Invited review

doi: 10.1111/j.1365-3121.2008.00843.x

Plate tectonics, flood basalts and the evolution of Earth's oceans

Jun Korenaga

Department of Geology and Geophysics, Yale University, PO Box 208109, New Haven, CT 06520-8109, USA

ABSTRACT

The chemical composition of the bulk silicate Earth (BSE) indicates that the present-day thermal budget of Earth is likely to be characterized by a significant excess of surface heat loss over internal heat generation, indicating an important role of secular cooling in Earth's history. When combined with petrological constraints on the degree of secular cooling, this thermal budget places a tight constraint on permissible heat-flow scaling for mantle convection, along with implications for the operation of plate tectonics on Earth, the history of mantle plumes and flood basalt magmatism, and the origin and evolution of Earth's oceans. In the presence of plate tectonics, hotter mantle may have convected more slowly because it generates thicker dehydrated lithosphere, which could slow down subduction. The intervals of globally synchronous orogenies are consistent with the predicted variation of plate velocity for the last 3.6 Gyr. Hotter mantle also produces thicker, buoyant basaltic crust, and the subductability of oceanic lithosphere is a critical factor regarding the emergence of plate tectonics before the Proterozoic. Moreover, sluggish convection in the past is equivalent to reduced secular cooling, thus suggesting a more minor role of mantle plumes in the early Earth. Finally, deeper ocean basins are possible with slower plate motion in the past, and Earth's oceans in the Archean is suggested to have had about twice as much water as today, and the mantle may have started as dry and have been gradually hydrated by subduction. The global water cycle may thus be dominated by regassing, rather than degassing, pointing towards the impact origin of Earth's oceans, which is shown to be supported by the revised composition of the BSE.

Terra Nova, 20, 419–439, 2008

Introduction

The thermal history of Earth, after the initial period of global magma ocean, is governed by the balance of surface heat loss and internal heat generation. An initially hot Earth has been gradually cooled down with time because, for most of Earth's history, surface heat loss has been greater than internal heat supply by radiogenic isotopes. This simple energy balance places first-order constraints on various aspects of physical and chemical processes operating on the surface of Earth and within its deep interior. There have been a number of theoretical studies on Earth's thermal history (e.g. McKenzie and Weiss, 1975; Davies, 1980; Schubert *et al.*, 1980; Christensen, 1985; Richter, 1985; Solomatov, 2001; Korenaga, 2003; Grigne *et al.*, 2005; Labrosse and Jaupart, 2007), and a review on so far published models on Earth's thermal evolution has recently been published (Korenaga, 2008b). Estimating a thermal history involves both geophysics (the physics of convective heat loss) and geochemistry (the abundance of heat-producing elements), and many published models do not account for both aspects simultaneously. Available geochemical and geological data place a strict constraint on the permissible range of thermal history, and the most likely scenario is briefly reviewed in the next section.

Correspondence: Jun Korenaga, Department of Geology and Geophysics, Yale University, PO Box 208109, New Haven, CT 06520-8109, USA. Tel.: +1 (203) 432 7381; fax: +1 (203) 432 3134; e-mail: jun.korenaga@yale.edu

This article was commissioned by the Editors of *Terra Nova* to mark the International Year of Planet Earth.

The purpose of this review article is to discuss some of major geological processes that are directly related to the thermal evolution of Earth, on the basis of this latest understanding. Topics covered include the onset of plate tectonics, the history of mantle plumes, continental growth, and the origin and evolution of Earth's oceans. They are difficult problems, and little can be said with certainty at present. Because they are intimately coupled with thermal evolution, however, we can still derive a few important constraints. These constraints may be counter-intuitive or not obvious at first sight. Mantle plumes, for example, are often thought to be more active in the past, but such temporal trend is in direct conflict with what thermal evolution indicates. Also, the volume of oceans is unlikely to be constant, and it is probably decreasing with time. Whereas these topics tend to be considered independently, each of them represents a different aspect on the evolution of the integrated Earth system, so a proper understanding of Earth's thermal history is essential to make self-consistent predictions about them.

Thermal history of earth

Global heat balance equations

The thermal history of Earth may be modelled by integrating the following global energy balance (e.g. Stevenson *et al.*, 1983):

$$C_m \frac{dT_m(t)}{dt} = H_m(t) - Q(t) + Q_{CMB}(t), \quad (1)$$

and

$$C_c \frac{dT_c(t)}{dt} = (L + E_g)\frac{dm_{ic}(t)}{dt} + H_c(t) - Q_{CMB}(t), \quad (2)$$

where the subscripts m and c denote the mantle and core components, respectively, C_i is heat capacity, $T_i(t)$ is average temperature, and $H_i(t)$ is internal heat production. Core heat flux, $Q_{CMB}(t)$, represents heat exchange between the mantle and the core, and $Q(t)$ is the surface heat flux. The mass of the inner core is denoted by $m_{ic}(t)$, and the first term on the right-hand side of Eq. (2) describes the release of latent heat and gravitational energy associated with the growth of the inner core. As mantle evolution is tracked by a single average temperature, this formulation corresponds to whole-mantle convection. Though layered-mantle convection has been a popular concept (e.g. Jacobsen and Wasserburg, 1979; Richter, 1985; Allegre *et al.*, 1996; Kellogg *et al.*, 1999), whole-mantle convection is probably sufficient to explain available geophysical and geochemical data if the uncertainties of these data and the limitation of theoretical predictions are taken into account (Lyubetskaya and Korenaga, 2007b; Korenaga, 2008b). Note that layered mantle convection is still *possible*. Because it is not well defined (Korenaga, 2008b), however, exploring its implication for Earth's evolution would require us to deal with more degrees of freedom. For simplicity, therefore, we restrict ourselves to whole-mantle convection in this review, which should serve as a reference when considering more complex convection models (Table 1 presents a summary of major assumptions made in this article).

Internal heat production in the core is controversial (e.g. Gessmann and Wood, 2002; McDonough, 2003; Rama Murthy *et al.*, 2003; Lee *et al.*, 2004; Lassiter, 2006), but even if the core contains potassium at 100 p.p.m. level, H_c would be only ~0.7 TW at present. The energy release due to the inner core growth is similarly small, being on the order of ~2 TW (e.g. Stevenson *et al.*, 1983). Thus, the first two terms on the right-hand side of Eq. (2) may be neglected for simplicity, and Eqs (1) and (2) may be combined to yield:

$$C\frac{dT_m(t)}{dt} \approx H_m(t) - Q(t) + C_c\frac{d\Delta T_{CMB}(t)}{dt}, \quad (3)$$

where C is the heat capacity of the entire Earth [$= C_m + C_c$, $\sim 7 \times 10^{27}$ J K^{-1} (Stacey, 1981)] and $\Delta T_{CMB}(t)$ is the temperature contrast at the core–mantle boundary. The present-day contrast is estimated to be on the order of 1000 K (e.g. Boehler, 1996; Williams, 1998), but how it should change with time is not understood well. To estimate the temporal variability of the temperature contrast, we need to calculate the core heat flux $Q_{CMB}(t)$, which depends critically on the poorly known rheology of the lowermost mantle (e.g. Solomatov, 1996; Karato, 1998; Korenaga, 2005a) and also on other complications such as the presence of chemical heterogeneities (e.g. Farnetani, 1997; Jellinek and Manga, 2002), the post-perovskite phase transition (e.g. Murakami *et al.*, 2004; Oganov and Ono, 2004), and drastic changes in thermal properties (e.g. Badro *et al.*, 2004; Lin *et al.*, 2005). We do not even know why the present-day contrast happens to be ~1000 K. Thus, it may be reasonable to treat $d\Delta T_{CMB}(t)/dt$ as a free parameter and see how thermal evolution would be affected by this term. As the simplest option, it is set to zero here, and other possibilities will be explored later (see Core heat flux and the possibility of superheated core). With these simplifying assumptions, the global heat balance can be expressed as:

$$C\frac{dT_m(t)}{dt} = H_m(t) - Q(t). \quad (4)$$

Table 1 List of major assumptions employed in this article.

Assumption	Relevant section
Earth's mantle convects as a single layer	Thermal History of Earth
Internal heat production in the core is negligible	Thermal History of Earth
Mantle dehydration by melting controls global mantle dynamics	Thermal History of Earth
Mean sea level has always been close to mean continental level	History of Ocean Volume
Seafloor age–area relationship follows a triangular distribution	History of Ocean Volume
Zero-age depth of seafloor has been approximately constant	History of Ocean Volume
Average thickness of continental crust has been approximately constant	History of Ocean Volume

As the physical state of the early Earth is uncertain, the above differential equation is integrated backward in time, starting with the present-day condition.

Present-day thermal budget

The present-day thermal budget, i.e. $Q(0)$ and $H_m(0)$, may be estimated as follows. The global heat flux is estimated to be 46 ± 3 TW (Jaupart *et al.*, 2007), which is the sum of heat flux due to mantle convection, $Q(0)$, and radiogenic heat production in continental crust, $H_{cc}(0)$. The latter is estimated to be 7.5 ± 2.5 TW (Rudnick and Gao, 2003), so we may write

$$H_{cc}(0) = 7.5 + 2.5\varepsilon_1 \quad [\mathrm{TW}], \quad (5)$$

and

$$Q(0) = 46 - H_{cc}(0) + 3\varepsilon_2 \quad [\mathrm{TW}], \quad (6)$$

where ε_1 and ε_2 are random variables following the Gaussian distribution of zero mean and unit standard deviation. The use of two independent random variables signifies that the uncertainty of the global heat flux and that of continental heat production are not correlated. Assuming whole-mantle convection, heat production in the bulk silicate Earth (BSE) (i.e. crust and mantle) is estimated to be 16 ± 3 TW (Lyubetskaya and Korenaga, 2007b), so heat production in the convecting mantle may be expressed as:

$$H_m(0) = 16 - H_{cc}(0) + 3\varepsilon_3 \quad [\mathrm{TW}], \quad (7)$$

where ε_3 is another random variable following the same Gaussian distribution, and the 'convective Urey ratio' (Korenaga, 2008b) is defined as:

$$\mathrm{Ur}(0) = H_m(0)/Q(0). \quad (8)$$

These uncertainties are relatively large, allowing unrealistic values [e.g. negative values for $H_m(0)$], so the following additional constraints are imposed: 5 TW ≤ $H_{cc}(0)$ ≤ 14 TW and $H_m(0)$ ≥ 3 TW. These *a priori* bounds are based on published composition models for the continental crust and the convecting mantle (e.g. Jochum *et al.*, 1983; Taylor and McLennan, 1985; Wedepohl, 1995; Rudnick and Gao, 2003; Salters and Stracke, 2004; Workman and Hart, 2005). The resulting probability distribution functions are shown in Fig. 1. It can

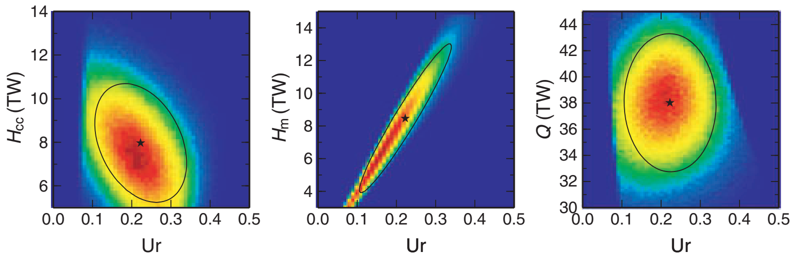

Fig. 1 Joint probability distribution functions for continental crust heat production $H_{cc,}$ mantle heat production H_m, convective heat flux Q, and the convective Urey ratio Ur, based on Eqs (5)–(8) with the *a priori* bounds described in the text. Brighter colour denotes higher probability. Stars correspond to mean values, and ellipses to 68% confidence regions assuming the normal distribution.

be seen that $H_{cc}(0) \sim 8$ TW, $H_m(0) \sim 8.5$ TW, $Q(0) \sim 38$ TW, and $\mathrm{Ur}(0) \sim 0.22$. For the sake of simplicity, thermal evolution models in this paper are all calculated with these mean values, but it is important to keep in mind that the present-day thermal budget has non-trivial uncertainties and that some uncertainties are strongly correlated [e.g. $H_m(0)$ and $\mathrm{Ur}(0)$]. See Korenaga (2008b) for the effects of such uncertainties on the prediction of thermal histories.

Internal heating and surface heat flux

At present, the convecting mantle contains only ~50% of heat-producing elements in the BSE, but it must have had a larger fraction when the mass of continental crust was smaller in the past. This may be expressed as:

$$H_m(t) = H_{m0}(t) + (1 - F_C(t))H_{cc}(t), \tag{9}$$

where $H_{m0}(t)$ and $H_{cc}(t)$ denote the (hypothetical) evolution of internal heating in the mantle and the continental crust, respectively, without mass transfer between them, and $F_C(t)$ is the mass fraction of continental crust with respect to the present-day value [assuming $F_C(t) \le 1$ for all t]. The calculation of these internal heating functions is straightforward once the present-day values are specified (e.g. Spohn and Breuer, 1993; Grigne and Labrosse, 2001).

The evolution of surface heat flux $Q(t)$ has to be estimated through heat-flow scaling for mantle convection as there are no direct observations on global heat flow in the past. This scaling issue has been controversial (e.g. Schubert *et al.*, 1980; Christensen, 1985; Gurnis, 1989; Solomatov, 1995; Conrad and Hager, 1999b; Korenaga, 2003). Figure 2 shows three representative scaling laws (see Korenaga (2006) for their derivations). The 'conventional' scaling predicts higher heat flux for hotter mantle, and the temperature dependency is dictated by the activation energy of mantle viscosity [a few hundred kJ mol^{-1}, (Karato and Wu, 1993)]. Even in the limit of zero activation energy ('isoviscous convection' in Fig. 2), the temperature dependency of heat flow is still positive because heat flow is also proportional to the temperature difference between the surface and the interior. It has been repeatedly shown that the positive temperature dependency does not produce a sensible thermal history consistent with the present-day thermal budget (e.g. Christensen, 1985; Korenaga, 2003). One could resolve this problem by postulating a high Urey ratio at present (e.g. ~0.7) (Davies, 1980, 1993, 2007; Schubert *et al.*, 1980; Schubert and Reymer, 1985; Williams and Pan, 1992; McNamara and van Keken, 2000), but such high Urey ratio appears to be in conflict with geochemical data (McDonough and Sun, 1995; Lyubetskaya and Korenaga, 2007a). Future antineutrino data should give direct constraints on mantle radioactivity (Araki *et al.*, 2005; McDonough, 2005). Recently, there were some attempts to rescue the conventional scaling by invoking temporal fluctuations in plate tectonics (e.g. Grigne *et al.*, 2005; Silver and Behn, 2008), but these studies have been suggested to be inconsistent either with known sea level changes (Korenaga, 2007a) or with the physics of heat transfer (Korenaga, 2008a). It is also important to understand what underlies the conventional scaling (e.g. Howard, 1966);

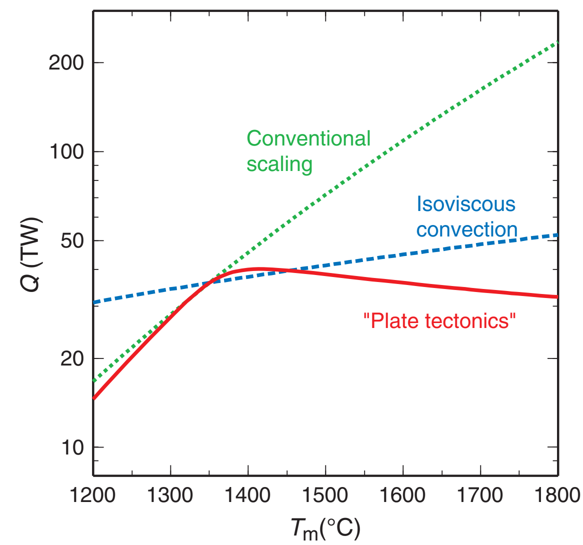

Fig. 2 Three classes of heat-flow scaling for mantle convection. 'Conventional' and 'isoviscous' parametrizations are both based on the heat-flow scaling of simple thermal convection. Mantle viscosity is temperature-dependent for the former, with the activation energy of 300 kJ mol^{-1}, and it is constant for the latter. 'Plate tectonics' scaling incorporates the effects of mantle melting beneath mid-ocean ridges such as dehydration stiffening and compositional buoyancy. See Korenaga (2006) for derivations.

hotter mantle is predicted to yield higher heat flux because the boundary layer (i.e. plates) becomes thinner owing to greater convective instability. Conventional scaling and substantial boundary-layer growth are thus mutually exclusive, but mixing these opposing concepts seems to be required if intermittent plate tectonics were to moderate surface heat flux.

