# Initiation and Evolution of Plate Tectonics on Earth: Theories and Observations

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mantle convection, oceanic lithosphere, subduction, continental growth, thermal history, oceans

## Abstract

The inception of plate tectonics on Earth and its subsequent evolution are discussed on the basis of theoretical considerations and observational constraints. The likelihood of plate tectonics in the past depends on what mechanism is responsible for the relatively constant surface heat flux that is indicated by the likely thermal history of Earth. The continuous operation of plate tectonics throughout Earth's history is possible if, for example, the strength of convective stress in the mantle is affected by the gradual subduction of surface water. Various geological indicators for the emergence of plate tectonics are evaluated from a geodynamical perspective, and they invariably involve certain implicit assumptions about mantle dynamics, which are either demonstrably wrong or yet to be explored. The history of plate tectonics is suggested to be intrinsically connected to the secular evolution of the atmosphere, through sea-level changes caused by ocean-mantle interaction.

## 1. INTRODUCTION

> "It is the larger conception which determines the expression of the details."
>
> —*Joseph Barrell* (Barrell 1919, p. 282)

Five decades after the advent of the plate tectonics theory (e.g., Hess 1962, Vine & Matthews 1963, Wilson 1965), our understanding of geology seems to have matured enough to discuss the initiation of plate tectonics in Earth's history, which might have been regarded in the past century as too speculative to be legitimate. In recent years, quite a few papers have been published to suggest when plate tectonics started, with proposed timings covering almost the entire history of Earth (**Figure 1**). The diversity of opinions results from ambiguities in the interpretation of relevant geological observations as well as different weightings on different kinds of data. Stern (2005), for example, suggests that modern-style plate tectonics started around the beginning of the Neoproterozoic era [~1 billion years ago (1 Gya)] on the basis of the absence of ultrahigh-pressure

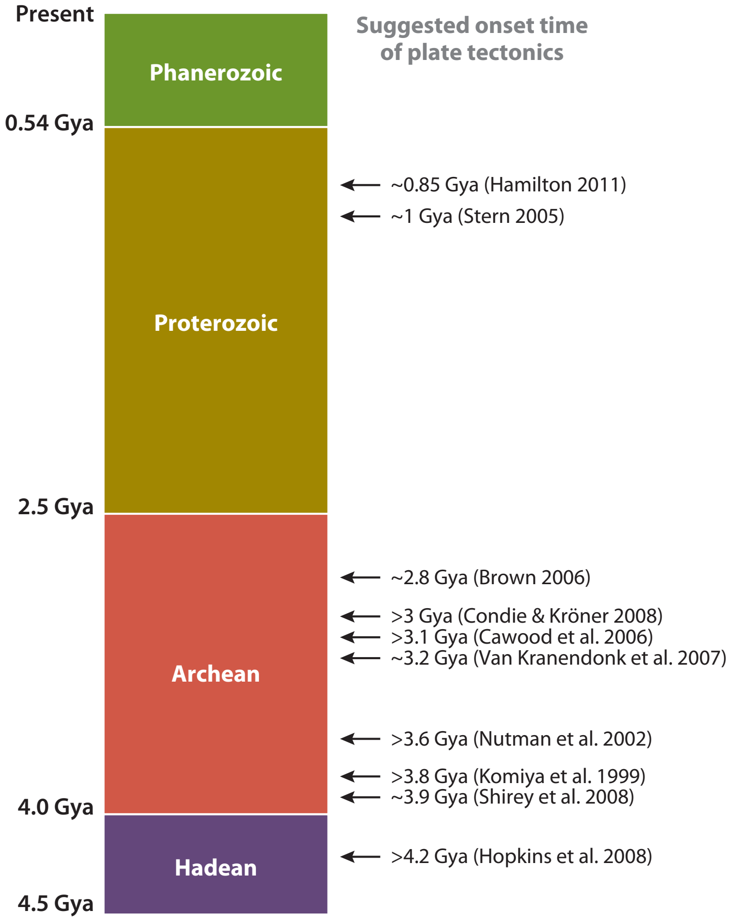


**Figure 1**

Geologic timescale and suggestions for the onset time of plate tectonics. Suggestions shown here merely demonstrate the diversity of opinions published in the past decade or so and are not meant to be a comprehensive compilation of recent literature.

metamorphism and ophiolites before the era, both of which are considered to be prima facie evidence for the operation of plate tectonics. Many other geologists, however, prefer the onset of plate tectonics sometime during the Archean eon (e.g., Komiya et al. 1999, Brown 2006, Cawood et al. 2006, Van Kranendonk et al. 2007, Shirey et al. 2008, Condie & Kröner 2008) because other indicators for plate tectonics such as orogens, accretionary prisms, and paired metamorphic belts can be traced back at least to the late Archean. Some authors suggest that, on the basis of the geochemistry of Hadean zircons, plate tectonics may have been in action already in the Hadean (e.g., Hopkins et al. 2008, 2010).

