.

Honda S (1995) A simple paramaterized model of Earth's thermal history with the transition from layered to whole mantle convection. *Earth and Planetary Science Letters* 131: 357–369.

Honda S, Yuen DA, Balachandar S, and Reuteler D (1993) Three-dimensional instabilities of mantle convection with multiple phase transitions. *Science* 259: 1308–1311.

Isacks B, Oliver J, and Sykes LR (1968) Seismology and the new global tectonics. *Journal of Geophysical Research* 73: 5855–5899.

Jeffreys H (1976) *The Earth, Its Origin, History and Physical Constitution*. Cambridge: Cambridge University Press.

Jochum KP, Hofmann AW, Ito E, Seufert HM, and White WM (1983) K, U and Th in mid-ocean ridge basalt glasses and heat production, K/U and K/Rb in the mantle. *Nature* 306: 431–436.

Kellogg LH, Hager BH, and van der Hilst RD (1999) Compositional stratification in the deep mantle. *Science* 283: 1881–1884.

Kelvin L (1862) On the age of the sun's heat. *Macmillans Magazine* 5: 288.

Kelvin L (1863) On the secular cooling of the Earth. *Philosophical Magazine (Series 4)* 25: 1.

Kelvin L (1899) The age of the Earth as an abode fitted for life. *Journal of Victoria Institute* 31: 11.

Korenaga J (2006) *The Archean Earth*. Washington, DC: American Geophysical Union.

Kröner A (*ed*.) (1981) *Precambrian Plate Tectonics*, pp. 59–60. Amsterdam: Elsevier.

Labrosse S and Macouin M (2003) The inner core and the geodynamo. *Comptes Rendus Geoscience* 335: 37–50.

Machatel P and Weber P (1991) Intermittent layered convection in a model mantle with an endothermic phase change at 670 km. *Nature* 350: 55–57.

Manga M, Weeraratne D, and Morris SJS (2001) Boundary layer thickness and instabilities in Benard convection of a liquid with temperature-dependent viscosity. *Physics of the Fluids* 13: 802–805.

McCulloch MT and Bennett VC (1994) Progressive growth of the Earth's continental crust and depleted mantle: Geochemical constraints. *Geochimica et Cosmochimica Acta* 58: 4717–4738.

McDonough WF and Sun S-s (1995) The composition of the Earth. *Chemical Geology* 120: 223–253.

McGovern PJ and Schubert G (1989) Thermal evolution of the Earth: Effects of volatile exchange between atmosphere and interior. *Earth and Planetary Science Letters* 96: 27–37.

McKenzie DP and Bickle MJ (1988) The volume and composition of melt generated by extension of the lithosphere. *Journal of Petrology* 29: 625–679.

McKenzie DP, Roberts JM, and Weiss NO (1974) Convection in the Earth's mantle: Towards a numerical solution. *Journal of Fluid Mechanics* 62: 465–538.

McKenzie DP and Weiss N (1975) Speculations on the thermal and tectonic history of the Earth. *Geophysical Journal of the Royal Astronomical Society* 42: 131–174.

Mitrovica JX (1996) Haskell (1935) revisited. *Journal of Geophysical Research* 101: 555–569.

Mitrovica JX and Forte AM (1997) Radial profile of mantle viscosity: Results from the joint inversion of convection and postglacial rebound observables. *Journal of Geophysical Research* 102: 2751–2769.

Nakakuki T, Sato H, and Fujimoto H (1994) Interaction of the upwelling plume with the phase and chemical boundary at the 670 km discontinuity: Effects of temperature-dependent viscosity. *Earth and Planetary Science Letters* 121: 369–384.

Nimmo F (2007) Thermal and compositional evolution of the core this volume.

Nimmo F, Price GD, Brodholt J, and Gubbins D (2004) The influence of potassium on core and geodynamo evolution. *Geophysical Journal International* 156: 363–376.

O'Connell RJ (1977) On the scale of mantle convection. *Tectonophysics* 38: 119–136.

Palme H and O'Neill HSC (2004) Cosmochemical estimates of mantle composition. In: Carlson RW, Holland HD, and Turekian KK (eds.) *Treatise on Geochemistry*, pp. 1–38. Oxford: Elsevier.

Patterson C (1956) Age of meteorites and of the Earth. *Geochimica et Cosmochimica Acta* 10: 230–237.

Ratcliffe JT, Tackley PJ, Schubert G, and Zebib A (1997) Transitions in thermal convection with strongly variable viscosity. *Physics of the Earth and Planetary Interiors* 102: 201–212.

Roberts PH, Jones CA, and Calderwood AR (2003) *Earth's Core and Lower Mantle*. London: Taylor and Francis.

Rossby HT (1969) A study of Benard convection with and without rotation. *Journal of Fluid Mechanics* 36: 309–336.

Rudnick RL and Fountain DM (1995) Nature and composition of the continental crust: A lower crustal perspective. *Reviews of Geophysics* 33: 267–309.

Safronov VS (1978) The heating of the Earth during its formation. *Icarus* 33: 8–12.