It appears, therefore, that some kind of *negative* temperature dependency is essential to prevent the so-called thermal catastrophe (Fig. 3a), and such scaling may be possible if the effects of grain growth kinetics (Solomatov, 1996, 2001) or mantle melting associated with plate tectonics (Korenaga, 2003, 2006) are important for global mantle dynamics. Whereas the importance of grain growth kinetics in mantle convection (particularly in the lower mantle) is not well understood yet, heat-flow scaling with mantle melting depends mainly on upper mantle properties, which are reasonably well known. Thus, the scaling of Korenaga (2006) ('plate tectonics' in Fig. 3) is adopted for thermal history calculations here. Note that the concept of hotter and stiffer mantle (Solomatov, 1996) could enhance the negative temperature dependency further. As will be suggested later, the mantle may have been drier on average, which would also help suppressing convective heat flux from a hotter mantle in the past. These suggestions for unconventional heat-flow scaling are far from being established and subject to further investigation, including fully self-consistent numerical modelling. Important points are that there is no *a priori* reason for Earth's mantle to follow the conventional scaling, and that we now have some physically plausible mechanisms that may modify drastically the apparent temperature sensitivity of convective heat flux.

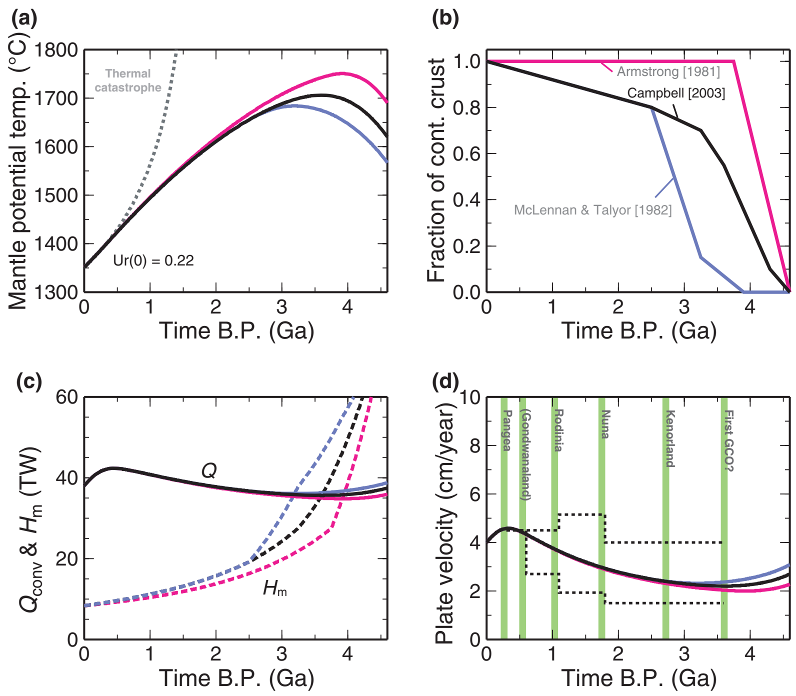

Fig. 3 Thermal evolution modelling with continental growth, starting with the present-day mantle temperature of 1350 °C and the present-day convective Urey ratio of 0.22. The global heat balance of Eq. (4) is assumed, and the details of integration procedure can be found in Korenaga (2006). The heat-flow scaling of plate tectonics (Fig. 2) is adopted. (a) The history of mantle potential temperature T_m. The result with the conventional heat-flow scaling is also shown by dashed line. (b) Three models of continental growth: 'instantaneous' growth (Armstrong, 1981), 'gradual' growth (McLennan and Taylor, 1982), and somewhere in-between (Campbell, 2003). (c) Predicted history of convective heat flux (solid) and mantle internal heating (dashed). Different colours correspond different models of continental growth. It may be seen that the Urey ratio, $\mathrm{Ur}(t) = H_m(t)/Q(t)$, gradually increases from the present-day value of 0.22 to higher values in the past. (d) Predicted history of average plate velocity. Dashed lines denote the range of geological estimate based on the intervals between supercontinent assemblies or globally synchronous orogenies (Korenaga, 2006), the timing of which are shown by green bars (Hoffman, 1997; Nutman, 2006).

The calculated thermal history is presented with three different models of continental growth (Fig. 3). Table 2 lists key model parameters and their (present-day) values adopted here. The present-day mantle potential temperature is assumed to be 1350 °C (e.g. Kinzler, 1997; Herzberg *et al.*, 2007). It can be observed that the detail of continental growth is not important; radically different growth models give rise to only ~100 K difference in the early Earth (Fig. 3a). Perhaps the most important feature is the robustness of model predictions owing to a negative feedback implemented by the adopted heat-flow scaling. Lower heat flux for hotter mantle implies that internal heating may have been higher than surface heat flux sometime in the past (Fig. 3c), and also that plate tectonics was more sluggish in the past (Fig. 3d).

Before interpreting these predictions any further, at least two issues need to be discussed. The first one is the validity of the assumed heat-flow scaling for plate tectonics. The physics of plate tectonics is not fully understood yet (e.g. Bercovici *et al.*, 2000; Bercovici, 2003). We still do not know, for example, why Earth exhibits plate tectonics and other terrestrial planets do not. There are various suggestions (e.g. Tozer, 1985; Regenauer-Lieb *et al.*, 2001; Korenaga, 2007b), many of which call for the existence of surface water, but we do not have a quantitative understanding for under what conditions plate tectonics can take place. Without such fundamental understanding, then, why should we trust the proposed scaling for plate tectonics? The answer lies in the nature of scaling laws. Heat-flow scaling prescribes the sensitivity of surface heat flow to variations in mantle temperature, and all of scaling laws in Fig. 2 are normalized with respect to the present-day convective heat flow. We know plate tectonics is operating today, and we can estimate

Table 2 Key parameters used in mantle and ocean evolution modelling.

Parameter	Definition	Value*	Note
C	Heat capacity of the entire Earth	7×10^{27} J K^{-1}	Eqs (3) and (4)
$T_m(t)$	Mantle potential temperature	1350 °C	Eqs (3) and (4)
$H_m(t)$	Heat production in the mantle†	~8.5 TW	Eqs (3), (4) and (7); Fig. 1
$Q(t)$	Convective heat flux‡	~38 TW	Eqs (3), (4) and (6); Fig. 1
C_c	Heat capacity of the core	1.3×10^{27} J K^{-1}	Eq. (3)
$\Delta T_{CMB}(t)$	Temperature contrast at CMB§	~1000 K	Eq. (3)
$H_{cc}(t)$	Heat production in continental crust¶	~8 TW	Eq. (9); Fig. 1
$\tau_{max}(t)$	Maximum age of seafloor**	180 Ma	Eq. (23)
$v(t)$	Average plate velocity††	4 cm yr^{-1}	Eq. (23)
$A_o(t)$	Total area of ocean basins	3.1×10^{14} m^3	Eqs (16) and (23)
$G(t)$	Plate creation rate	3.45 km^2 yr^{-1}	Eqs (22), (15) and (23)
$V(t)$	Total ocean volume	1.51×10^{18} m^3	Eq. (18)
$d_0(t)$	Zero-age depth of seafloor	2654 m‡‡	Eq. (19)
$b(t)$	Seafloor subsidence rate	323 m Myr$^{-1/2}$	Eqs (19) and (20)

*For parameters that depend on time t, present-day values are listed.

†Past heat production is calculated based on the present-day value and U : Th : K of 1 : 4 : 1.27×10^4 [see Eq. (3) and Table 1 of Korenaga (2006)]. The effect of continental growth is taken into account by Eq. (9).

‡Past convective heat flux is calculated using the 'plate tectonics' scaling law shown in Fig. 2.

§This is assumed to be constant in Figs 3, 4 and 9. The effect of its possible temporal variation on mantle evolution is minor (see Section on Core heat flux and the possibility of superheated core), though it is important for the thermal history of the core (Fig. 5).

¶Past heat production in the continental crust is calculated in a similar way to that in the mantle but with U : Th : K of 1 : 5 : 10^4.

**The triangular distribution of seafloor age-depth is assumed [Eq. (15)].

††Past plate velocity is calculated using its relationship to convective heat flux, $v \propto (Q/T_m)^2$ [Eq. (18) of Korenaga (2006)].

‡‡This is assumed to be constant in Fig. 9.

how potential energy release is balanced by viscous dissipation for the present-day mantle (e.g. Conrad and Hager, 1999a). Heat-flow scaling is constructed by considering how this energy balance should change for different temperatures. As far as plate tectonics is operating, therefore, the assumed heat flow scaling is unlikely to be grossly in error, though the details of viscous dissipation in the mantle could be debatable. This brings up the second issue, which is the onset of plate tectonics on Earth. Figure 3 shows the backward integration of global heat balance to the beginning of Earth's history (4.6 Ga), and this is equivalent to assuming that plate tectonics started as soon as Earth was created. However, if plate tectonics started at 3 Ga, for example, model predictions for older times bear little significance. This issue is discussed in detail next.

Onset of plate tectonics

Geological and geochemical evidence

The assembly of continental masses and associated collisional processes require the closure of ocean basins that existed between different continents. The subduction of oceanic lithosphere should be involved in closing ocean basins, so the timing of ancient orogenies provides important constraints on the first emergence of plate tectonics on Earth. The oldest 'supercontinent' Kenorland was assembled at *c.* 2.7 Ga (Hoffman, 1997), so it would be reasonable to assume the operation of plate tectonics in the late Archean. Whether Kenorland was a true supercontinent or not (Bleeker, 2003) is not so important in the current context; there is field evidence for globally synchronous orogeny at *c.* 2.7 Ga, which is sufficient for discussion here. Also, recycling of oceanic crust by subduction back to *c.* 3 Ga is supported by the age of eclogite xenoliths from subcontinental lithospheric mantle (e.g. Jacob *et al.*, 1994; Shirey *et al.*, 2001). As noted by Korenaga (2006), the time intervals of so far known five supercontinents (Pangea, Gondwanaland, Rodinia, Nuna and Kenorland) may be used to estimate the spatially and temporally averaged rate of plate motion, and such estimate appears to be consistent with the prediction based on thermal evolution modelling (Fig. 3d). One could question this connection between the history of supercontinents and the vigour of plate tectonics by suggesting that plate tectonics could have been proceeding vigorously for long periods with no significant change in continental configuration, but as noted earlier (see Internal heating and surface heat flux), more vigorous convection (i.e. higher heat flow) in the past may not allow us to construct a reasonable thermal history. Nonetheless, the speed of continental motion may be better regarded as a minimum estimate on the global average, so the agreement with model prediction should not be taken at face value.

The interpretation of geological records prior to *c.* 3 Ga has been controversial. It is important to keep in mind that, even if plate tectonics was operating in the early Archean, it does not have to resemble modern-style plate tectonics. The most important ingredient is the subduction of oceanic lithosphere, and other aspects of plate tectonics could have been different. It follows that a mere change in the style of tectonics [e.g. at *c.* 3.2 Ga as suggested by the geology of the Pilbara Craton in Australia (Van Kranendonk *et al.*, 2007)] does not necessarily coincide with the emergence of plate tectonics. Plates can coexist with vertical tectonics, as currently observed at the Basin and Range (Sleep, 2007). In fact, the operation of plate tectonics as early as 3.7–3.8 Ga was suggested by Komiya *et al.* (1999) based on the Isua supracrustal belt in Greenland, and the age of episodic granite production recorded in the Istaq Gneiss Complex of Greenland (Nutman *et al.*, 2002) as well as the Narryer Gneiss Complex of Western Australia (Kinny and Nutman, 1996) may point to an early global collisional orogeny at *c.* 3.6 Ga (Nutman, 2006). Interestingly, this timing is consistent with what thermal evolution indicates (Fig. 3d).

The possibility of plate tectonics in even earlier times has also been widely debated, and it is centred on the interpretation of the Hadean (> 4 Ga) detrital zircons (e.g. Mojzsis *et al.*, 2001; Wilde *et al.*, 2001; Harrison *et al.*, 2005; Valley *et al.*, 2005; Coogan and Hinton, 2006; Grimes *et al.*, 2007). Some authors have suggested the existence and recycling of continental crust as early as 4.5 Ga (Harrison *et al.*, 2005), which implies the operation of plate tectonics (or just recycling of oceanic lithosphere) shortly after the formation of Earth. The interpretation of out-of-context detrital zircons is, however, not unique especially for the Hadean era (e.g. Sleep, 2007), unless we invoke Uniformitarianism for relevant geological processes. What the Hadean zircons actually constrain seems to be an open question. A related issue is that, if continental crust existed in the Hadean, how voluminous it could have been then, and it will be discussed later (see Continental growth: instantaneous, gradual, or discontinuous?).

A note on late heavy bombardment

On the basis of the lunar cratering record, it has been suggested that Earth experienced a spike of meteorite bombardment at 4.0–3.8 Ga (Hartmann *et al.*, 2000; Ryder, 2002). This event is called as the late heavy bombardment, and it is sometimes discussed as if it could have had a strong influence on the stability of lithosphere (and thus the operation of plate tectonics) (e.g. Shirey *et al.*, 2008). The total mass of impactors that hit the Moon during this period is estimated to be $\sim 3 \times 10^{18}$ kg (Ryder, 2002), and by extrapolation, Earth is expected to have been hit by a few big (>300 km diameter) impactors and numerous smaller ones (about a dozen of $\sim$200-km-diameter impactors, a hundred of $\sim$100-km-diameter impactors, and so on) (e.g. Sleep, 2007). The mass of a 300-km-diameter impactor would be $\sim 4 \times 10^{19}$ kg, and its influence of the thermal state of Earth's lithosphere may be estimated by equating its potential energy and the energy required to heat up the lithosphere by ΔT as:

$$M_L c_p \Delta T \sim \frac{G M_E m}{r_E}, \qquad (10)$$

where M_L is the mass of lithosphere ($\sim 1.7 \times 10^{23}$ kg for the thickness of 100 km), c_p is its specific heat ($\sim 10^3$ J kg^{-1} K^{-1}), G is the gravitational constant (6.67×10^{-11} m^3 kg^{-1} s^{-2}), M_E is the mass of Earth ($\sim 6 \times 10^{24}$ kg), m is the mass of the impactor, and r_E is the radius of Earth ($\sim 6.4 \times 10^6$ m). For a 300-km-diameter impactor, ΔT is only $\sim$15 K. Of course, impact heating is a highly localized phenomenon, and a much greater temperature rise is expected in the vicinity of an impact site. The late heavy bombardment is certainly important for surface conditions as it could vaporize at least a part of oceans. The above order-of-magnitude estimate is probably sufficient, however, to show that the effect of even the biggest impactor during the late bombardment would not be significant for *global* mantle dynamics such as the operation of plate tectonics.

Geodynamical considerations

Estimating the onset of plate tectonics on a purely theoretical ground is difficult because the very physics of plate-tectonic convection is not well understood (e.g. Bercovici *et al.*, 2000). There still exist, however, certain physical constraints that have to be considered when discussing the style and vigour of mantle convection in the early Earth. In particular, the interpretation of geochemical data is often assisted by or compared to geodynamical considerations (sometimes subconsciously), and this is where a preconception could potentially lead to biased interpretations. For example, vigorous convection is often assumed for the Hadean dynamics (e.g. Sleep and Zahnle, 2001), but it is not clear if such vigour can really be attained in the early Earth. Though convection in a magma ocean should have been extremely vigorous (Solomatov, 2007), the transition from such intense liquid convection to solid-state mantle convection is currently poorly understood [see Sleep (2007) for the possibility of 'mush ocean']. If mantle convection was in the mode of plate tectonics, for example, the scaling of Korenaga (2006) might be applicable, which predicts slower, not faster, plate motion. Even without calling for this particular scaling, however, a simple physical argument centred on the subductability of oceanic lithosphere (e.g. Bickle, 1986; Davies, 1992) seems to question rapid convection when Earth was much hotter than today. For the top boundary layer to be recycled into the interior by free convection, the boundary layer must be negatively buoyant, at least on average, with respect to the hot interior. This buoyancy constraint is not an issue for simple thermal convection with a homogeneous material because the top thermal boundary layer is always denser than the hot interior. Convection in Earth's mantle, however, induces pressure-release melting and differentiates the shallow mantle into basaltic crust and depleted mantle lithosphere, both of which are less dense than the mantle before melting if compared at the same temperature. For the top boundary layer to become denser than the interior, therefore, it has to be cooled for a sufficiently long time so that negative buoyancy due to thermal contraction overcomes the intrinsic chemical buoyancy [note that basalt is transformed to denser eclogite at $\sim$60 km depth (Ringwood and Green, 1964), so this chemical buoyancy issue is important only for the initiation of subduction, not for its continuous operation]. For the present-day condition, this time-scale is estimated to be *c.* 20 Myr, but when the mantle is hotter than present by 200–300 K, the time-scale could be on the order of 100 Myr (Davies, 1992). That is, when the mantle was hotter in the Hadean, it must have had thicker buoyant crust as well, which could return to the interior only after *c.* 100-Myr long surface cooling. This time-scale is comparable to the present-day time-scale to renew ocean basins. Vigorously convecting hotter mantle could be at odds with more copious melting expected for such mantle.