The difficulty of finding unambiguous geological evidence for the onset of plate tectonics may be appreciated from how these geological eons are defined. Unlike the Proterozoic-Phanerozoic boundary, which is marked by the appearance of abundant fossil life, the other two boundaries are defined by the scarcity of geological samples. The Archean-Proterozoic boundary is defined by the relative abundance of rocks—i.e., rocks of Archean ages are far rarer than those of Proterozoic ages—and the Hadean-Archean boundary is marked by the oldest rock on Earth. There is no rock sample found from the Hadean, and the mineral zircon is currently the only way to probe this deepest eon. As we try to explore more deeply in time, therefore, geological data preserved to the present become more limited in space and more sporadic in time. The reconstruction of the history of plate tectonics on Earth needs to deal with this fundamental limitation on geological observations.

This review therefore places an emphasis on a theoretical approach, with the following premise: Plate tectonics is currently taking place on Earth, so if we understand why it is happening, we may be able to infer the past by extrapolating from the present. Having a theoretical framework for Earth's evolution also allows us to better interpret geological data and appreciate their significance. The importance of such theoretical underpinning may also be understood from a planetary science perspective. The recent discovery of Earth-like planets in other solar systems (e.g., Rivera et al. 2005, Borucki et al. 2011) has invigorated the discussion of habitable planets (e.g., Gaidos et al. 2005, Zahnle et al. 2007), and the operation of plate tectonics is often considered to be essential for habitability (e.g., Kasting & Catling 2003). The physical theory of plate tectonics thus has important applications to planetary habitability and the origins of life in the universe (e.g., Korenaga 2012). However, if we do not have a theory to explain why plate tectonics initiated on Earth and how it evolved with time, we cannot apply our understanding to other planets under different physical conditions. Finding convincing geological evidence for the onset of plate tectonics is one thing, but resolving its underlying physics is another. Compared with other planets within and outside our solar system, Earth is immensely more accessible, so a future theory for planetary evolution depends on how well we can decipher the history of plate tectonics on Earth through both observational and theoretical efforts.

The structure of this review is as follows. I first go through various theoretical considerations, starting with a brief summary of the modes of mantle convection and the condition for plate tectonics. A simple theoretical argument suggests that the operation of plate tectonics is more likely in the past than at the present, but when considered jointly with the thermal evolution of Earth, the likelihood of plate tectonics in the past becomes uncertain. The thermal evolution of Earth nonetheless provides a useful platform to consider a variety of theoretical issues and observational constraints in a coherent manner, and as an example, I discuss the buoyancy of oceanic lithosphere, which is often considered a major obstacle for plate tectonics in the Precambrian (i.e., prior to the Phanerozoic). The theoretical part ends with a summary of the possibility of plate tectonics in the early Earth. In light of this theoretical understanding, I then discuss relevant geological observations, such as plate tectonics indicators and the secular evolution of metamorphism, together with the possibility of intermittent plate tectonics. The observational part concludes with some remarks on preservation bias and its possible causes. The scarcity of geological samples from the Hadean and Archean eons may be a natural consequence of the peculiar thermal evolution of Earth. In the remaining part of this review, I discuss some outstanding issues that are important not only for the history of plate tectonics but also for the evolution of Earth as a whole.

# 2. THEORETICAL CONSIDERATIONS

## 2.1. Why Does Plate Tectonics Happen on Earth?

In our solar system, Earth is the only planet that exhibits plate tectonics. Other planets such as Venus and Mars are believed to be in the mode of stagnant lid convection (**Figure 2*a***). The absence of plate tectonics on those planets is easier to explain than its presence on Earth because stagnant lid convection is the most natural mode of convection with strongly temperature-dependent viscosity (Solomatov 1995). The viscosity of silicate rocks that constitute a planetary mantle is extremely sensitive to temperature, as indicated by the following Arrhenius relation:

$$\eta(T) \propto \exp\left(\frac{E}{RT}\right), \tag{1}$$

where $E$ is the activation energy (typically on the order of a few hundred kilojoules), $R$ is the universal gas constant, and $T$ is the absolute temperature. With this temperature dependency, the viscosity varies over many orders of magnitude across the top thermal boundary layer (**Figure 2*d***), leading to a single rigid lid covering an entire planetary surface, and convection takes place only beneath the rigid lid.

In plate tectonics, the top thermal boundary layer (or equivalently, lithosphere or plates) can deform considerably and sink into the deep mantle (**Figure 2*b***). The recycling of the top boundary layer is what distinguishes between these two modes of mantle convection, as it enables geochemical cycles between the surface and the deep interior, whereas stagnant lid convection allows only one-way mass transfer from the mantle to the surface by magmatism. In this review article, therefore, the term plate tectonics is used in a broad sense, referring to mantle convection with the subduction of surface plates. From a dynamical point of view, subduction is such a drastic difference that further classification based on the details of plate tectonics, such as the rigidity of individual plates, is secondary, although such details may sometimes matter when interpreting geological data.