Schubert G (1979) Subsolidus convection in the mantles of terrestrial planets. *Annual Review of Earth and Planetary Sciences* 7: 289–342.

Schubert G, Cassen P, and Young RE (1979a) Core cooling by subsolidus mantle convection. *Physics of the Earth and Planetary Interiors* 20: 194–208.

Schubert G, Cassen P, and Young RE (1979b) Subsolidus convective cooling histories of terrestrial planets. *Icarus* 38: 192–211.

Schubert G, Stevenson DJ, and Cassen P (1980) Whole planet cooling and the radiogenic heat source contents of the Earth and moon. *Journal of Geophysical Research* 85: 2531–2538.

Schubert G, Turcotte DL, and Olson P (2001) *Mantle Convection in the Earth and Planets*. Cambridge: Cambridge University Press.

Schubert G, Yuen DA, and Turcotte DL (1975) Role of phase transitions in a dynamic mantle. *Geophysical Journal of the Royal Astronomical Society* 42: 705–735.

Sclater JG, Jaupart C, and Galson D (1980) The heat flow through the oceanic and continental crust and the heat loss of the Earth. *Reviews of Geophysics* 18: 269–312.

Shankland TJ, Nitsan U, and Duba AG (1979) Optical absorption and radiative heat transport in olivine at high temperature. *Journal of Geophysical Research* 84: 1603–1610.

Sharpe HN and Peltier WR (1978) Paramaterized mantle convection and the Earth's thermal history. *Geophysical Research Letters* 5: 737–740.

Sharpe HN and Peltier WR (1979) A thermal history model for the Earth with parameterised convection. *Geophysical Journal of the Royal Astronomical Society* 59: 171–203.

Sleep NH (1990) Hotspots and mantle plumes: Some phenomenology. *Journal of Geophysical Research* 95: 6715–6736.

Solheim LP and Peltier WR (1994a) Avalanche effects in phase transition modulated thermal convection: A model of Earth's mantle. *Journal of Geophysical Research* 99: 6997–7018.

Solheim LP and Peltier WR (1994b) Correction to Avalanche effects in phase transition modulated thermal convection: A model of Earth's mantle. *Journal of Geophysical Research* 99: 18203.

Solomatov VS (1995) Scaling of temperature- and stress-dependent viscosity convection. *Physics of the Fluids* 7: 266–274.

Spohn T and Schubert G (1982) Modes of mantle convection and the removal of heat from the Earth's interior. *Journal of Geophysical Research* 87: 4682–4696.

Stacey FD (1980) The cooling earth: A reappraisal. *Physics of the Earth and Planetary Interiors* 22: 89–96.

Stacey FD (1992) *Physics of the Earth*. Brisbane: Brookfield Press.

Stacey FD and Loper DE (1984) Thermal histories of the core and mantle. *Physics of the Earth and Planetary Interiors* 36: 99–115.

Stern RJ (2005) Evidence from ophiolites, blueschists, and ultrahigh-pressure metamorphic terranes that the modern episode of subduction tectonics began in Neoproterozoic time. *Geology* 33: 557–560.

Stevenson DJ and Schubert G (1983) Magnetism and thermal evolution of the terrestrial planets. *Icarus* 54: 466–489.

Stevenson DJ and Turner JS (1979) Fluid models of mantle convection. In: McElhinny M (ed.) *The Earth, Its Origin, Evolution and Structure*, pp. 227–263. New York: Academic Press.

Tackley PJ, Stevenson DJ, Glatzmaier GA, and Schubert G (1993) 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.

Tackley PJ, Xie S, Nakagawa T, and Hernlund JW (2005) Numerical and laboratory studies of mantle convection: Philosophy, accomplishments, and thermochemical structure and evolution. In: van der Hilst RD, Bass JD, Matas J, and Trampert J (eds.) *Earth's Deep Mantle: Structure, Compositon, and Evolution*, pp. 83–99. Washington, DC: American Geophysical Union.

Tozer DC (1965) Heat transfer and convection currents. *Philosophical Transactions of the Royal Society of London, Series A* 258: 252–271.

Tozer DC (1972) The present thermal state of the terrestrial planets. *Physics of the Earth and Planetary Interiors* 6: 182–197.

Turcotte DL and Oxburgh ER (1967) Finite amplitude convection cells and continental drift. *Journal of Fluid Mechanics* 28: 29–42.

Turcotte DL and Schubert G (2001) *Geodynamics: Applications of Continuum Physics to Geological Problems*. Cambridge: Cambridge University Press.

van Hunen J, van den Berg AP, and Vlaar NJ (2002) On the role of subducting oceanic plateaus in the development of shallow flat subduction. *Tectonophysics* 352: 317–333.

Weinstein S (1995) The effects of a deep mantle endothermic phase change on the structure of thermal convection in silicate planets. *Journal of Geophysical Research* 100: 11719–11728.

Wetherill GW (1985) Occurrence of giant impacts during the growth of the terrestrial planets. *Science* 228: 877–879.