The subductability argument may be countered because thick oceanic crust such as oceanic plateaus does subduct at present (e.g. Sleep and Windley, 1982), but such thick crust is localized at present as opposed to its likely global occurrence in the Archean. The argument may also be questioned because the seafloor is currently subducting irrespective of its age (Parsons, 1982). Indeed, near zero-age crust is subducting under the western margin of the North American plate, whereas the oldest crust in the Atlantic (*c.* 180 Ma) is not subducting at all, so the simple buoyancy argument may seem to be irrelevant to actual plate dynamics. Young subducting seafloor is, however, attached to an older, already subducted plate, and the presence of old, non-subducting seafloor tells us that negative buoyancy alone is not sufficient to initiate subduction. Probably a more adequate measure is the global average of the age of subducting seafloor, which is *c*. 49 Ma (rate average) or *c*. 61 Ma (area average) [based on Table 1 of Parsons (1982)]. In either case, it is old enough to achieve sufficient negative buoyancy under the present-day condition.

The concept of subductability depends on two factors: thermal diffusivity and intrinsic chemical buoyancy. Whereas the former is virtually constant under plausible mantle conditions, the chemical buoyancy is a function of mantle temperature, and the functionality is less certain. To predict the density of oceanic crust, one has to first calculate the composition of primary mantle melt for a range of mantle temperature, and then calculate the density of mineral aggregates expected to be solidified from a given melt composition. The current understanding of mid-ocean ridge magmatism is probably sufficient to conduct the first step with moderate accuracy (e.g. Langmuir *et al*., 1992; Kinzler, 1997; Walter, 1998), but the second step is more uncertain because the details of crystallizing phases depend on the temperature and pressure conditions within newly forming crust and thus on how exactly crust is constructed. Even for the present-day oceanic crust, the physical mechanism of crustal accretion is still under debates (e.g. Phipps Morgan and Chen, 1993; Korenaga and Kelemen, 1998; Wilson *et al*., 2006), and we do not know, from first principles, how to construct much thicker crust corresponding to hotter mantle. The subductability calculation by Korenaga (2006), for example, depends on the crustal density parametrization of Korenaga *et al*. (2002), which assumes low-pressure crystallization. Future studies on the crustal structure of large igneous provinces may provide important field constraints on this issue.

Provided that intrinsic chemical buoyancy remains significant for hotter mantle, subductability is a robust physical constraint, but its use needs some care. Davies (1992), for example, argued that plate tectonics was unlikely when the mantle was much hotter because the time-scale for negative buoyancy would be too long (i.e. *c*. 100 Myr) to be achieved in vigorously convecting mantle, but this argument depends on the conventional heat-flow scaling (Fig. 2). Also, van Thienen *et al*. (2004) suggested that plate tectonics may not be possible when the mantle potential temperature is higher than 1500 °C based on subductability, but their calculation assumes the so-called plate model for the evolution of oceanic lithosphere (Parsons and Sclater, 1977; Stein and Stein, 1992), in which the growth of lithosphere is inhibited after *c*. 80 Myr. Though convective instability can limit the growth of lithosphere (Parsons and McKenzie, 1978), this instability is a function of mantle viscosity, and it is difficult to justify the use of a constant maximum thickness over a range of mantle temperature (cf. Korenaga, 2003). Note that the plate model contains an artificial bottom boundary condition to suppress cooling, and a recent global analysis of seafloor topography casts a doubt on the observational basis for this model (Korenaga and Korenaga, 2008).

Recently, Davies (2006) proposed that the subductability issue would not present a major obstacle for the operation of plate tectonics if the mantle was already depleted in the early Earth by earlier melting events. In his numerical model, subduction is forced by a surface velocity condition, and subducted oceanic crust segregates from depleted mantle lithosphere and sinks to the lower mantle, leading to a gradual depletion of the upper mantle, the melting of which does not yield thick oceanic crust. This scenario, however, requires some kind of tectonics (other than plate tectonics) that can subduct oceanic crust, and a time-scale to achieve the required mantle depletion is uncertain [in the model of Davies (2006), subduction was achieved by assuming the continuous operation of plate tectonics]. The once highly depleted upper mantle would also need to be refertilized later to explain the present-day upper mantle. In other words, the upper mantle needs to change its composition so that melting always yields relatively thin oceanic crust regardless of its temperature. The plausibility of this mechanism seems to hinge on a delicate balance between mantle mixing and secular cooling.

Subductability thus remains to be an important factor to be considered, and the backward integration of Earth's thermal history seems to provide an important perspective related to subductability, namely, the role of internal heating on the initiation of plate tectonics. Figure 4 shows two kinds of lithospheric thickness. One is the minimum thickness of subductable oceanic lithosphere [which is equivalent to the 'critical thickness' defined by Korenaga (2006)]; lithosphere must exceed this thickness to become negatively buoyant. This thickness is a function of mantle temperature, so its temporal variation is determined by an assumed thermal history (Fig. 3a). The other is the equilibrium thickness of a hypothetical stagnant lid, which is a function of internal heat generation as (Korenaga, 2006):

$$\bar{h}_{\mathrm{SL}} \sim \frac{kAT_{\mathrm{m}}}{H_{\mathrm{m}}}, \tag{11}$$

where k is thermal conductivity and A is the total surface area of Earth. At a thermal equilibrium, lithosphere should become thinner for higher internal heating, and Fig. 4 indicates that the equilibrium thickness may have been less than the minimum thickness before *c*. 2.5–3 Ga. In the Archean and the Hadean, the amount of internal heat generation in the mantle was greater than present by a factor of >2 (Fig. 3c), and this high internal heating may have suppressed the growth of the top thermal boundary layer and thus prevented lithosphere to become negatively buoyant. This argument is, however, probably too simplistic because the equilibrium thickness is unlikely to be achieved instantaneously. The thermal adjustment time-scale can be long, on the order of 1 Gyr (Daly, 1980), and this long time-scale in a sense justifies Eq. (1), in which surface heat flux $Q(t)$ is usually parameterized as a function of mantle temperature (Fig. 2) and can vary independently from internal heat production $H_{\mathrm{m}}(t)$. The physics of thermal adjustment is, however, not fully explored for mantle convection with realistic rheology, and the significance of the thickness crossover in Fig. 4 is an open question. As noted by Sleep (2000) and Stevenson (2003), a change in the mode of mantle convection may lead to non-monotonic thermal histories.

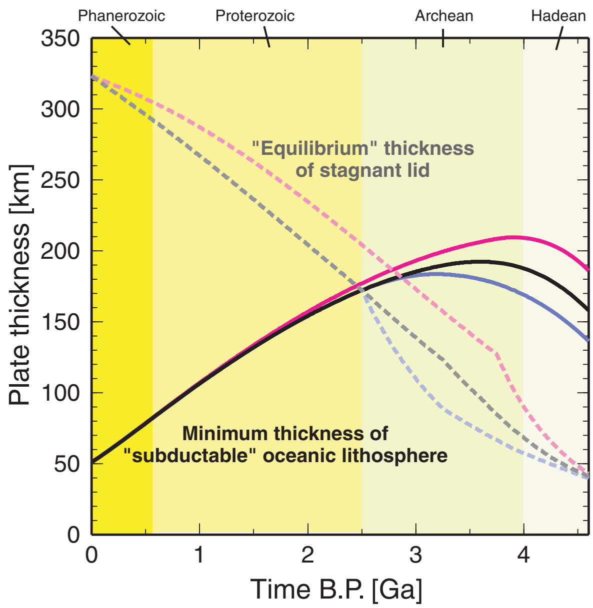

Fig. 4 The predicted variation of the minimum thickness of subductable oceanic lithosphere and the equilibrium thickness of stagnant lid, based on the thermal history shown in Fig. 3. The equilibrium lid thickness is thinner in the past due to higher mantle heat production, and is thinner than the minimum subductable thickness before the Proterozoic, suggesting that the subductability of oceanic lithosphere is a key factor for the emergence of plate tectonics in the early Earth.

Though it could be a coincidence, the timing of crossover is close to the Archean-Proterozoic boundary, at which surviving continental crust began to emerge in abundance. A quantitative relationship between mantle convection and continental recycling throughout Earth's history is one of the subjects that should be explored by future geodynamical studies.

To go beyond this simple buoyancy argument, it is imperative to advance our understanding of the physics of plate tectonics. Though plate-tectonic-like convection can be successfully simulated in numerical models (e.g. Moresi and Solomatov, 1998; Tackley, 2000; Richards *et al.*, 2001; Ogawa, 2003; Gurnis *et al.*, 2004; Stein *et al.*, 2004), currently available models treat the strength of oceanic lithosphere as a free parameter, which must be adjusted to achieve plate tectonics. It is difficult to discuss the onset of plate tectonics on the basis of those models because the free parameter may not be constant over the geological time. It would desirable to predict the strength of lithosphere from first principles based on tangible physical processes; such research effort is still in its infancy (e.g. Korenaga, 2007b).

Secular cooling and flood basalt volcanism

Did mantle plumes exist in the Archean?

Apart from metal-silicate segregation that took place within the first hundred million years of Earth's history (e.g. Halliday, 2003), chemical differentiation in Earth's interior refers to the melting of silicate rocks. The melting of shallow upper mantle usually takes one of the following three types of surface manifestation: mid-ocean ridge magmatism, arc magmatism and hotspot magmatism (e.g. Wilson, 1989; McBirney, 1993). The first two are associated with plate tectonics. Here, the term hotspot magmatism is used in a broad sense to cover not only hotspots such as Hawaii and Iceland but also continental and oceanic flood basalt provinces such as the Deccan Traps and the Ontong Java Plateau (Coffin and Eldholm, 1994). This type of magmatism is commonly explained by the upwelling of mantle plumes (e.g. Morgan, 1971; Richards *et al.*, 1989; Campbell and Griffiths, 1990; White and McKenzie, 1995) though the origin of hotspot magmatism has been controversial (e.g. Anderson, 1998; Foulger *et al.*, 2005). Thermal anomalies such as mantle plumes are certainly one way to generate hotspot islands and flood basalts, but not the only way because chemical and/or dynamical anomalies may also result in similar magmatic activities (e.g. Sleep, 1984; Tackley and Stevenson, 1993; Korenaga and Jordan, 2002; Anderson, 2005; Ito and Mahoney, 2005; Korenaga, 2005b).

For the sake of discussion, however, let us assume that most of hotspots and flood basalts are formed by the melting of mantle plumes that originate in the core–mantle boundary region. In this case, the thermal evolution of Earth suggests that hotspot magmatism should have been more reduced in the past. As will be explained shortly, this is a straightforward consequence of a geologically plausible thermal history (Sleep *et al.*, 1988), though this fact does not seem to be widely recognized. For example, a plume-dominated regime is often suggested as an alternative mode of mantle convection in the Archean (i.e. instead of the plate-tectonic regime) (e.g. Fyfe, 1978; Van Kranendonk *et al.*, 2007), and some models for continental growth call for a prominent role of flood basalts or mantle plumes in the early Earth (e.g. Abbott and Mooney, 1995; Albarède, 1998). Along with the notion of more vigorous convection, plume activities in the Archean are commonly assumed to have been similar to or higher than today. Indeed, the vigor of mantle convection and the intensity of plumes may be related through the thermal budget of Earth, but their temporal variations do not have to be positively correlated. It is possible, for example, to have a reduced plume flux while maintaining the vigor of convection, and the thermal history shown in Fig. 3 indicates that such possibility is likely. To discuss this further, we need to relax one of the assumptions behind Eq. (4) and consider the possibility of differential core cooling using Eq. (3).

Core heat flux and the possibility of superheated core

The reconstructed thermal history of Fig. 3 is based on Eq. (4), in which the mantle and the core are assumed to have cooled at the same rate. That is, the temperature contrast at the core–mantle boundary is assumed to have been constant. In this case, the core heat flux is directly related to the secular cooling of the mantle as:

$$Q_{\mathrm{CMB}}(t) = -C_c \frac{\mathrm{d}T_m(t)}{\mathrm{d}t} \qquad (12)$$

[from Eq. (2)]. The history of core heat flux with this assumption is shown as case 1 in Fig. 5. The present-day core heat flux is estimated to be ~6 TW, and more important, the past core heat flux is lower than the present-day value and is predicted to be negative before the early Archean (i.e. core heating instead of cooling should take place then). With the nearly constant surface heat flux but with gradually decaying radiogenic heat source (Fig. 3c), the secular cooling of Earth and thus the core heat flux are likely to have been lower in the past.

The details of the predicted core evolution are, however, subject to large uncertainties. The temperature contrast ΔT_{CMB}, which is assumed to be constant, is almost a free parameter given our limited knowledge of the dynamics of the core–mantle boundary region; the rheology of the lowermost mantle is currently unknown. To explore the significance of time-dependent ΔT_{CMB}, two different variations are considered (cases 2 and 3 in Fig. 5b). In case 2, the contrast decreases linearly by 500 K over the entire Earth history, whereas in case 3 it decreases quadratically by a similar amount. Thermal evolution was solved again, but using Eq. (3), and the corresponding core heat flux was calculated as:

$$Q_{\mathrm{CMB}}(t) = -C_c \frac{\mathrm{d}T_m(t)}{\mathrm{d}t} - C_c \frac{\mathrm{d}\Delta T_{\mathrm{CMB}}(t)}{\mathrm{d}t}. \qquad (13)$$

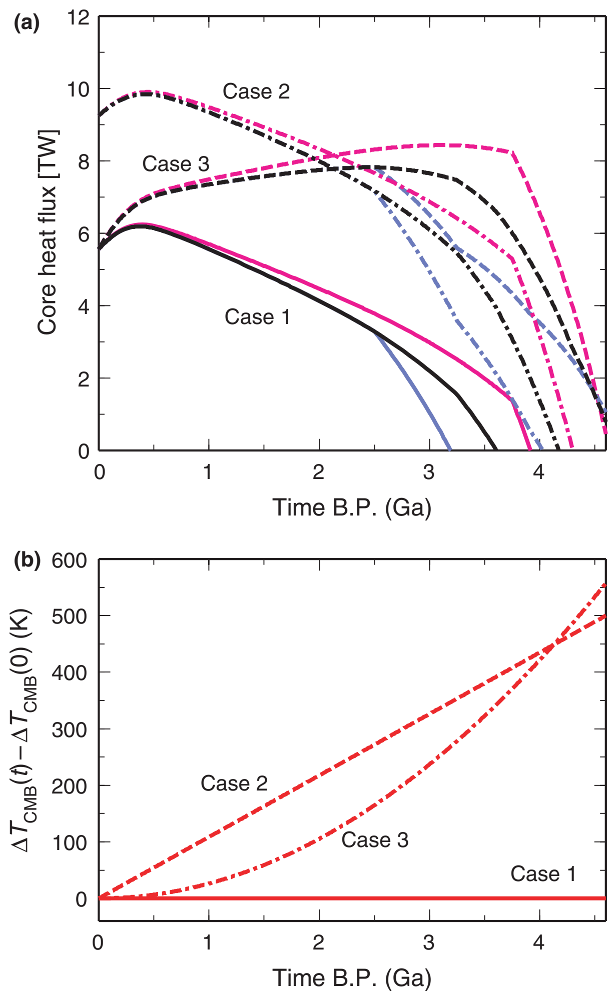

Fig. 5 (a) Predicted history of core heat flux Q_{CMB} incorporating the effect of differential core cooling. Different colours correspond to different models of continental growth (Fig. 3b). (b) Three cases of differential cooling are shown: 1, no differential cooling (the core cools at the same rate as the mantle), 2, constant rate of differential cooling (temperature contrast at the core–mantle boundary linearly decrease by 500 K over 4.6 Gyr), and 3, decreasing rate of differential cooling (heat flux due to differential cooling decreases linearly from 10 TW at 4.6 Ga to zero at present).