What makes plate tectonics possible on Earth? An obvious candidate to compensate for temperature-dependent viscosity is the brittle deformation mechanism, which is effective under low temperatures and pressures. The brittle strength of rocks is limited by frictional resistance (Scholz 2002), with the yield stress

$$\tau_y \sim \tau_0 + \mu \rho g z, \tag{2}$$

where $\tau_0$ is the cohesive strength, $\mu$ is the coefficient of friction, $\rho$ is the rock density, and $z$ is the depth. The yield stress is proportional to lithostatic pressure, and because the friction coefficient for rocks is generally on the order of unity (Byerlee 1978), the stress quickly increases with depth and does not help reduce the overall strength of the lithosphere (**Figure 2*e***). For brittle failure to compensate for temperature-dependent viscosity, therefore, the friction coefficient has to be reduced by an order of magnitude (or more), and this is in fact an approach commonly taken in the numerical simulation of plate tectonics (e.g., Moresi & Solomatov 1998, Richards et al. 2001, Stein et al. 2004, van Heck & Tackley 2008, Korenaga 2010). At lithospheric scale, the cohesive strength $\tau_0$ is practically zero (Byerlee 1978). By setting $\mu$ to zero and varying $\tau_0$, some numerical studies explored depth-independent yield stress, which seems important in generating

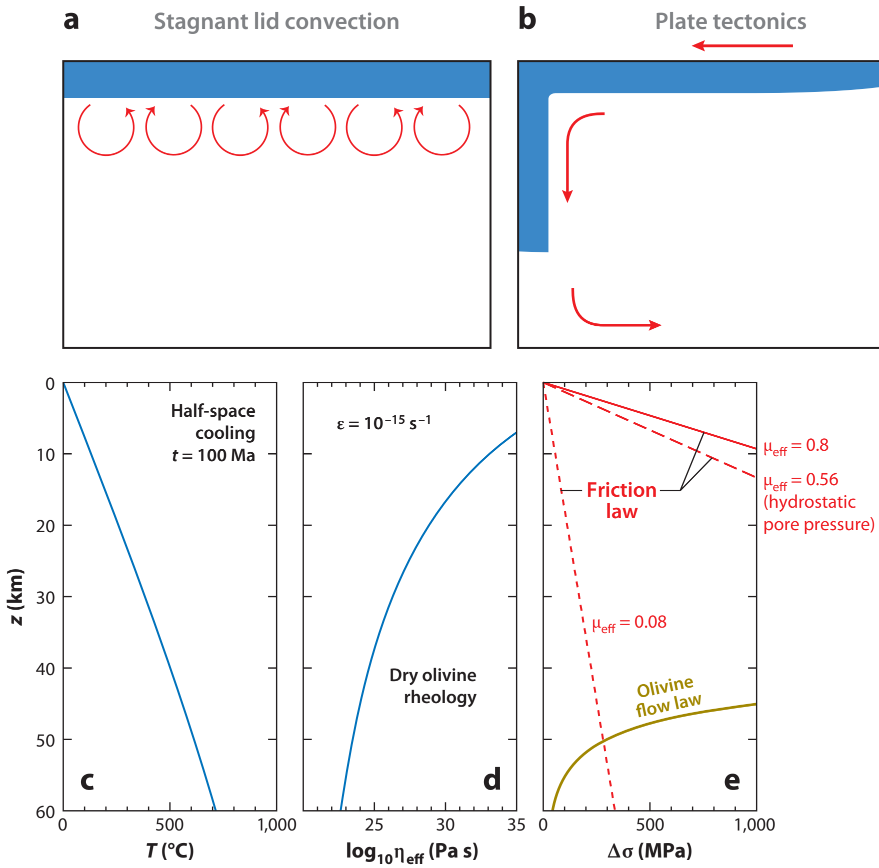


**Figure 2**

Two contrasting modes of mantle convection: (*a*) stagnant lid convection and (*b*) plate tectonics. In the latter, the top thermal boundary layer is continuously recycled back to the mantle. The operation of plate tectonics requires some mechanism that can compensate for the strongly temperature-dependent viscosity of silicate rocks, as illustrated by (*d*) effective viscosity (calculated with a geological strain rate of $10^{-15}$ $s^{-1}$) and (*e*) corresponding yield stress across mature (100-Ma-old) oceanic lithosphere, the temperature profile of which is shown in panel *c*. Also shown in panel *e* are three cases for brittle yield stress: dry faulting with $\mu = 0.8$, wet faulting with hydrostatic pore pressure ($\mu_{eff} = 0.56$), and wet faulting with high pore pressure ($\mu_{eff}$ in this case is arbitrarily reduced to 0.08). These brittle yield stresses are calculated with a formula for optimal thrust faulting (Turcotte & Schubert 1982), which is more precise than Equation (2). After Korenaga (2007).

an intermediate mode of convection termed intermittent plate tectonics (oscillating between plate tectonics and stagnant lid convection) (e.g., Moresi & Solomatov 1998). The use of such constant yield stress is, however, difficult to justify because one has to explain why the cohesive strength can be significant and, at the same time, why the friction coefficient can be ignored. The effect of brittle failure can thus be succinctly represented by the friction coefficient $\mu$ alone.