It may be seen that differential core cooling could modify substantially predicted core heat flux (Fig. 5a). On the other hand, the thermal history of the mantle (not shown) is not affected much by this modification. This is expected because the differential core cooling of case 2, for example, provides additional core heat flux of ~4 TW, which is significant for the core thermal budget, but not for the mantle thermal budget. Cases 1 through 3 are all consistent with the observational constraints on the present-day core heat flux [6–12 TW, (Buffett, 2003)].

At the core–mantle boundary, the core side is considered to be hotter than the mantle side by at least ~1000 K at present (Williams, 1998). The present-day contrast is uncertain because the estimate of the core-side temperature is based on the phase diagram of the core, which depends on its chemical composition and in particular on the (uncertain) light element composition (e.g. McDonough, 2003). Nevertheless, a temperature contrast more than a few hundred K is probably robust, and this suggests either (1) that there was no contrast at the beginning of the Earth history, but the mantle cooled faster than the core, or (2) that the core was initially hotter than the mantle, and the temperature contrast has not been entirely removed. The first scenario seems unlikely because it would predict core heat flux lower than case 1 and thus becomes incompatible with the present-day core heat flux estimate and also with the history of the geomagnetic field (e.g. McElhinny and Senanayake, 1980; Buffett, 2002;

Tarduno *et al.*, 2007). The second possibility is physically plausible because either the gravitational segregation of the core or the Moon-forming giant impact is expected to deposit a considerable amount of heat into the core (Solomatov, 2007). The initial temperature of this superheated core and its later evolution are uncertain, but it would be fortuitous if the initial temperature contrast has been maintained to the present day; it is possible, but it would require a specific dynamics of the core-boundary boundary region. Figure 5(a) suggests that the possibility of differential core cooling (or the current uncertainty of the core–mantle boundary dynamics) provides important degrees of freedom for the coupled core–mantle evolution. Explaining both the present-day core heat flux and the history of the geomagnetic field has been a conundrum for the thermal history of Earth's core (e.g. Labrosse *et al.*, 2001; Buffett, 2003; Nimmo *et al.*, 2004; Butler *et al.*, 2005), and these extra degrees of freedom may help to resolve it.

A linear decrease in the temperature contrast (case 2) shifts core heat flux almost uniformly, thus unaffecting the trend of lower heat flux in the past. To reverse this trend, we need to invoke a greater temporal variation in the past (e.g. case 3), but there is a negative feedback mechanism to suppress core heat flux in the early Earth. Higher differential core cooling is equivalent to higher internal heating in the mantle [Eq. (3)], which would then reduce the secular cooling of the entire Earth (i.e. including the core). This is why even case 3 predicts vanishing core heat flux in the early Earth (Fig. 5a). It appears that core heat flux was probably lower than (cases 1 and 2) or similar to (case 3) the present-day value, and substantially higher core heat flux in the early Earth probably requires an unrealistic degree of differential core cooling.

Origins of flood basalts and preservation bias

With reduced core heat flux, it would be difficult to expect more vigorous plume activities in the early Earth. Furthermore, plume heat flux is likely to be smaller than the total core heat flux (e.g. Davies, 1993; Labrosse, 2002). If the average plume size remains similar through time, then, the number of plumes should have been lower, and if we instead assume the constant frequency of plume formation, the average plume size should have been smaller. In either case, we expect a reduced volume of flood basalt magmatism in the past as a corollary of the mantle plume hypothesis. This does not necessarily mean that we cannot expect greater flood basalt magmatism in the past. First of all, 'normal' mantle in the Archean is likely to be $\sim$300 K hotter than present (Fig. 3a), so flood basalt magmatism in the Archean may not necessarily require an unusual source mantle. High temperature alone, however, is probably insufficient to explain focused magmatic events such as flood basalts, and there are a few non-plume mechanisms as discussed below.

One popular concept is the episodic overturn of layered-mantle convection (e.g. Stein and Hofmann, 1994; Condie, 1998; Rino *et al.*, 2004). In the layered-mantle convection mode, the lower mantle cools less efficiently and thus becomes hotter than the upper mantle. Numerical modelling in the early 1990s suggested that, whereas it is difficult to maintain a purely layered state, episodically layered convection may take place with the endothermic phase change at the base of the mantle transition zone (i.e. at the 660-km discontinuity) (e.g. Machetel and Weber, 1991; Honda *et al.*, 1993; Tackley *et al.*, 1993; Solheim and Peltier, 1994). When a temporally layered state is broken, a large volume of the hot lower mantle material can be brought to the surface, potentially resulting in massive melting events. Note that the plausibility of episodic overturns depends critically on the magnitude of the (negative) Clapeyron slope for the endothermic phase change. To insulate the lower mantle from cooling due to subducted slabs and make it substantially hotter than the upper mantle, subducted slabs must be supported *globally* by the phase change for several hundred million years. Recent experimental studies suggest that the Clapeyron slope is not as strongly negative as previously thought (Katsura *et al.*, 2003; Fei *et al.*, 2004), potentially undermining the physical basis for episodically layered convection.

Another mechanism that may produce large igneous provinces is the upwelling of chemically anomalous mantle that has been fertilized by recycled oceanic or continental crust (e.g. Korenaga and Kelemen, 2000; Yaxley, 2000; Anderson, 2005). More fertile mantle is usually intrinsically denser (O'Hara, 1975), so its upwelling probably requires special tectonic environments (Korenaga, 2004, 2005b). Compensating chemical density anomalies by thermal buoyancy is possible, but it would require unrealistically hotter mantle [e.g. ΔT $\sim$500–600 K (Lin and van Keken, 2005)], which may not be consistent with available petrological constraints ($\Delta T \sim$ 100–300 K) (e.g. White and McKenzie, 1995; Herzberg, 2004). Setting aside this dynamical difficulty, it is also unclear how abundant such fertile mantle would have been in the past. For one thing, the recycling rate of oceanic crust is controlled by plate motion, which may not have been different from present as discussed earlier (Internal heating and surface heat flux). On the other hand, the recycling of continental crust is free from this constraint, and primordial heterogeneities created during the magma ocean may have been abundant in the early Earth. The possibility of fertile mantle is a wild card, as our understanding of the dynamics of chemically heterogeneous mantle is still immature (Korenaga, 2008b).

Geological indicators for flood basalts are common in Archean terranes (e.g. Campbell *et al.*, 1989; Ernst and Buchan, 2003; Sandiford *et al.*, 2004), which is probably the basis for the notion of more active plume activities. One way to reconcile the apparent discrepancy between the theoretical expectation and the field observation is to call for preservation bias; the continental crust with flood basalts may have better survived for the following reason. The genesis of flood basalts, produced by either thermal or chemical anomalies, involves the melting of a large volume of the mantle. Because mantle melting also dehydrates and stiffens the residual mantle (Karato, 1986; Hirth and Kohlstedt, 1996), this large-scale melting would produce a voluminous, stiff mantle root, which could protect the overlying continental crust from tectonic disturbances. It is unclear how much of such dehydrated residual mantle has contributed to what we call today as continental tectosphere (Jordan, 1988; Pearson, 1999), but the role of dehydrated mantle in ancient continental dynamics is an important dynamical problem to consider (e.g. Doin *et al.*, 1997; Lenardic and Moresi, 1999). Oceanic crust older than *c.* 180 Myr old is all subducted, and we have only continental crust to discuss anything older. It is natural to hope for a minimal preservation bias, but when interpreting billion-years-old continental crust, it would be hard to overestimate preservation bias.

Continental growth and the history of ocean volume

Continental growth: instantaneous, gradual, or discontinuous?

When the continental crust emerged in the Earth history and how it has evolved to its preset figure have been debated over several decades (e.g. Hurley and Rand, 1969; O'Nions *et al.*, 1979; DePaolo, 1980; Armstrong, 1981; Taylor and McLennan, 1985; Jacobsen, 1988; Collerson and Kamber, 1999; Campbell, 2003; Harrison *et al.*, 2005), and continental growth is still a highly controversial topic. Being highly enriched in heat-producing elements, its growth history can influence the thermal evolution of Earth by depleting the convecting mantle [Eq. (9)], but this type of influence is of relatively minor importance (Fig. 3). Probably a more important aspect is whether the production of continental crust has been continuous or discontinuous. Gradual growth models (e.g. McLennan and Taylor, 1982; Jacobsen, 1988; Campbell, 2003) are obviously continuous, and instantaneous growth models (e.g. Armstrong, 1981; Harrison *et al.*, 2005) are also mostly continuous in this sense, because the constant continental mass is assumed to have been maintained by balancing continuous production and destruction. These 'continuous' growth models are compatible with the continuous operation of plate tectonics. 'Discontinuous' growth models (e.g. Rino *et al.*, 2004; Hawkesworth and Kemp, 2006; Parman, 2007), on the other hand, suggest that the mode of mantle convection may have changed at least a few times in the past, and the use of single heat-flow scaling in thermal evolution modelling, as done in most of previous studies, may become overly simplistic in this case. Thus, the debate over continental growth has a critical relevance to the theoretical formulation of thermal evolution modelling.

The episodic overturn of layered-mantle convection has been a popular concept as a plausible mechanism that may explain the episodic growth of continental growth (e.g. Stein and Hofmann, 1994), but as discussed earlier, this idea is based on early numerical convection models with a strong endothermic phase transition, the assumption of which does not seem to be valid in light of recent experimental studies. Recently, O'Neill *et al.* (2007) proposed that plate tectonics itself might have been intermittent in the Precambrian. Whereas their argument using palaeomagnetic data is weak given the likelihood of true power wander (Evans, 2003), they offer a plausible dynamical reasoning. Mantle convection models with pseudo-plastic rheology are known to exhibit three modes of convection (stagnant-lid, intermittent plate tectonics, and continuous plate tectonics), depending on the assumed strength of lithosphere (or its maximum yield strength) (e.g. Moresi and Solomatov, 1998; Stein *et al.*, 2004). When the mantle was hotter in the past, its internal viscosity may be lower due to temperature dependency, so convective stress could become too low to sustain the continuous operation of plate tectonics. In this case, intermittent plate tectonics is possible for a certain range of maximum yield stress. Unlike the model proposed by Silver and Behn (2008), however, this original version of intermittent plate tectonics proposed by Moresi and Solomatov (1998) hardly modifies conventional heat-flow scaling, because low heat flux during the stagnant-lid mode is compensated by high heat flux during the plate tectonics mode [see, for example, Figure 3 of Moresi and Solomatov (1998)]. So it may be able to explain continental growth but not thermal evolution. Also, as noted earlier, the pseudo-plastic model of Moresi and Solomatov (1998) needs to assume the strength of lithosphere (i.e. its maximum yield stress), which may not be constant over Earth's history. Equally important, mantle viscosity is a function of not only temperature, but also other parameters such as grain size and water content (e.g. Karato and Wu, 1993; Hirth and Kohlstedt, 2003; Korenaga and Karato, 2008). Solomatov (1996), for example, suggested that hotter mantle could become more viscous if grain-size-dependent viscosity is considered. Assuming weaker convective stress from a hotter mantle is equivalent to holding these other variables constant through time, which may not be warranted. In fact, as shown in the next section, the present-day thermal budget indicates that the mantle was probably drier in the past, which may compensate a decrease in viscosity due to temperature dependency.

The diversity of continental growth models is partly because different geochemists place different weights on relevant geochemical data such as $^{143}Nd/^{144}Nd$ and Nb/Th. For example, a significant volume of continental crust in the Hadean, as implied by the hafnium isotope data of ancient zircons (Harrison *et al.*, 2005), seems to be incompatible with the Th–U–Nb systematics of depleted-mantle-derived rocks (Collerson and Kamber, 1999). The interpretation of geochemical data in terms of geological processes, however, often depends on simple box models (e.g. DePaolo, 1980; Jacobsen, 1988), which in turn assumes rapid mantle mixing (Coltice *et al.*, 2000). Also, geochemical interpretations usually require the chemical or isotopic composition of the BSE as a reference baseline, but the uncertainty of such reference value is not trivial. Continental growth models based on the Nd isotope evolution (e.g. Bennett, 2003), for example, assumes that the Sm/Nd ratio of the BSE is identical to that of chondrites within 1% uncertainty [the 'strong' version of chondrite assumption (Korenaga, 2008b)]. Available isotopic data from terrestrial samples, on the other hand, require only the 'weak' version of chondrite assumption, i.e. the ratios of refractory lithophile elements such as Sm and Nd should not be different from the chondritic average more than a few per cents (Lyubetskaya and Korenaga, 2007b). This type of uncertainty may be essential when constructing a geochemical model that can reconcile different geochemical data simultaneously.

Global water cycle and net water influx

Frequent inundations throughout the Phanerozoic suggest that the mean sea level has always been close to the mean continental level, which is known as the constant freeboard (Wise, 1974). This constancy is at least geologically reasonable; continental crust is subject to erosion when it is above the sea level, and when a part of continents subsides below the sea level, it would likely be the locus of deposition. The abundant occurrence of submarine flood basalt magmatism in the Archean and Proterozoic (Arndt, 1999), however, implies that the mean sea level had been high enough to inundate a substantial fraction of continents through the Precambrian. The constant freeboard is thus probably too simplified an assumption, but the following point seems to be robust: there has always been a sufficient volume of water to fill up the ocean basins, *at least* to the mean continental level. In other words, the constant freeboard is still useful to quantify the lower bound on the ocean volume.

Reymer and Schubert (1984) coupled the constant freeboard with the thermal evolution of Earth and estimated the history of continental growth assuming that the ocean volume has been constant (Fig. 6a). Their model is based on the conventional heat-flow scaling (Fig. 2), which predicts faster plate motion (i.e. higher heat flow) for hotter mantle in the past. Faster plate motion means younger and thus shallower seafloor, and to maintain the constant freeboard without changing the ocean volume, continental mass needs to be reduced. Note that the conventional scaling cannot reconstruct a reasonable thermal history unless internal heat production in the convecting mantle is much higher than geochemical constraints (see Internal heating and surface heat flux). This important fact tends to be overlooked, and the notion of more vigorous convection in the past has been entrenched in the literature on the continental freeboard or the history of ocean volume (e.g. Galer, 1991; Kasting and Holm, 1992; Galer and Mezger, 1998; Harrison, 1999; Hynes, 2001; Rüpke *et al.*, 2004; Kasting *et al.*, 2006).

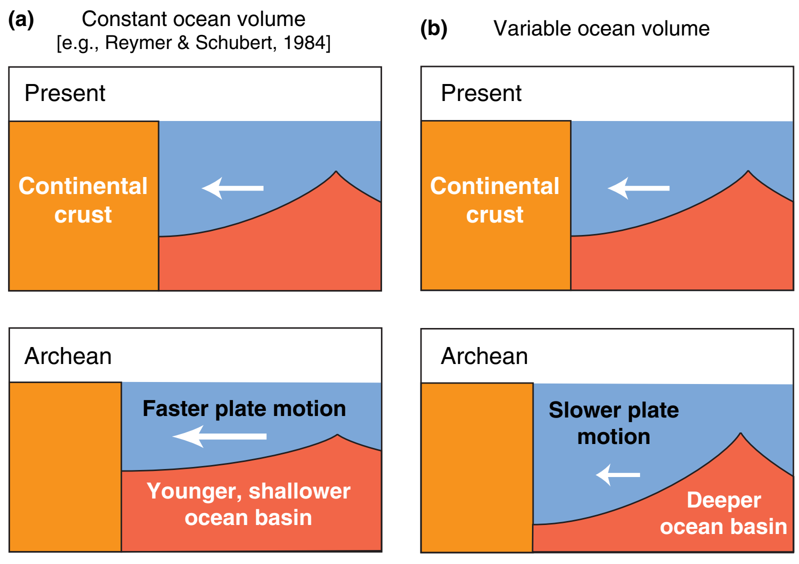

Fig. 6 Cartoon illustrating possible relations among heat-flow scaling, continental growth, and ocean volume, under the assumption of the constant freeboard. (a) When the ocean volume is assumed to be constant, plate motion must have been faster to create younger and shallower seafloor when there was less continental crust. Faster plate motion (i.e. higher heat flux) in the past, however, results in thermal catastrophe. (b) Slower plate motion, which is more consistent with the present-day thermal budget, predicts a greater ocean volume in the past.

The present-day thermal budget characterized by a low Urey ratio instead suggests less vigorous convection thus slower plate motion, the corollary of which is older and deeper ocean basins in the past (Fig. 6b). In this case, the volume of Earth's oceans may have been greater to maintain the constant freeboard, as many geochemical studies suggest smaller continental mass in the Archean. This interesting possibility has not been seriously considered probably because it is difficult to test. It can be demonstrated, however, that the constant ocean volume is inconsistent with the present-day thermal budget, without using any heat-flow scaling law.

A key concept is that, for a given continental growth history, the assumptions of constant freeboard and constant ocean volume are sufficient to determine oceanic heat flux, which represents a major fraction of convective heat flux. In general, oceanic heat flux at a time t, $Q_o(t)$, may be expressed as:

$$Q_o(t) = \int_0^{\tau_{max}} \frac{dA_o}{d\tau}(\tau, t) q(\tau, t) d\tau, \quad (14)$$

where $dA_o/dt(\tau)$ is the area–age distribution of seafloor, and $q(\tau)$ is heat flow from a seafloor of age τ. Both of them can also be a function of a geological time t. The present-day area–age distribution is best approximated by a triangular distribution (i.e. the area of younger seafloor is greater than that of older seafloor), and we assume this distribution for older times because it arises naturally from subduction irrespective of seafloor age (Parsons, 1982). Thus we have

$$\frac{dA_o}{d\tau}(\tau, t) = G(t)\left(1 - \frac{\tau}{\tau_{max}(t)}\right), \quad (15)$$

where $\tau_{max}(t)$ is the age of the oldest seafloor. The coefficient $G(t)$ is constrained by the total area of ocean basins, $A_o(t)$, as:

$$A_o(t) = A - A_c(t) = \frac{G(t)}{2}\tau_{max}(t), \quad (16)$$

where A is the total area of Earth's surface and $A_c(t)$ is the total area of continents. Seafloor heat flow is assumed to follow half-space cooling as:

$$q(\tau, t) = \frac{B}{\sqrt{\tau}} \frac{T_m(t)}{T_m(0)}, \quad (17)$$

where B is 550 mW m^{-2} Myr$^{-1/2}$ (Korenaga and Korenaga, 2008). At present (i.e. $t = 0$), $G(0) = 3.45$

$km^2\ yr^{-1}$ and $\tau_{max}(0) = 180$ Myr (Parsons, 1982), and Eq. (14) gives ~34 TW for the total oceanic heat flux, which is in good agreement with the actual estimate [32 ± 2 TW (Pollack *et al.*, 1993; Jaupart *et al.*, 2007)]. Similarly, assuming the constant freeboard, the total ocean volume may be expressed as:

$$V(t) = \int_0^{\tau_{max}} \frac{dA_o}{d\tau} d(\tau, t) d\tau, \qquad (18)$$

where $d(\tau)$ is the seafloor depth as a function of seafloor age,

$$d(\tau, t) = d_0(t) + b(t)\sqrt{\tau}, \qquad (19)$$

Here d_0 is the zero-age depth (average depth to mid-ocean ridge axis), and b is the subsidence rate due to half-space cooling. At the present, we have $d_0(0)$ = 2654 m and $b(0)$ = 323 m $Myr^{-1/2}$ (Korenaga and Korenaga, 2008). The subsidence rate is linearly proportional to a temperature contrast between the surface and the interior (e.g. Turcotte and Schubert, 1982), so we have

$$b(t) = b(0)\frac{T_m(t)}{T_m(0)}. \qquad (20)$$

By assuming the constant ocean volume, therefore, we can solve Eq. (18) for the maximum seafloor age as:

$$\tau_{max}(t) = \left[\frac{15}{8b(t)}\left(\frac{V(0)}{A_o(t)} - d_0(t)\right)\right]^2, \qquad (21)$$

and $Q(t)$ can then be calculated by combining Eqs (15)–(17). The total area of continents in Eq. (16) is calculated from a given history of continental growth, assuming that the average thickness of continental crust has been approximately constant at ~35–45 km (Durrheim and Mooney, 1991; Galer and Mezger, 1998). Note that Galer and Mezger (1998) suggested that continental crust could have been thicker in the Archean than at present by ~5 km, based on original burial pressures estimated for exposed Archean granite-greenstone segments. Thicker crust in the past would only substantiate the following argument because it means more ocean basins [Eq. (16)] and flatter seafloor topography to satisfy the constant freeboard.

We used the oceanic heat flux of Eq. (14) as the lower bound on the total convective heat flux (i.e. subcontinental heat flux is neglected), and solved the heat balance of Eq. (4) starting with the present-day Urey ratio of 0.22. Here the zero-age seafloor depth d_0 is assumed to be constant. This assumption may be justified because the existence of volcanogenic massive sulphide deposits since the Archean (e.g. Nisbet *et al.*, 1987) indicates that the depth of mid-ocean ridges should always be greater than ~2 km below sea level (Ohmoto, 1996). As shown in Fig. 7a, this calculation leads to thermal catastrophe even for the instantaneous growth model, in which no net continental growth takes place during the last 4 Gyr. This is because hotter mantle by itself results in greater thermal subsidence [Eq. (20)], so to keep the same ocean volume, the maximum seafloor age should decrease (Fig. 7c), i.e. seafloor should become younger on average, resulting in an increase in oceanic heat flux (Fig. 7b). This positive correlation between mantle temperature and oceanic heat flux may be summarized as 'empirical' heat-flow scaling (Fig. 7d). Alternatively, we may fix the maximum seafloor age as 180-Myr-old, and solve Eq. (18) for the zero-age depth. Results for this case are shown in Fig. 8. The temperature of the Archean mantle is still too high, and the predicted zero-age depth in the Archean is probably too shallow (Ohmoto, 1996; Kitajima *et al.*, 2001).

The constant ocean volume, therefore, seems to be incompatible with the thermal budget of Earth, and the possibility of time-dependent ocean volume deserves some attention. As a preliminary attempt, the thermal history of Fig. 3 may be used to calculate the history of ocean volume (Fig. 9). First, noting that the coefficient $G(t)$ in the area–age distribution can be expressed as:

$$G(t) = Lv(t), \qquad (22)$$

where L is the total length of divergent plate boundaries and $v(t)$ is the average spreading rate, the maximum age

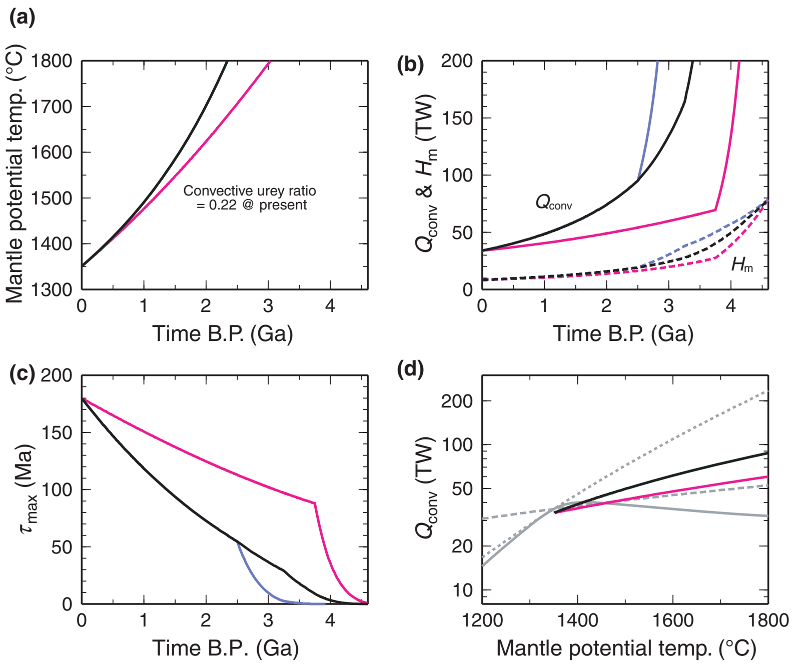

Fig. 7 A possible consequence of the constant ocean volume, with the assumption of the constant zero-age depth. The modelling procedure is identical to that used for Fig. 3, expect that convective heat flux is based on Eq. (14). (a) Mantle potential temperature as a function of time. (b) Convective heat flux and internal heating. (c) Maximum age of seafloor. (d) *A posteriori* relationship between convective heat flux and mantle temperature. For comparison, three heat-flow scaling laws of Fig. 2 are shown in grey. Legend is the same as Fig. 3; note that the cases of McLennan and Taylor (1982) (blue) and Campbell (2003) (black) are identical in (a) and (d).

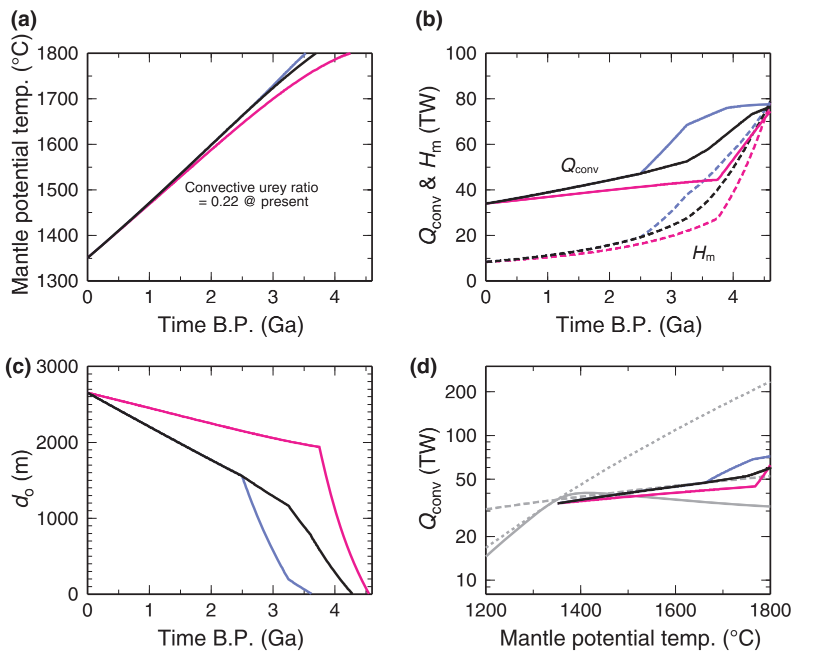

Fig. 8 Same as Fig. 7, but with the assumption of the constant maximum age of seafloor. The history of zero-age depth is shown in (c).

of seafloor may be calculated from the predicted plate velocity as:

$$\tau_{\max}(t) = \frac{2A_{\mathrm{o}}(t)}{G(0)}\frac{v(0)}{v(t)}. \qquad (23)$$

Here, L is assumed to be constant because the average size of plates is unlikely to change substantially with time (Korenaga, 2006). The total ocean volume can then be calculated from Eq. (18), assuming the constant freeboard as well as the constant zero-age depth (Fig. 9b). Results indicate that Archean ocean basins may have been deep enough to hold twice as much water as today.

Obviously, the predicted history of ocean volume should not be taken at face value, as it is based on a number of assumptions with varying credibility (Table 1). At the same time, however, losing one-ocean-worth water since the Archean is not surprising, or may even be expected in light of the global water cycle. It has long been known that water flux into the mantle from the hydrosphere by subduction exceeds water flux out of the mantle by mid-ocean magmatism by an order of magnitude (e.g. Ito *et al.*, 1983; Jarrard, 2003). Most of such water influx is usually believed to quickly return to the surface through arc magmatism, so the *net* water influx may close to be zero. It is difficult, however, to completely dehydrate subducting slabs because nominally anhydrous minerals can hold a non-trivial amount of water (e.g. Hirschmann *et al.*, 2005). Continuous subduction at the current rate could easily bring one-ocean-worth water into the deep mantle over a few billion years (Smyth and Jacobsen, 2006). The average rate of the predicted influx for the last 2 Gyr is also compatible with the estimated range of present-day influx (Jarrard, 2003) (Fig. 9c).

The continental crust contains ~1 wt% water on average (e.g. Wedepohl, 1995), equivalent to ~0.18 ocean. The MORB-source mantle is estimated to have 142 ± 85 p.p.m. water (Saal *et al.*, 2002), and the global mass balance of silicate reservoirs, on the basis of the new composition model of Earth's primitive mantle (Lyubetskaya and Korenaga, 2007a), indicates that the MORB-source mantle is representative for the bulk of the mantle (Lyubetskaya and Korenaga, 2007b), so the mantle may hold ~0.42 ± 0.25 ocean of water. Thus, the amount of water currently stored in the BSE (crust and mantle), ~0.5–1 ocean, is similar to what is estimated to have been lost from Earth's oceans (Fig. 9b). This coincidence may suggest that the mantle was dry in the Hadean and has been progressively hydrated by subduction. As mentioned earlier, such gradual hydration may be important for the continuous operation of plate tectonics. Note that the dry Archean mantle does not preclude the 'wet' origin of komatiite (Allegre, 1982; Parman *et al.*, 1997; Grove and Parman, 2004), because arc environments can be locally hydrated by subduction. The regassing-dominated global water cycle suggests that the most of terrestrial water may have originated at Earth's surface, pointing towards the impact origin of Earth's oceans. As discussed next, this issue has an intriguing connection to the thermal budget of Earth.

The origin of terrestrial water

The origin of Earth's oceans, or how this planet acquired its present amount of water in the early solar system, depends on a number of factors that are currently not known precisely enough, such as the relative timing of Earth accretion, core segregation, the dissipation of the solar nebula, and the disappearance of the massive asteroid belt (e.g. Abe *et al.*, 2000; Marty and Yokochi, 2006). The hydrogen isotope ratio (D/H) is ~150×10^{-6} for Earth's oceans, ~25×10^{-6} for the solar nebula, ~310×10^{-6} for comets and ~130–180×10^{-6} for chondrites, and with this information alone, Earth's oceans could have been derived simply from the accretion of chondritic materials, or from the solar nebula with isotopic fractionation, or from the mixing of multiple sources (e.g. Dauphas *et al.*, 2000; Drake, 2005; Genda and Ikoma, 2008). When combined with carbon and nitrogen isotope constraints, however, the late accretion (i.e. accretion after the core formation) of chondritic materials appear to be the most likely source of terrestrial water (Marty and Yokochi, 2006), and indeed, such late accretion with 0.5–1% Earth's mass is suggested by the abundance of highly siderophile elements (HSE) in Earth's mantle (e.g. Morgan, 1986) and is also physically plausible according to the dynamical simulations of planetary accretion (e.g. Morbidelli *et al.*,

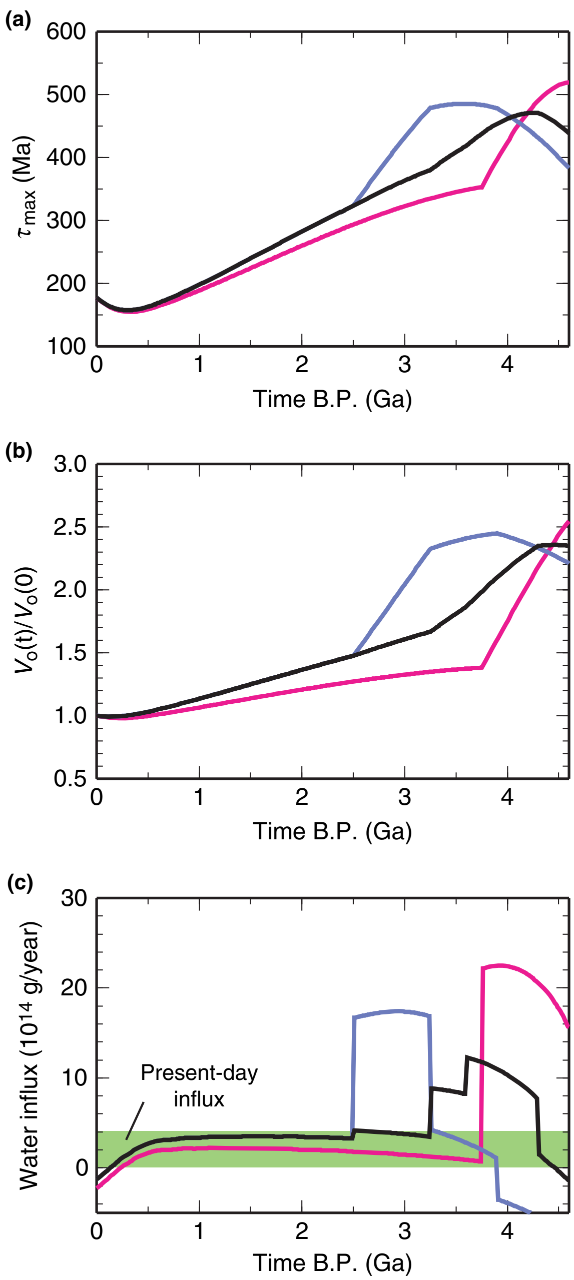

Fig. 9 (a) The maximum age of seafloor predicted from the thermal history of Fig. 3 [Eq. (23)]. (b) The predicted history of ocean volume normalized by the present-day volume. (c) Net water influx from the hydrosphere to the mantle, assuming that the predicted change of ocean volume shown in (b) is due to the hydration of the mantle by subduction. Legend is the same as in Fig. 3.