Is a reduction in the friction coefficient consistent with rock mechanics? A short answer is yes, although the issue is complicated. When the pore space within rocks is filled with water, the pore fluid pressure can reduce the shear strength of the rocks because the shear strength is equal to the friction coefficient times the difference between the lithostatic and pore pressures. The presence of water at depth could therefore facilitate frictional sliding, and the net effect is commonly expressed by the effective friction coefficient, $\mu_{eff}$. However, oceanic lithosphere, the deformation of which is central to the operation of plate tectonics, is dehydrated by melting at mid-ocean ridges (Hirth & Kohlstedt 1996, Evans et al. 2005), so to reduce its effective friction coefficient, there must be a process that can rehydrate the lithosphere deeply (e.g., down to ~50 km for 100-Ma-old lithosphere; see **Figure 2*e***). Plate bending at subduction zones, for example, may fracture oceanic lithosphere to substantial depths (Ranero et al. 2003). We should be aware, however, of the risk of a chicken-and-egg problem if we call for a process that can happen only within the mode of plate tectonics because such a process cannot be used when discussing how plate tectonics can be initiated. For the same reason, we cannot invoke a variety of dynamic weakening mechanisms suggested by the studies of earthquake dynamics, such as flash heating at highly stressed frictional microcontacts (e.g., Rice 2006), because all these mechanisms require already ongoing slip along a preexisting fault. In this regard, the deep hydration of oceanic lithosphere by thermal cracking appears promising because it requires only surface water as a prerequisite (Korenaga 2007). Thermal contraction within a cooling lithosphere is predicted to generate thermal stress high enough to deeply fracture the stiffest part of the lithosphere. Moreover, strongly temperature-dependent viscosity becomes beneficial in this mechanism; thermal stress is higher for greater temperature dependency. Cracking alone, however, does not reduce the effective friction coefficient sufficiently. If pore water is connected to surface water (i.e., open cracks), its pressure should be hydrostatic, and such pressure affects the brittle yield stress only slightly (**Figure 2*e***). For a drastic reduction in the friction coefficient, therefore, pore water should be isolated by serpentinization reaction (Korenaga 2007), but a likely path for the physicochemical evolution of thermally cracked lithosphere is yet to be investigated.

As an alternative to brittle failure, grain size reduction via deformational work has been proposed (Bercovici & Ricard 2005, Landuyt et al. 2008), but whether this mechanism can overcome strongly temperature-dependent viscosity is uncertain. Also, the absence of plate tectonics on Venus cannot be explained by this mechanism alone, but it can be explained by the lack of shear strength reduction with high pore fluid pressure, which can operate only with surface water. Landuyt & Bercovici (2009) suggested that rapid grain growth under the high surface temperature of Venus may counteract grain size reduction, but grain growth in the lithospheric mantle is likely to be inhibited by orthopyroxene pinning even at high temperatures (Hiraga et al. 2010, Chu & Korenaga 2012). Nonetheless, grain size reduction is an important mechanism that generates and preserves localized zones of weakness, and its significance may be better appreciated in combination with brittle failure such as thermal cracking. Also, the aforementioned dynamic weakening mechanisms can set in once plate tectonics is initiated. Modern-style plate tectonics surely exploits various kinds of preexisting weaknesses, e.g., by converting a fracture zone into a subduction zone (e.g., Hall et al. 2003). It is thus imperative to explore how the present-day oceanic lithosphere is damaged by observational means and identify a mechanism that can operate even in the absence of plate tectonics.

## 2.2. Likelihood of Plate Tectonics in the Past

Without the exact knowledge of the actual mechanism that reduces the strength of oceanic lithosphere, deriving a physical condition for plate tectonics is still possible by focusing on its most peculiar dynamics, i.e., the bending of a strong plate at subduction. Even with various weakening mechanisms considered, the effective viscosity of oceanic lithosphere, $\eta_L$, is likely to be higher than that of the convecting mantle, $\eta_i$, and the ratio of these two viscosities is denoted here by $\Delta\eta_L$. The stress required to bend a plate is proportional to lithospheric viscosity as well as to bending strain rate (Conrad & Hager 1999):

$$\tau_b \sim \eta_L \frac{vh}{R_{\text{eff}}^2} \propto \eta_L \frac{v^{1/2}}{R_{\text{eff}}^2}, \tag{3}$$

where $v$ is the plate velocity, $h$ is the plate thickness, and $R_{\text{eff}}$ is the effective bending curvature. [The notion of the "effective" bending curvature is introduced here to signify that, when realistic mantle rheology is considered, the bending stress is not simply proportional to the inverse square of geometrical bending curvature (Rose & Korenaga 2011).] The last proportionality holds because a plate grows by thermal diffusion, so its thickness is proportional to the square root of age, which is in turn inversely proportional to $\sqrt{v}$. The convective stress due to the negative buoyancy of a sinking plate scales with a temperature contrast across the plate, $\Delta T$, as