2000). Also, the Moon-forming giant impact could have been this late veneer if a fraction of the impactor's core were mechanically mixed with the Earth's mantle.

Osmium is one of those HSEs, however, and its isotopic ratio, $^{187}Os/^{188}Os$, of Earth's mantle is known to present a serious impediment to this late accretion hypothesis for Earth's oceans (Meisel *et al.*, 1996, 2001). Mantle samples such as mantle xenoliths and massif peridotites exhibit a linear correlation between this isotopic ratio and the index of melt depletion such as Al_2O_3 (Fig. 10), and the isotopic ratio for the BSE may be estimated based on the elemental composition model of BSE. With the conventional estimate for the BSE Al_2O_3 content (~4.2 wt%), the observed correlation suggests the BSE isotope ratio of 0.1289–0.1304 (95% confidence limit) (Meisel *et al.*, 2001). On the other hand, the so far observed range of the $^{187}Os/^{188}Os$ ratio is 0.1255–0.1270 for carbonaceous chondrites, 0.1270–0.1305 for ordinary chondrites and 0.1270–0.1290 for enstatite chondrites (Meisel *et al.*, 1996). Thus, Earth's $^{187}Os/^{188}Os$ is consistent with ordinary or enstatite chondrites, but not with carbonaceous chondrites. Ordinary and enstatite chondrites are, however, much drier (< 1% H_2O) than carbonaceous chondrites (~10% H_2O) (Robert, 2003), so the addition of 0.5–1% Earth's mass could account for only a small fraction of Earth's water budget. The osmium constraint has motivated a variety of more elaborate ways to deliver water to Earth and satisfy geochemical constraints at the same time (e.g. Dauphas and Marty, 2002; Drake and Righter, 2002; Marty and Yokochi, 2006).

This argument may not be so robust because, if the late accretion of HSE were made by the Moon-forming impactor (or similarly large impactors) and partial core addition, there would be no simple relationship between the added masses of HSE and water. Even if we limit ourselves to simple end-member mixing, however, the osmium constraint has one weakness that has been overlooked. The composition model of BSE is based primarily on noisy geochemical trends exhibited by mantle rocks, but the model uncertainty has not been well quantified. A new statistical method was recently built to address this issue, resulting in not only quantifying the uncertainty but also revising the model itself in a non-trivial manner (Lyubetskaya and Korenaga, 2007a). The new BSE model suggests the Al_2O_3 content of 3.52 ± 0.60 wt%. Revisiting the correlation, we would obtain the new 95% confidence limit of 0.1267–0.1277 for BSE $^{187}Os/^{188}Os$, which turns out to be consistent with *any* kind of chondrites (Fig. 10). Note that this confidence limit is derived from the uncertainty of the linear trend and does not reflect the uncertainty of the BSE model. With the

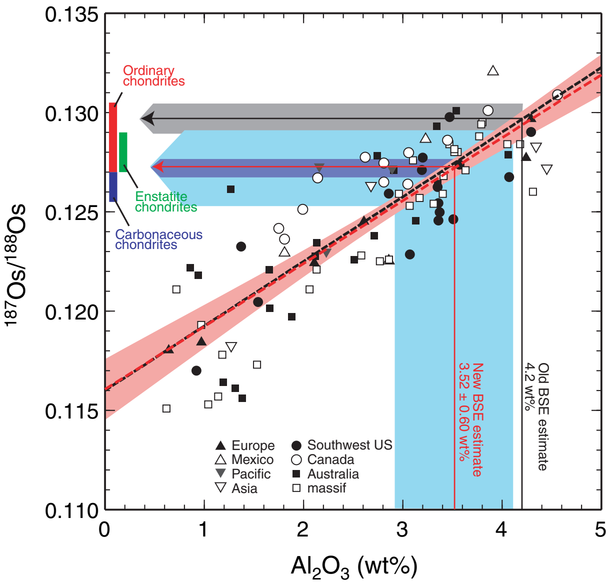

Fig. 10 Covariation of Al_2O_3 and $^{187}Os/^{188}Os$ for mantle xenoliths and massif peridotites [data are from Meisel *et al.* (2001) and references therein]. Different symbols denote different localities as indicated by the legend. Linear regression by Meisel *et al.* (2001) is shown by solid dash line ($y = 0.1160 + 0.003253x$). Linear regression by the bootstrap resampling method is shown by red dashed line ($y = 0.1161 + 0.003170x$) with pink shade for the 95% confidence limit. The small difference from the original regression probably reflects that unpublished Mexico samples used by Meisel *et al.* (2001) are not used here. The 95% confidence limit on the osmium isotope rate is shown for the case with the old BSE Al_2O_3 content (4.2 wt%) and the original regression (grey arrow) and for that with the new BSE Al_2O_3 content (3.52 wt%) and the bootstrap regression (blue arrow). The influence of the uncertainty of the new BSE model (± 0.60 wt%) is indicated by light blue arrow. The osmium isotope ranges observed for three major types of chondrites are also indicated (blue, carbonaceous; red, ordinary; green, enstatite).

uncertainty of ± 0.60 wt%, the likely range of the BSE $^{187}Os/^{188}Os$ would be widened as 0.1253–0.1291. The other potential water sources can still contribute to some degree, but the most dominant source may well be carbonaceous chondrites.

It is noted that efforts to quantify the reliability of the BSE model were motivated solely by a need to build a self-consistent geochemical model, which is important for long-standing debates over the structure and evolution of Earth's mantle (Lyubetskaya and Korenaga, 2007b; Korenaga, 2008b). The relevance to the late accretion hypothesis is entirely serendipitous. Though the new BSE model is not a final answer and should be revised in future when additional data become available, it does seem to bring us towards simpler, less arbitrary hypotheses for the structure of mantle convection as well as the origin of Earth's oceans.

Acknowledgements

The author thanks Alfred Kröner for invitation to write a review paper on the thermal evolution of Earth, and Kent Condie, Norm Sleep, Geoff Davies, and an anonymous reviewer for constructive reviews. This work was sponsored by the U.S. National Science Foundation under grant EAR-0449517.

References

Abbott, D. and Mooney, W., 1995. The structure and geochemical evolution of the continental crust: Support for the oceanic plateau model of continental growth. *Rev. Geophys. Suppl.*, **33**, 231–242.

Abe, Y., Ohtani, E., Okuchi, T., Righter, K. and Drake, M., 2000. Water in the early earth. In: *The Origin of Earth and Moon* (K. Righter and R.M. Canup, eds), pp. 413–433. University of Arizona, Tucson.

Albarède, F., 1998. The growth of continental crust. *Tectonophysics*, **296**, 1–14.

Allegre, C.J., 1982. Genesis of Archaean komatiites in a wet ultramafic subducted plate. In: *Komatiites* (N.T. Arndt and E.G. Nisbet, eds), pp. 495–500. George Allen & Unwin, London.

Allegre, C.J., Hofmann, A. and O'Nions, K., 1996. The argon constraints on mantle structure. *Geophys. Res. Lett.*, **23**, 3555–3557.

Anderson, D.L., 1998. *The EDGES of the Mantle, in the Core-Mantle Boundary Region*, pp. 255–271. American Geophysical Union, Washington, DC.

Anderson, D.L., 2005. Large igneous provinces, delamination, and fertile mantle. *Elements*, **1**, 271–275.

Araki, T., Enomoto, S., Furuno, K. *et al.*, 2005. Experimental investigation of geologically produced antineutrinos with KamLAND. *Nature*, **436**, 499–503.

Armstrong, R.L., 1981. Radiogenic isotopes: the case for crustal recycling on a near-steady-state no-continental-growth Earth. *Phil. Trans. R. Soc. Lond. A*, **301**, 443–472.

Arndt, N., 1999. Why was flood volcanism on submerged continental platforms so common in the Precambrian? *Precambrian Res.*, **97**, 155–164.

Badro, J., Rueff, J.-P., Vanko, G., Monaco, G., Figuet, G. and Guyot, F., 2004. Electronic transition in perovskite: possible non-convecting layers in the lower mantle. *Science*, **305**, 383–386.

Bennett, V.C., 2003. Compositional evolution of the mantle. In: *Treatise on Geochemistry* (H.D. Holland and K.K. Turekian, eds), Vol. 2, pp. 493–519. Elsevier, Amsterdam.

Bercovici, D., 2003. The generation of plate tectonics from mantle convection. *Earth Planet. Sci. Lett.*, **205**, 107–121.

Bercovici, D., Ricard, Y. and Richards, M.A., 2000. The relation between mantle dynamics and plate tectonics: a primer. In: *The History and Dynamics of Global Plate Motions* (M.A. Richards, R.G. Gordon, R.D. van der Hilst, ed.), pp. 5–46. American Geophysical Union, Washington, DC.

Bickle, M.J., 1986. Implications of melting for stabilisation of the lithosphere and heat loss in the Archean. *Earth Planet. Sci. Lett.*, **80**, 314–324.

Bleeker, W., 2003. The late Archean record: a puzzle in *ca.* 35 pieces. *Lithos*, **71**, 99–134.

Boehler, R., 1996. Melting temperature of the Earth's mantle and core: Earth's thermal structure. *Ann. Rev. Earth Planet. Sci.*, **24**, 15–40.

Buffett, B.A., 2002. Estimates of heat flow in the deep mantle based on the power requirements for the geodynamo. *Geophys. Res. Lett.*, **29**, Doi: 10.1029/2001GL014649.

Buffett, B.A., 2003. On the thermal state of Earth's core. *Science*, **299**, 1675–1676.

Butler, S.L., Peltier, W.R. and Costin, S.O., 2005. Numerical models of the Earth's thermal history: effects of inner-core solidification and core potassim. *Phys. Earth Planet. Int.*, **152**, 22–42.
Campbell, I.H., 2003. Constraints on continental growth models from Nb/U ratios in the 3.5 Ga Barberton and other Archaean basalt-komatiite suites. *Am. J. Sci.*, **303**, 319–351.
Campbell, I.H. and Griffiths, R.W., 1990. Implications of mantle plume structure for the evolution of flood basalts. *Earth Planet. Sci. Lett.*, **99**, 79–93.
Campbell, I.H., Griffiths, R.W. and Hill, R.I., 1989. Melting in an Archean mantle plume: heads it's basalts, tails it's komatiits. *Nature*, **339**, 697–699.
Christensen, U.R., 1985. Thermal evolution models for the Earth. *J. Geophys. Res.*, **90**, 2995–3007.
Coffin, M.F. and Eldholm, O., 1994. Large igneous provinces: crustal structure, dimensions, and external consequences. *Rev. Geophys.*, **32**, 1–36.
Collerson, K.D. and Kamber, B.S., 1999. Evolution of the continents and the atmosphere inferred from Th-U-Nb systematics of the depleted mantle. *Science*, **283**, 1519–1522.
Coltice, N., Albarede, F. and Gillet, P., 2000. ^{40}K-^{40}Ar constraints on recycling continental crust into the mantle. *Science*, **288**, 845–847.
Condie, K.C., 1998. Episodic continental growth and supercontinents: a mantle avalanche connection? *Earth Planet. Sci. Lett.*, **163**, 97–108.
Conrad, C.P. and Hager, B.H., 1999a. Effects of plate bending and fault strength at subduction zones on plate dynamics. *J. Geophys. Res.*, **104**, 17,551–17,571.
Conrad, C.P. and Hager, B.H., 1999b. The thermal evolution of an earth with strong subduction zones. *Geophys. Res. Lett.*, **26**, 3041–3044.
Coogan, L.A. and Hinton, R.W., 2006. Do the trace element compositions of detrital zircons require Hadean continental crust? *Geology*, **34**, 633–636.
Daly, S.F., 1980. Convection with decaying heat sources: constant viscosity. *Geophys. J. R. Astron. Soc.*, **61**, 519–547.
Dauphas, N. and Marty, B., 2002. Inference on the nature and the mass of Earth's late veneer from noble metals and gases. *J. Geophys. Res.*, **107**, 5129. Doi: 10.1029/2001JE001617.
Dauphas, N., Robert, F. and Marty, B., 2000. The late asteroidal and cometary bombardment of Earth as recorded in water deuterium to protium ratio. *Icarus*, **148**, 508–512.
Davies, G.F., 1980. Thermal histories of convective Earth models and constraints on radiogenic heat production in the earth. *J. Geophys. Res.*, **85**, 2517–2530.
Davies, G.F., 1992. On the emergence of plate tectonics. *Geology*, **20**, 963–966.
Davies, G.F., 1993. Cooling the core and mantle by plume and plate flows. *Geophys. J. Int.*, **115**, 132–146.
Davies, G.F., 2006. Gravitational depltion of the early Earth's upper mantle and the viability of early plate tectonics. *Earth Planet. Sci. Lett.*, **243**, 376–382.
Davies, G.F., 2007. Mantle regulation of core cooling: a geodynamo without core radioactivity? *Phys. Earth Planet. Int.*, **160**, 215–229.
DePaolo, D.J., 1980. Crustal growth and mantle evolution: inferences from models of element transport and Nd and Sr isotopes. *Geochim. Cosmochim. Acta*, **44**, 1185–1196.
Doin, M.-P., Fleitout, L. and Christensen, U., 1997. Mantle convection and stability of depleted and undepleted continental lithosphere. *J. Geophys. Res.*, **102**, 2772–2787.
Drake, M.J., 2005. Origin of water in the terrestrial planets. *Meteorit. Planet. Sci.*, **40**, 519–527.
Drake, M.J. and Righter, K., 2002. Determining the composition of the Earth. *Nature*, **416**, 39–44.
Durrheim, R.J. and Mooney, W.D., 1991. Archean and Proterozoic crustal evolution: evidence from crustal seismology. *Geology*, **19**, 606–609.
Ernst, R.E. and Buchan, K.L., 2003. Recognizing mantle plumes in the geological record. *Annu. Rev. Earth Planet. Sci.*, **31**, 469–523.
Evans, D.A.D., 2003. True polar wander and supercontinents. *Tectonophysics*, **362**, 303–320.
Farnetani, C.G., 1997. Exess temperature of mantle plumes: the role of chemical stratification across D''. *Geophys. Res. Lett.*, **24**, 1583–1586.
Fei, Y., Van Orman, J., Li, J., van Westrennen, W., Sanloup, C., Minarik, W., Hirose, K., Komabayashi, T., Walter, M. and Funakoshi, K., 2004. Experimentally determined postspinel transformation boundary in Mg_2SiO_4 using MgO as an internal pressure standard and its geophysical implications. *J. Geophys. Res.*, **109**, B02305. Doi: 10.1029/2003JB002652.
Foulger, G.R., Natland, J.H., Presnell, D.C. and Anderson, D.L., 2005. *Plates, Plumes, and Paradigms*. Geological Society of America, Boulder, CO.
Fyfe, W.S., 1978. The evolution of the earth's crust: Modern plate tectonics to ancient hot spot tectonics? *Chem. Geol.*, **23**, 89–114.
Galer, S.J.G., 1991. Interrelationships between continental freeboard, tectonics and mantle temperature. *Earth Planet. Sci. Lett.*, **105**, 214–228.
Galer, S.J.G. and Mezger, K., 1998. Metamorphism, denudation and sea level in the Archean and cooling of the Earth. *Precambrian Res.*, **92**, 389–412.
Genda, H. and Ikoma, M., 2008. Origin of the ocean on the Earth: early evolution of water D/H in a hydrogen-rich atmosphere. *Icarus*, **194**, 42–52.
Gessmann, C.K. and Wood, B.J., 2002. Potassium in the Earth's core? *Earth Planet. Sci. Lett.*, **200**, 63–78.
Grigne, C. and Labrosse, S., 2001. Effects of continents on Earth cooling: thermal blanketing and depletion in radioactive elements. *Geophys. Res. Lett.*, **28**, 2707–2710.
Grigne, C., Labrosse, S. and Tackley, P.J., 2005. Convective heat transfer as a function of wavelength: Implications for the cooling of the Earth. *J. Geophys. Res.*, **110**, B03409. Doi: 10.1029/2004JB003376.
Grimes, C.B., John, B.E., Kelemen, P.B., Mazdab, F.K., Wooden, J.L., Cheadle, M.J., Kanghøj, K. and Schwartz, J.J., 2007. Trace element chemistry of zircons from oceanic crust: a method for distinguishing detrital zircon provenance. *Geology*, **35**, 643–646.
Grove, T.L. and Parman, S.W., 2004. Thermal evolution of the Earth as recorded by komatiites. *Earth Planet. Sci. Lett.*, **219**, 173–187.
Gurnis, M., 1989. A reassessment of the heat transport by variable viscosity convection with plates and lids. *Geophys. Res. Lett.*, **16**, 179–182.
Gurnis, M., Hall, C. and Lavier, L., 2004. Evolving force balance during incipient subduction. *Geochem. Geophys. Geosys.*, **5**, Q07. Doi: 10.1029/2003GC000681.
Halliday, A.N., 2003. The origin and earliest history of the earth. In: *Treatise on Geochemistry* (H.D. Holland and K.K. Turekian, eds), Vol. 1, pp. 509–557, Elsevier, Amsterdam.
Harrison, C.G.A., 1999. Constraints on ocean volume change since the Archean. *Geophys. Res. Lett.*, **26**, 1913–1916.
Harrison, T.M., Blichert-Toft, J., M̈uller, W., Albarede, F., Holden, P. and Mojzsis, S.J., 2005. Heterogeneous Hadean hafnium: evidence of continental crust at 4.4 to 4.5 Ga. *Science*, **310**, 1947–1950.
Hartmann, W.K., Ryder, G., Dones, L. and Grinspoon, D., 2000. The time-dependent intense bombardment of the primordial Earth/Moon system. In: *Orign of the Earth and Moon* (R.M. Canup and K. Righter, eds), pp. 493–512. University of Arizona Press, Tucson, AZ.
Hawkesworth, C.J. and Kemp, A.I.S., 2006. Evolution of the continental crust. *Nature*, **443**, 811–817.
Herzberg, C., 2004. Partial melting below the Ontong Java Plateau. In: *Origin and Evolution of the Ontong Java Plateau* (G. Fitton, J. Mahoney, P. Wallace and

A. Saunders, eds), pp. 179–184. Geological Society of London, London.
Herzberg, C., Asimow, P.D., Arndt, N., Niu, Y., C.Lesher, M., Fitton, J.G., Cheadle, M.J. and Saunders, A.D., 2007. Temperatures in ambient mantle and plumes: constraints from basalts, picrites, and komatiites. *Geochem. Geophys. Geosys.*, **8**, Q02. Doi: 10.1029/2006GC001390.
Hirschmann, M.M., Aubaud, C. and Withers, A.C., 2005. Storage capacity of H_2O in nominally anhydrous minerals in the upper mantle. *Earth Planet. Sci. Lett.*, **236**, 167–181.
Hirth, G. and Kohlstedt, D.L., 1996. Water in the oceanic mantle: Implications for rheology, melt extraction, and the evolution of the lithosphere. *Earth Planet. Sci. Lett.*, **144**, 93–108.
Hirth, G. and Kohlstedt, D., 2003. Rheology of the upper mantle and the mantle wedge: a view from the experimentalists. In: *Inside the Subduction Factory* (J. Eiler, ed.), pp. 83–105. American Geophysical Union, Washington, DC.
Hoffman, P.F., 1997. Tectonic genealogy of North America. In: *Earth Structure: An Introduction to Structural Geology and Tectonics* (B.A. van der Pluijm and S. Marshak, eds), pp. 459–464. McGraw-Hill, New York.
Honda, S., Yuen, D.A., Balachandar, S. and Reuteler, D., 1993. Three-dimensional instabilities of mantle convection with multiple phase transitions. *Science*, **259**, 1308–1311.
Howard, L.N., 1966. Convection at high Rayleigh number. In: *Proceedings of the Eleventh International Congress of Applied Mechanics* (H. Gortler, ed.), pp. 1109–1115. Springer-Verlag, New York.
Hurley, P.M. and Rand, J.R., 1969. Pre-drift continental nuclei. *Science*, **164**, 1229–1242.
Hynes, A., 2001. Freeboard revisited: continental growth, crustal thickness change and Earth's thermal efficiency. *Earth Planet. Sci. Lett.*, **185**, 161–172.
Ito, G. and Mahoney, J.J., 2005. Flow and melting of a heterogeneous mantle: 2. implications for a chemically nonlayered mantle. *Earth Planet. Sci. Lett.*, **230**, 47–63.
Ito, E., Harris, D.M. and Anderson, A.T., 1983. Alteration of oceanic crust and geologic cycling of chlorine and water. *Geochim. Cosmochim. Acta*, **47**, 1613–1624.
Jacob, D., Jagoutz, E., Lowry, D., Mattey, D. and Kudrjavtseva, G., 1994. Diamondiferous eclogites from Siberia: remnants of Archean oceanic crust. *Geochim. Cosmochim. Acta*, **58**, 5191–5207.
Jacobsen, S.B., 1988. Isotopic constraints on crustal growth and recyling. *Earth Planet. Sci. Lett.*, **90**, 315–329.
Jacobsen, S.B. and Wasserburg, G.J., 1979. The mean age of mantle and crustal reservoirs. *J. Geophys. Res.*, **84**, 7411–7427.
Jarrard, R.D., 2003. Subduction fluxes of water, carbon dioxide, chlorine, and potassium. *Geochem. Geophys. Geosys.*, **4**, 8905. Doi: 10.1029/2002GC000392..
Jaupart, C., Labrosse, S. and Mareschal, J.-C., 2007. Temperatures, heat and energy in the mantle of the Earth. In: *Treatise on Geophysics* (G. Schubert, ed.), Vol. 7, pp. 253–303. Elsevier, Amsterdam.
Jellinek, A.M. and Manga, M., 2002. The influence of a chemical boundary layer on the fixity, spacing and lifetime of mantle plumes. *Nature*, **418**, 760–763.
Jochum, K.P., Hofmann, A.W., Ito, E., Seufert, H.M. and White, W.M., 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.
Jordan, T.H., 1988. Structure and formation of the continental tectosphere. *J. Petrol. Spec.*, 11–37.
Karato, S., 1986. Does partial melting reduce the creep strength of the upper mantle? *Nature*, **319**, 309–310.
Karato, S., 1998. Seismic anisotropy in the deep mantle, boundary layers and the geometry of mantle convection. *Pure Appl. Geophys.*, **151**, 565–587.
Karato, S. and Wu, P., 1993. Rheology of the upper mantle: a synthesis. *Science*, **260**, 771–778.
Kasting, J.F. and Holm, N.G., 1992. What determines the volume of the oceans? *Earth Planet. Sci. Lett.*, **109**, 507–515.
Kasting, J.F., Tazewell Howard, M., Wallmann, K., Veizer, J., Shields, G. and Jaffrés, J., 2006. Paleoclimates, ocean depth, and the oxygen isotopic composition of seawater. *Earth Planet. Sci. Lett.*, **252**, 82–93.
Katsura, T., Yamada, H., Shinmei, T., Kubo, A., Ono, S., Kanzaki, M., Yoneda, A., Walter, M.J., Ito, E., Urakawa, S., Funakoshi, K. and Utsumi, W., 2003. Post-spinel transition in Mg_2SiO_4 determined by high $P-T$ in situ X-ray deffractometry. *Phys. Earth Planet. Int.*, **136**, 11–24.
Kellogg, L.H., Hager, B.H. and van der Hilst, R.D., 1999. Compositional stratification in the deep mantle. *Science*, **283**, 1881–1884.
Kinny, P.D. and Nutman, A.P., 1996. Zirconology of the Meeberrie gneiss, Yilgarn Craton, Western Australia: an early Archaean migmatite. *Precambrian Res.*, **78**, 165–178.
Kinzler, R.J., 1997. Melting of mantle peridotite at pressures approaching the spinel to garnet transition: application to mid-ocean ridge basalt petrogenesis. *J. Geophys. Res.*, **102**, 852–874.
Kitajima, K., Maruyama, S., Utsunomiya, S. and Liou, J.G., 2001. Seafloor hydrothermal alteration at an Archaean mid-ocean ridge. *J. Metamor. Geol.*, **19**, 583–599.
Komiya, T., Maruyama, S., Masuda, T., Nohda, S., Hayashi, M. and Okamotmo, K., 1999. Plate tectonics at 3.8–3.7 Ga: field evidence from the Isua accretionary complex, southern west Greenland. *J. Geol.*, **107**, 515–554.
Korenaga, J., 2003. Energetics of mantle convection and the fate of fossil heat. *Geophys. Res. Lett.*, **30**, 1437. Doi: 10.1029/2003GL016982.
Korenaga, J., 2004. Mantle mixing and continental breakup magmatism. *Earth Planet. Sci. Lett.*, **218**, 463–473.
Korenaga, J., 2005a. Firm mantle plumes and the nature of the core-mantle boundary region. *Earth Planet. Sci. Lett.*, **232**, 29–37.
Korenaga, J., 2005b. Why did not the Ontong Java Plateau form subaerially? *Earth Planet. Sci. Lett.*, **234**, 385–399.
Korenaga, J., 2006. Archean geodynamics and the thermal evolution of Earth. In: *Archean Geodynamics and Environments* (K. Benn, J.-C. Mareschal and K. Condie, eds), pp. 7–32. American Geophysical Union, Washington, DC.
Korenaga, J., 2007a. Eustasy, supercontinental insulation, and the temporal variability of terrestrial heat flux. *Earth Planet. Sci. Lett.*, **257**, 350–358.
Korenaga, J., 2007b. Thermal cracking and the deep hydration of oceanic lithosphere: a key to the generation of plate tectonics? *J. Geophys. Res.*, **112**, B05408. Doi: 10.1029/2006JB004502.
Korenaga, J., 2008a. Comment on "Intermittent plate tectonics?". *Science*, **320**, 1291a.
Korenaga, J., 2008b. Urey ratio and the structure and evolution of Earth's mantle. *Rev. Geophys.*, **46**, RG2007. Doi: 10.1029/2007RG000241.
Korenaga, J. and Jordan, T.H., 2002. On the state of sublithospheric upper mantle beneath a supercontient. *Geophys. J. Int.*, **149**, 179–189.
Korenaga, J. and Karato, S., 2008. A new analysis of experimental data on olivine rheology. *J. Geophys. Res.*, **113**, B02403. Doi: 10.1029/2007JB005100.
Korenaga, J. and Kelemen, P.B., 1998. Melt migration through the oceanic lower crust: a constraint from melt percolation modeling with finite solid diffusion. *Earth Planet. Sci. Lett.*, **156**, 1–11.
Korenaga, J. and Kelemen, P.B., 2000. Major element heterogeneity of the mantle source in the North Atlantic igneous province. *Earth Planet. Sci. Lett.*, **184**, 251–268.
Korenaga, T. and Korenaga, J., 2008. Subsidence of normal oceanic

lithosphere, apparent thermal expansivity, and seafloor flattening. *Earth Planet. Sci. Lett.*, **268**, 44–51.
Korenaga, J., Kelemen, P.B. and Holbrook, W.S., 2002. Methods for resolving the origin of large igneous provinces from crustal seismology. *J. Geophys. Res.*, **107**, 2178. Doi: 10.1029/2001B001030.
Labrosse, S., 2002. Hotspots, mantle plumes and core heat loss. *Earth Planet. Sci. Lett.*, **199**, 147–156.
Labrosse, S. and Jaupart, C., 2007. Thermal evolution of the Earth: secular changes and fluctuations of plate characteristics. *Earth Planet. Sci. Lett.*, **260**, 465–481.
Labrosse, S., Poirier, J.-P. and Mouel, J.-L.L., 2001. The age of the inner core. *Earth Planet. Sci. Lett.*, **190**, 111–123.
Langmuir, C.H., Klein, E.M. and Plank, T., 1992. Petrological systematics of mid-ocean ridge basalts: constraints on melt generation beneath ocean ridges. In: *Mantle Flow and Melt Generation at Mid-Ocean Ridges, Geophys. Monogr. Ser.* (J. Phipps Morgan, D.K. Blackman and J.M. Sinton, eds), Vol. 71, pp. 183–280. AGU, Washington, DC.
Lassiter, J.C., 2006. Constraints on the coupled thermal evolution of the Earth's core and mantle, the age of the inner core, and the origin of the $^{186}Os/^{188}Os$ "core signal" in plume-derived lavas. *Earth Planet. Sci. Lett.*, **250**, 306–317.
Lee, K.K.M., Steinle-Neumann, G. and Jeanloz, R., 2004. Ab-initio high-pressure alloying of iron and potassium: implications for the Earth's core. *Geophys. Res. Lett.*, **31**, L11603. Doi: 10.1029/2004GL019839.
Lenardic, A. and Moresi, L.-N., 1999. Some thoughts on the stability of cratonic lithosphere: effects of buoyancy and viscosity. *J. Geophys. Res.*, **104**, 12,747–12,758.
Lin, S.-C. and van Keken, P.E., 2005. Multiple volcanic episodes of flood basalts caused by thermochemical mantle plumes. *Nature*, **436**, 250–252.
Lin, J., Struzhkin, V., Jacobsen, S., Hu, M., Chow, P., Kung, J., Liu, H., Mao, H. and Hemley, R., 2005. Spin transition of iron in magnesiowüste in the Earth's lower mantle. *Nature*, **436**, 377–380.
Lyubetskaya, T. and Korenaga, J., 2007a. Chemical composition of Earth's primitive mantle and its variance, 1, methods and results. *J. Geophys. Res.*, **112**, B03211. Doi: 10.1029/2005JB004223.
Lyubetskaya, T. and Korenaga, J., 2007b. Chemical composition of Earth's primitive mantle and its variance, 2, implications for global geodynamics. *J. Geophys. Res.*, **112**, B03212. Doi: 10.1029/2005JB004224.
Machetel, P. and Weber, P., 1991. Intermittent layered convection in a model with an endothermic phase change at 670 km. *Nature*, **350**, 55–57.
Marty, B. and Yokochi, R., 2006. Water in the early Earth. *Rev. Mineral. Geochem.*, **62**, 421–450.
McBirney, A.R., 1993. *Igneous Petrology*, 2nd edn. Jones and Bartlett Publishers, Boston, MA.
McDonough, W.F., 2003. Compositional model for the Earth's core. In: *Treatise on Geochemistry* (H.D. Holland and K.K. Turekian, eds), Vol. 2, pp. 547–568. Elsevier, Amsterdam.
McDonough, W.F., 2005. Ghosts from within. *Nature*, **436**, 467–468.
McDonough, W.F. and Sun, S.-s., 1995. The composition of the Earth. *Chem. Geol.*, **120**, 223–253.
McElhinny, M.W. and Senanayake, W.E., 1980. Paleomagnetic evidence for the existence of the geomagnetic field 3.5 Ga ago. *J. Geophys. Res.*, **85**, 3523–3528.
McKenzie, D. and Weiss, N., 1975. Speculations on the thermal and tectonic history of the Earth. *Geophys. J. R. Astron. Soc.*, **42**, 131–174.
McLennan, S.M. and Taylor, R.S., 1982. Geochemical constraints on the growth of the continental crust. *J. Geol.*, **90**, 347–361.
McNamara, A.K. and van Keken, P.E., 2000. Cooling of the Earth: a parameterized convection study of whole versus layered models. *Geochem. Geophys. Geosys.*, **1**, 2000GC000045.
Meisel, T., Walker, R.J. and Morgan, J.W., 1996. The osmium isotopic composition of the Earth's primitive upper mantle. *Nature*, **383**, 517–520.
Meisel, T., Walker, R.J., Irving, A.J. and Lorand, J.-P., 2001. Osmium isotopic compositions of mantle xenoliths: a global perspective. *Geochim. Cosmochim. Acta*, **65**, 1311–1323.
Mojzsis, S.J., Harrison, T.M. and Pidgeon, R.T., 2001. Oxygen-isotope evidence from ancient zircons for liquid water at the Earth's surface 4,300 Myr ago. *Nature*, **409**, 178–181.
Morbidelli, A., Chambers, J., Lunine, J.I., Petit, J.M., Robert, F., Valsecchi, G.B. and Cyr, K.E., 2000. Source regions and timescales for the delivery of water to the Earth. *Meteorit. Planet. Sci.*, **35**, 1309–1320.
Moresi, L. and Solomatov, V., 1998. Mantle convection with a brittle lithosphere: thoughts on the global tectonic styles of the Earth and Venus. *Geophys. J. Int.*, **133**, 669–682.
Morgan, W.J., 1971. Convection plumes in the lower mantle. *Nature*, **230**, 42–43.
Morgan, J.W., 1986. Ultramafic xenoliths – clues to Earth's late accretional history. *J. Geophys. Res.*, **91**, 2375–2387.
Murakami, M., Hirose, K., Kawamura, K., Sata, N. and Ohishi, Y., 2004. Post-perovskite phase transition in MgSiO3. *Science*, **304**, 855–858.
Nimmo, F., Price, G.D., Brodholt, J. and Gubbins, D., 2004. The influence of potassium on core and geodynamo evolution. *Geophys. J. Int.*, **156**, 363–376.
Nisbet, E.G., Arndt, N.T., Bickle, M.J., Cameron, W.E., Chauvel, C., Cheadle, M., Hegner, E., Kyser, T.K., Martin, A., Renner, R. and Roedder, E., 1987. Uniquely fresh 2.7 Ga komatiites from the Belingwe greenstone belt, Zimbabwe. *Geology*, **15**, 1147–1150.
Nutman, A.P., 2006. Antiquity of the oceans and continents. *Elements*, **2**, 223–227.
Nutman, A.P., Friend, C.R.L. and Bennett, V.C., 2002. Evidence for 3650–3600 Ma assembly of the northern end of the Itsaq Gneiss Complex, Greenland: implication for early Archean tectonics. *Tectonics*, **21**, 1005. Doi: 10.1029/2000TC001203.
O'Hara, M.J., 1975. Is there an icelandic mantle plume? *Nature*, **253**, 708–710.
O'Neill, C., Lenardic, A., Moresi, L., Torsvik, T.H. and Lee, C.-T., 2007. Episodic Precambrian subduction. *Earth Planet. Sci. Lett.*, **262**, 552–562.
O'Nions, R.K., Evensen, N.M. and Hamilton, P.J., 1979. Geochemical modeling of mantle differentiation and crustal growth. *J. Geophys. Res.*, **84**, 6091–6101.
Oganov, A.R. and Ono, S., 2004. Theoretical and experimental evidence for a post-perovskite phase of $MgSiO_3$ in Earth's D'' layer. *Nature*, **430**, 445–448.
Ogawa, M., 2003. Plate-like regime of a numerically modeled thermal convection in a fluid with temperature-, pressure-, and stress-history-dependent viscosity. *J. Geophys. Res.*, **108**, 2067. Doi: 10.1029/2000JB000069.
Ohmoto, H., 1996. Formation of volcanogenic massive sulfide deposits: the Kuroko perspective. *Ore Geol. Rev.*, **10**, 135–177.
Parman, S.W., 2007. Helium isotopic evidence for episodic mantle melting and crustal growth. *Nature*, **446**, 900–903.
Parman, S.W., Dann, J.C., Grove, T.L. and de Wit, M.J., 1997. Emplacement conditions of komatiite magmas from the 3.49 Ga Komati Formation, Barberton Greenstone Belt, South Africa. *Earth Planet. Sci. Lett.*, **150**, 303–323.
Parsons, B., 1982. Causes and consequences of the relation between area and age of the ocean floor. *J. Geophys. Res.*, **87**, 289–302.
Parsons, B. and McKenzie, D., 1978. Mantle convection and the thermal structure of the plates. *J. Geophys. Res.*, **83**, 4485–4496.
Parsons, B. and Sclater, J.G., 1977. An analysis of the variation of ocean floor