$$\tau_c \sim \alpha \rho g D \Delta T, \tag{4}$$

where $\alpha$ is the thermal expansivity and $D$ is the mantle depth. The comparison of these two stress estimates is facilitated by nondimensionalizing them first, using the internal stress scale $\tau_i = \eta_i \kappa / D^2$, where $\kappa$ is the thermal diffusivity. The scaling for nondimensional bending stress is given by

$$\tau_b^* \propto \Delta\eta_L^{2/3} \text{Ra}_i^{1/3} \left( \frac{D}{R_{\text{eff}}} \right)^2, \tag{5}$$

where $\text{Ra}_i$ is the internal Rayleigh number defined as

$$\text{Ra}_i = \frac{\alpha \rho g \Delta T\, D^3}{\kappa \eta_i}, \tag{6}$$

and the following scaling for plate velocity is used (Korenaga 2010):

$$v \propto \left( \frac{\kappa}{D} \right) \text{Ra}_i^{2/3} \Delta\eta_L^{-2/3}. \tag{7}$$

The nondimensional convective stress is simply

$$\tau_c^* \sim \text{Ra}_i. \tag{8}$$

To sustain plate tectonics, the convective stress has to be high enough to overcome the bending stress, i.e., $\tau_c^* > \tau_b^*$. Compared with the bending stress, the convective stress increases more rapidly with $\text{Ra}_i$, so for a given $\Delta\eta_L$, satisfying the above condition by raising $\text{Ra}_i$ is always possible. That is, plate tectonics becomes more feasible at higher Rayleigh numbers. A similar conclusion can be derived from the scaling analysis of Solomatov (2004), who focused on stresses in stagnant lid convection.

The above scaling argument is also supported by recent numerical simulations (Korenaga 2010), which suggest the following scaling for the critical Rayleigh number:

$$\text{Ra}_{i,\text{crit}} \sim 16 \Delta\eta_L^2. \tag{9}$$

Equivalently, for a given Rayleigh number, the critical viscosity contrast, above which plate tectonics is inhibited, scales as

$$\Delta\eta_{L,\text{crit}} \sim \frac{1}{4} \text{Ra}_i^{1/2}, \tag{10}$$

as shown in **Figure 3**. These scalings imply that the effective bending curvature is weakly sensitive to the internal Rayleigh number as $(R_{\text{eff}}/D) \propto \text{Ra}_i^{-1/6}$. The important feature of the numerical work of Korenaga (2010) is the use of realistic temperature-dependent viscosity. As seen in Equation (1), the temperature dependency is characterized by the activation energy $E$, or its nondimensionalized

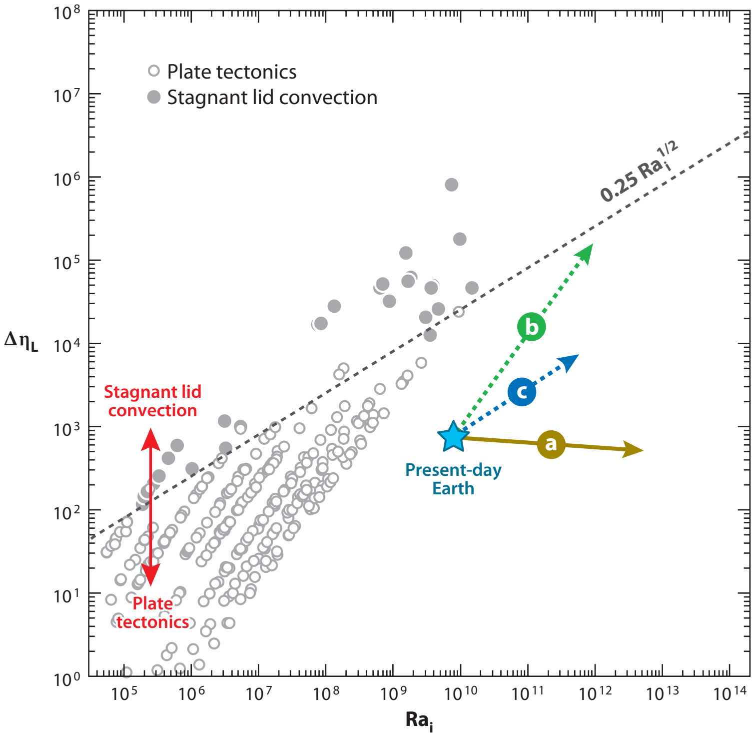


**Figure 3**

Regime diagram for plate tectonics and stagnant lid convection, in the parameter space of internal Rayleigh number $Ra_i$ and effective viscosity contrast across oceanic lithosphere $\Delta\eta_L$. Symbols denote numerical results from Korenaga (2010), with open and solid circles for plate tectonics and stagnant lid convection, respectively. The dashed line represents an approximate boundary between these two modes of convection [Equation (10)]. Also shown are, in a schematic manner, the likely location of present-day Earth and its possible historical paths: simple evolution (arrow *a*), evolution with mantle melting (arrow *b*), and evolution with mantle melting and with interaction with oceans (arrow *c*). Arrows point to the past from the present.

form

$$\theta = \frac{E\Delta T}{R(T_s + \Delta T)^2}, \tag{11}$$

which is known as the Frank-Kamenetskii parameter. Here $T_s$ denotes the surface temperature. The strongly temperature-dependent viscosity of silicate rocks corresponds to $\theta \sim 20$, but most early numerical studies used much lower values (typically up to ~7, when evaluated for the top thermal boundary layer) (e.g., Moresi & Solomatov 1998, Lenardic et al. 2004, Stein et al. 2004). Even with such a low $\theta$ value, it is possible to explore different modes of convection by varying $\mu_{eff}$, but having an Earth-like $\theta$ makes theoretical conjectures more geologically relevant. For the simulation results shown in **Figure 3**, $\theta$ ranges from ~10 to ~30. The systematic exploration of the parameter space with varying $Ra_i$ and $\Delta\eta_L$ allows us to extrapolate to Earth-like conditions at high Rayleigh numbers. On the basis of these simulation results, Korenaga (2010) also derived the following relation for the effective viscosity contrast across the lithosphere:

$$\Delta\eta_L \sim \exp(0.327\gamma^{0.647}\theta), \tag{12}$$

where $\gamma$ is the normalized friction coefficient defined as

$$\gamma = \frac{\mu_{\mathrm{eff}}}{\alpha \Delta T}. \tag{13}$$

Because plate tectonics is currently taking place on Earth, the present-day state should be found somewhere within the regime of plate tectonics in the parameter space of $\mathrm{Ra_i}$ and $\Delta\eta_{\mathrm{L}}$, and the above scaling argument implies that Earth may always have been in the regime of plate tectonics. Earth's mantle was generally hotter in the past than at the present, so $\Delta T$ was greater, and $\eta_i$ was smaller, making $\mathrm{Ra_i}$ higher in the past [Equation (6)]. Both the Frank-Kamenetskii parameter $\theta$ and the normalized friction coefficient $\gamma$ decrease (slightly) with increasing $\Delta T$ [Equations (11) and (13)], so $\Delta\eta_{\mathrm{L}}$ should have been lower for a given $E$ and $\mu_{\mathrm{eff}}$ [Equation (12)]. It then follows that plate tectonics may have been even more likely in the past (**Figure 3**, arrow *a*), suggesting the continuous operation of plate tectonics throughout Earth's history.

This argument may seem too simplistic, and the next section shows that such a simple scenario is insufficient to explain the thermal evolution of Earth. Earth's mantle is not a simple fluid, and some of its realistic complications cannot be ignored. The essence of the above scaling argument will, however, remain valid, and a good understanding of physical scaling is important in identifying conflicting arguments in the existing literature. O'Neill et al. (2007), for example, suggest that plate tectonics may have been less likely in the Precambrian because lower viscosity in a hotter mantle gives rise to lower convective stress in the case of stagnant lid convection. More important, however, is how the critical yield stress of lithosphere, above which plate tectonics does not occur, scales with the mantle temperature. The work of Solomatov (2004) suggests that the stress scales as (from his equation 30)

$$\tau_{\mathrm{crit,SL}} \propto \frac{\alpha \rho g D \Delta T}{\theta^2}. \tag{14}$$

Thus, the regime of plate tectonics should expand with increasing $\Delta T$, and this expansion is similar to that indicated in **Figure 3**. Unfortunately, the reason for the discrepancy between Solomatov (2004) and O'Neill et al. (2007) is hard to identify because the description of numerical modeling in the latter lacks some fundamental details.

The possibility of plate tectonics in the past is sometimes discounted on the basis of the chemical buoyancy of oceanic lithosphere (e.g., Davies 1992). Understanding the physics behind the thermal evolution of Earth is essential for understanding the buoyancy issue, so buoyancy will be discussed after the next section.

## 2.3. Energy Balance and Thermal Evolution

The appearance of plate tectonics and its subsequent evolution has a first-order impact on the cooling history of Earth, and by the same token, the history of plate tectonics should satisfy energetic constraints from the thermal evolution of Earth. A review on the theoretical foundation of thermal evolution is available in Korenaga (2008b), so only a brief summary along with some illustrative examples is given here.

The thermal history of Earth is controlled mostly by a balance between internal heating by radioactive elements in the mantle, $H$, and surface heat loss by mantle convection, $Q$, as

$$C \frac{dT_{\mathrm{i}}}{dt} = H(t) - Q(t), \tag{15}$$

where $C$ is the heat capacity of the whole Earth, $T_{\mathrm{i}}$ is the average mantle temperature, and $t$ is the time. The effect of core heat flux is implicit in this formulation (Korenaga 2008b, section 3.5), and the evolution of $T_{\mathrm{i}}$ can be approximated by that of mantle potential temperature $T_{\mathrm{p}}$, which is the temperature expected at the surface after correcting for adiabatic cooling. Once the