bathymetry and heat flow with age. *J. Geophys. Res.*, **82**, 803–827.
Pearson, D.G., 1999. The age of continental roots. *Lithos*, **48**, 171–194.
Phipps Morgan, J. and Chen, Y., 1993. The genesis of oceanic crust: Magma injection, hydrothermal circulation, and crustal flow. *J. Geophys. Res.*, **98**, 6283–6297.
Pollack, H.N., Hurter, S.J. and Johnson, J.R., 1993. Heat flow from the Earth's interior: analysis of the global data set. *Rev. Geophys.*, **31**, 267–280.
Rama Murthy, V., van Westrenen, W. and Fei, Y., 2003. Experimental evidence that potassium is a substantial radioactive heat source in planetary cores. *Nature*, **423**, 163–165.
Regenauer-Lieb, K., Yuen, D.A. and Branlund, J., 2001. The initiation of subduction: criticality by addition of water? *Science*, **294**, 578–580.
Reymer, A. and Schubert, G., 1984. Phanerozoic addition rates to the continental crust and crustal growth. *Tectonics*, **3**, 63–77.
Richards, M.A., Duncan, R.A. and Courtillot, V.E., 1989. Flood basalts and hot-spot tracks: plume heads and tails. *Science*, **246**, 103–107.
Richards, M.A., Yang, W.-S., Baumgardner, J.R. and Bunge, H.-P., 2001. Role of a low-viscosity zone in stabilizing plate tectonics: implications for comparative terrestrial planetology. *Geochem. Geophys. Geosys.*, **2**, 2000GC000115.
Richter, F.M., 1985. Models for the archean thermal regime. *Earth Planet. Sci. Lett.*, **73**, 350–360.
Ringwood, A.E. and Green, D.H., 1964. Experimental investigations bearing on the nature of the Mohorovicic discontinuity. *Nature*, **201**, 566–567.
Rino, S., Komiya, T., Windley, B.F., Katayama, I., Motoki, A. and Hirata, T., 2004. Major episodic increases of continental crustal growth determined from zircon ages of river sands; implications for mantle overturns in the Early Precambrian. *Phys. Earth Planet. Int.*, **146**, 369–394.
Robert, F., 2003. The D/H ratio in chondrites. *Space Sci. Rev.*, **106**, 87–101.
Rudnick, R.L. and Gao, S., 2003. Composition of the continental crust. In: *Treatise on Geochemistry* (H.D. Holland and K.K. Turekian, eds), Vol. 3, pp. 1–64. Elsevier, Amsterdam.
Rüpke, L.H., Phipps Morgan, J., Hort, M. and Connolly, J.A.D., 2004. Serpentine and the subduction zone water cycle. *Earth Planet. Sci. Lett.*, **223**, 17–34.
Ryder, G., 2002. Mass flux in the ancient Earth-Moon system and benign implications for the origin of life on Earth. *J. Geophys. Res.*, **107**, 5022. Doi: 10.1029/2001JE001583.
Saal, A.E., Hauri, E.H., Langmuir, C.H. and Perfit, M.R., 2002. Vapour undersaturation in primitive mid-ocean-ridge basalt and the volatile content of Earth's upper mantle. *Nature*, **419**, 451–455.
Salters, V.J.M. and Stracke, A., 2004. Composition of the depleted mantle. *Geochem. Geophys. Geosys.*, **5**, Q05004. Doi: 10.1029/2003GC000597.
Sandiford, M., Van Kranendonk, M.J. and Bodorkos, S., 2004. Conductive incubation and the origin of dome-and-keel structure in Archean granite-greenstone terrains: a model based on the eastern Pilbara Craton, Western Australia. *Tectonics*, **23**, TC1009. Doi: 10.1029/2002TC001452.
Schubert, G. and Reymer, A.P.S., 1985. Continental volume and freeboard through geological time. *Nature*, **316**, 336–339.
Schubert, G., Stevenson, D. and Cassen, P., 1980. Whole planet cooling and the radiogenic heat source contents of the earth and moon. *J. Geophys. Res.*, **85**, 2531–2538.
Shirey, S.B., Carlson, R.W., Richardson, S.H., Menzies, A., Gurney, J., Pearson, D.G., Harris, J.W. and Wiechert, U., 2001. Archean emplacement of eclogite components into the lithospheric mantle during formation of the Kaapvaal Craton. *Geophys. Res. Lett.*, **28**, 2509–2512.
Shirey, S.B., Kamber, B.S., Whitehouse, M.J., Mueller, P.A. and Basu, A.R., 2008. A review of the isotopic evidence for mantle and crustal processes in the Hadean and Archian: implications for the onset plate tectonic subduction. In: *When Did Plate Tectonics Start on Earth?* (K.C. Condie and V. Pease, eds), pp. 1–29. Geological Society of America, Boulder, CO.
Silver, P.G. and Behn, M.D., 2008. Intermittent plate tectonics? *Science*, **319**, 85–88.
Sleep, N.H., 1984. Tapping of magmas from ubiquitous mantle heterogeneities: an alternative to mantle plumes? *J. Geophys. Res.*, **89**, 10,029–10,041.
Sleep, N.H., 2000. Evolution of the mode of convection within terrestrial planets. *J. Geophys. Res.*, **105**, 17,563–17,578.
Sleep, N.H., 2007. Plate tectonics through time. In: *Treatise on Geophysics* (G. Schubert, ed.), Vol. 9, pp. 101–117. Elsevier, Amsterdam.
Sleep, N.H. and Windley, B.F., 1982. Archean plate tectonics: constraints and inferences. *J. Geol.*, **90**, 363–379.
Sleep, N.H. and Zahnle, K., 2001. Carbon dioxide cycling and implications for climate on ancient Earth. *J. Geophys. Res.*, **106**, 1373–1399.
Sleep, N.H., Richards, M.A. and Hager, B.H., 1988. Onset of mantle plumes in the presence of preexisting convection. *J. Geophys. Res.*, **93**, 7672–7689.
Smyth, J.R. and Jacobsen, S.D., 2006. Nominally anhydrous minerals and Earth's deep water cycle. In: *Earth's Deep Water Cycle* (S. Jacobsen and S. van der Lee), pp. 1–11. American Geophysical Union, Washington, DC.
Solheim, L.P. and Peltier, W.R., 1994. Avalanche effects in phase transition modulated thermal convection. *J. Geophys. Res.*, **99**, 6997–7018.
Solomatov, V.S., 1995. Scaling of temperature- and stress-dependent viscosity convection. *Phys. Fluids*, **7**, 266–274.
Solomatov, V.S., 1996. Can hotter mantle have a larger viscosity? *Geophys. Res. Lett.*, **23**, 937–940.
Solomatov, V.S., 2001. Grain size-dependent viscosity convection and the thermal evolution of the earth. *Earth Planet. Sci. Lett.*, **191**, 203–212.
Solomatov, V., 2007. Magma oceans and primordial mantle differentiation. In: *Treatise on Geophysics* (G. Schubert, ed.), Vol. 9, pp. 91–119. Elsevier, Amsterdam.
Spohn, T. and Breuer, D., 1993. Mantle differentiation through continental growth and recycling and the thermal evolution of the Earth. In: *Evolution of the Earth and Planets, Geophys. Monogr. Ser*. (E. Takahashi, R. Jeanloz and R. Rudie, eds), pp. 55–71. AGU, Washington, DC.
Stacey, F.D., 1981. Cooling of the Earth – a constraint on paleotectonic hypotheses. In: *Evolution of the Earth* (R.J. O'Connell and W.S. Fyfe, eds), pp. 272–276. American Geophysical Union, Washington, DC.
Stein, M. and Hofmann, A.W., 1994. Mantle plumes and episodic crustal growth. *Nature*, **372**, 63–68.
Stein, C.A. and Stein, S., 1992. A model for the global variation in oceanic depth and heat flow with lithospheric age. *Nature*, **359**, 123–129.
Stein, C., Schmalzl, J. and Hansen, U., 2004. The effect of rheological parameters on plate behavior in a self consistent model of mantle convection. *Phys. Earth Planet. Int.*, **142**, 225–255.
Stevenson, D.J., 2003. Styles of mantle convection and their influence on planetary evolution. *C. R. Geosci.*, **335**, 99–111.
Stevenson, D.J., Spohn, T. and Schubert, G., 1983. Magnetism and thermal evolution of the terrestrial planets. *Icarus*, **54**, 466–489.
Tackley, P.J., 2000. Mantle convection and plate tectonics: toward an integrated physical and chemical theory. *Science*, **288**, 2002–2007.
Tackley, P.J. and Stevenson, D., 1993. A mechanism for spontaneous self-perpet-

uating volcanism on the terrestorial planets. In: *Flow and Creep in the Solar System: Observations, Modeling and Theory* (D.B. Stone and S.K. Runcorn, eds), pp. 307–321. Kluwer Academic, Dordrecht.

Tarduno, J.A., Cottrel, R.D., Watkeys, M.K. and Bauch, D., 2007. Geomagnetic field strength 3.2 billion years ago recorded by single silicate crystals. *Nature*, **446**, 657–660.

Taylor, S.R. and McLennan, S.M., 1985. *The Continental Crust: its Composition and Evolution*. Blackwell, Boston.

van Thienen, P., Vlaar, N.J. and van den Berg, A.P., 2004. Plate tectonics on the terrestrial planets. *Phys. Earth Planet. Int.*, **142**, 61–74.

Tozer, D.C., 1985. Heat transfer and planetary evolution. *Geophys. Surv.*, **7**, 213–246.

Turcotte, D.L. and Schubert, G., 1982. *Geodynamics: Applications of Continuum Physics to Geological Problems*. John Wiley, New York.

Valley, J.W., Lackey, J.S., Clechenko, A.J.C.C.C., Spicuzza, M.J., Bindeman, M.A.S.B.I.N., Ferreira, V.P., Sial, A.N., Peck, E.M.K.W.H., Sinha, A.K. and Wei, C.S., 2005. 4.4 billion years of crustal maturation: oxygen isotope ratios of magmatic zircon. *Contrib. Mineral. Petrol.*, **150**, 561–580.

Van Kranendonk, M.J., Smithies, R.H., Hickman, A.H. and Champion, D., 2007. Review: secular tectonic evolution of Archean continental crust: interplay between horizontal and vertical processes in the formation of the Pilbara Craton, Australia. *Terra Nova*, **19**, 1–38.

Walter, M.J., 1998. Melting of garnet peridotite and the origin of komatiite and depleted lithosphere. *J. Petrol.*, **39**, 29–60.

Wedepohl, K.H., 1995. The composition of the continental crust. *Geochim. Cosmochim. Acta*, **59**, 1217–1232.

White, R.S. and McKenzie, D., 1995. Mantle plumes and flood basalts. *J. Geophys. Res.*, **100**, 17,543–17,585.

Wilde, S.A., Valley, J.W., Peck, W.H. and Graham, C.M., 2001. Evidence from detrial zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago. *Nature*, **409**, 175–178.

Williams, Q., 1998. The temperature contrast across D''. In: *The Core-Mantle Boundary Region* (M. Gurnis, M.E. Wysession, E. Knittle and B.A. Buffett, eds), pp. 73–81. American Geophysical Union, Washington, DC.

Williams, D.R. and Pan, V., 1992. Internally heated mantle convection and the thermal and degassing history of the Earth. *J. Geophys. Res.*, **97**, 8937–8950.

Wilson, M., 1989. *Igneous Petrogenesis*. Springer, Netherlands.

Wilson, D.S., Teagle, D.A.H., Alt, J.C. *et al.*, 2006. Drilling to gabbro in intact ocean crust. *Science*, **312**, 1016–1020.

Wise, D.U., 1974. Continental margins, freeboard and the volumes of continents and oceans through time. In: *Geology of Continental Margins* (C.A. Burk and C.L. Drake, eds), pp. 45–58. Springer, New York.

Workman, R.K. and Hart, S.R., 2005. Major and trace element composition of the depleted MORB mantle (DMM). *Earth Planet. Sci. Lett.*, **231**, 53–72.

Yaxley, G.M., 2000. Experimental study of the phase and melting relations of homogeneous basalt + peridotite mixtures and implications for the petrogenesis of flood basalts. *Contrib. Mineral. Petrol.*, **139**, 326–338.

Received 2 April 2008; revised version accepted 5 September 2008