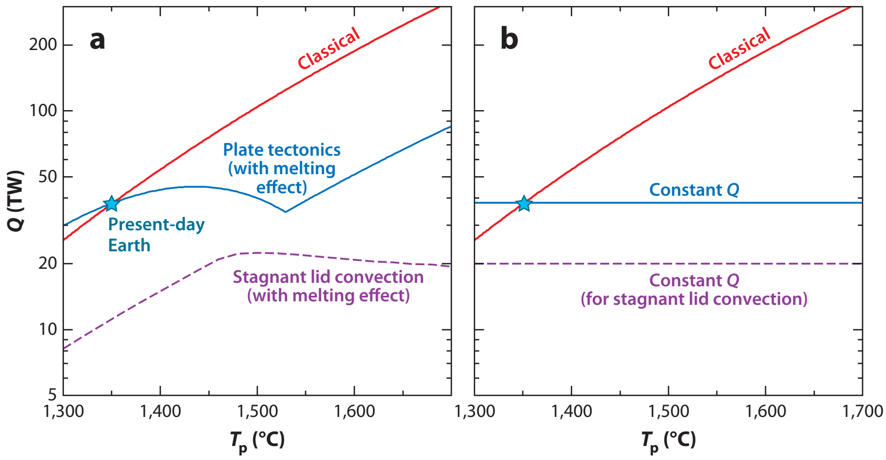


**Figure 4**

Some representative heat-flow scaling laws for mantle convection. (*a*) Solid curves denote calculations based on the scaling of plate tectonics with pseudoplastic rheology (Korenaga 2010), with the activation energy $E$ of 300 kJ mol$^{-1}$, the reference viscosity of $10^{19}$ Pa s at 1,350°C, and the viscosity contrast due to dehydration stiffening of 1 (i.e., no melting effect; *red*) and 100 (*blue*). The effective friction coefficient $\mu_{\text{eff}}$ is set to 0.025 (*red*) and 0.02 (*blue*), respectively, to reproduce the present-day mantle heat flux of 38 terawatts (TW) (Korenaga 2008b) at the present-day $T_p$ of 1,350°C (Herzberg et al. 2007). The red curve corresponds to classical scaling used in previous studies (e.g., Schubert et al. 1980). The dashed purple curve represents the scaling for stagnant lid convection with the effect of mantle melting (Korenaga 2009), using the same parameters assumed for the blue curve except for $\mu_{\text{eff}}$. (*b*) Simplified scaling used for examples shown in **Figures 5–7**.

present-day heat production $H(0)$ is given, calculating the past values of $H(t)$ is easy because the relative abundance of radioactive elements and their half-lives are known. Even for the same $H(t)$, however, vastly different thermal histories can result from different assumptions about $Q(t)$. As the mantle heat flux is generally parameterized as a function of mantle potential temperature, the following discussion focuses on the functionality of $Q(T_p)$.

It is usually thought that the mantle heat flux should increase with $T_p$ (e.g., Schubert et al. 2001) because more vigorous convection is expected for a hotter mantle characterized by lower viscosity (**Figure 4*a***, classical scaling). The viscosity of the mantle, however, depends not only on temperature but also on other factors such as grain size and water content (e.g., Karato & Wu 1993, Hirth & Kohlstedt 2003), and under certain conditions, a hotter mantle may become more viscous (e.g., Solomatov 1996, Korenaga 2003) (**Figure 4*a***, plate tectonics scaling). To illustrate how such a difference in heat-flow scaling affects thermal evolution, I first consider the consequence of constant heat-flux scaling (**Figure 4*b***), with different assumptions about the present-day internal heating $H(0)$. The simplicity of this temperature-independent scaling helps clarify the relation between heat balance and thermal history. The amount of internal heating is often expressed in terms of its relative contribution to mantle heat flux as $Ur(t) \equiv H(t)/Q(t)$, which is termed the (convective) Urey ratio, and two values (0.34 and 0.84) are tested for the present-day Urey ratio $Ur(0)$ (**Figure 5**). Whereas the thermal evolution with the low Urey ratio is geologically reasonable, that with the high Urey ratio results in too cold of a mantle in the past. Both thermal histories are concave downward because, going backward in time, internal heating eventually exceeds mantle heat flux, so thermal evolution has to switch from cooling down ($dT_i/dt < 0$) to warming up ($dT_i/dt > 0$). The timing of this transition is controlled by the amount

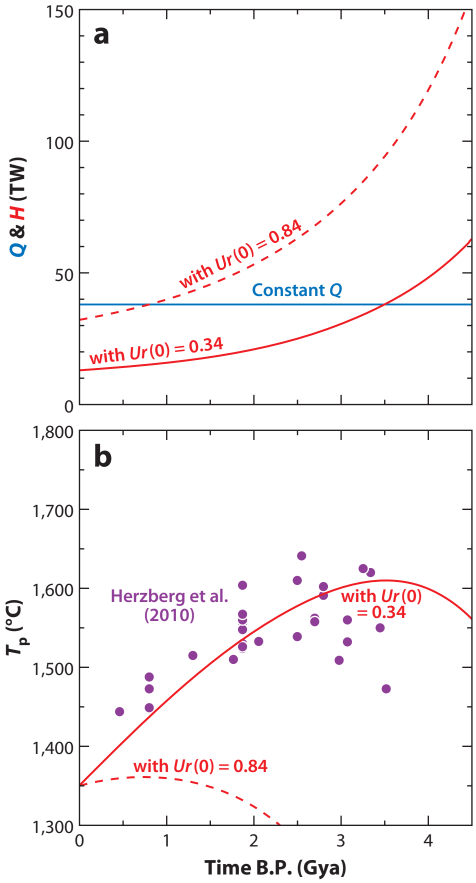


**Figure 5**

Thermal evolution modeling with constant heat flux. Equation (15) is integrated backward in time, starting with the present-day temperature of 1,350°C (Herzberg et al. 2007) and with the heat capacity $C$ of $7 \times 10^{27}$ J $K^{-1}$ (Stacey 1981). No effect of continental growth is considered in this simple example. (*a*) Mantle heat flux $Q$ (*blue*) and mantle heat production $H$ (*red*). Low and high heat production scenarios are tested using the present-day Urey ratios of 0.34 (*solid*) and 0.84 (*dashed*), respectively. (*b*) Corresponding predictions for thermal history. Solid circles denote petrological estimates on past potential temperature (Herzberg et al. 2010). Abbreviation: B.P., before present.



of internal heating, and to have a thermal maximum at ~3 Gya as suggested by the petrological estimate of Herzberg et al. (2010), the present-day Urey ratio has to be in the vicinity of ~0.3.

Let us consider next the case of the classical heat-flow scaling (**Figure 6**). Because the mantle heat flux increases with temperature in this case, a crossover between heat flux and heat production does not take place even for the high Urey ratio. On the contrary, the present-day difference between heat output and input is amplified in the past; such a difference has to be compensated for by the secular cooling of Earth, which means a hotter mantle and a higher heat flux in the past. This increase in heat flux is more rapid than that in heat production, leading to a positive feedback. The highly divergent thermal history seen for the low Urey ratio is known as thermal catastrophe (Christensen 1985), and to avoid such an unrealistic solution, researchers have assumed a high Urey ratio in the modeling of thermal evolution (e.g., Davies 1980, Schubert et al. 1980,

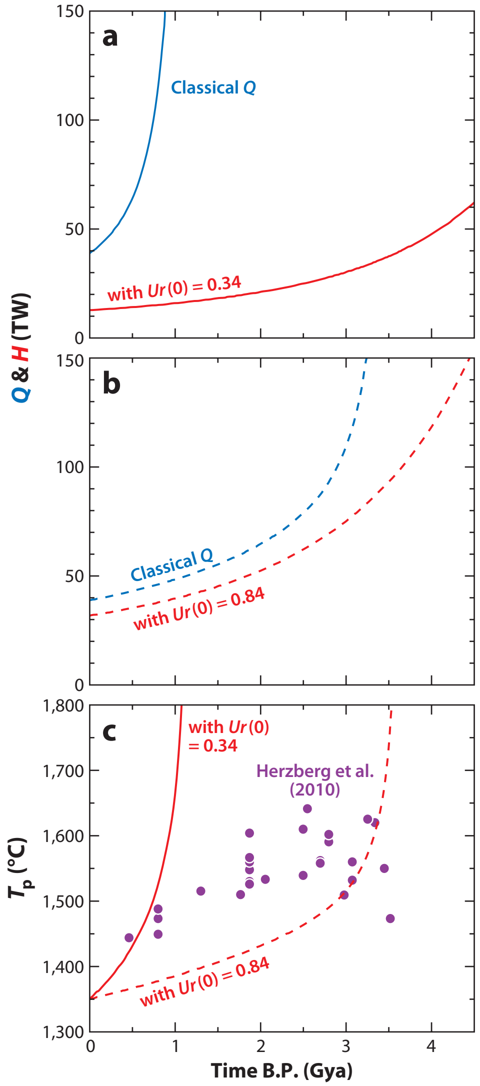


**Figure 6**

Similar to the thermal evolution modeling shown in **Figure 5** but with the classical heat-flow scaling shown in **Figure 4**. (*a*) Evolution of mantle heat flux $Q$ (*blue*) and mantle heat production $H$ (*red*) with low present-day Urey ratio (0.34). (*b*) Same as panel *a* but with high Urey ratio (0.84). (*c*) Corresponding thermal histories. Solid circles denote petrological estimates on past potential temperature (Herzberg et al. 2010). Abbreviation: B.P., before present.

McGovern & Schubert 1989, Franck & Bounama 1997, Davies 2009). There are, however, at least two problems with this assumption. First, internal heat production results from the decay of radioactive elements, the budget of which is constrained by the chemical composition of Earth. Composition models for Earth's mantle suggest that the convective Urey ratio at the present should be ~0.3 and cannot be as high as 0.8 (e.g., McDonough & Sun 1995; Lyubetskaya & Korenaga 2007a,b). Second, even though the high Urey ratio suppresses a catastrophic solution,

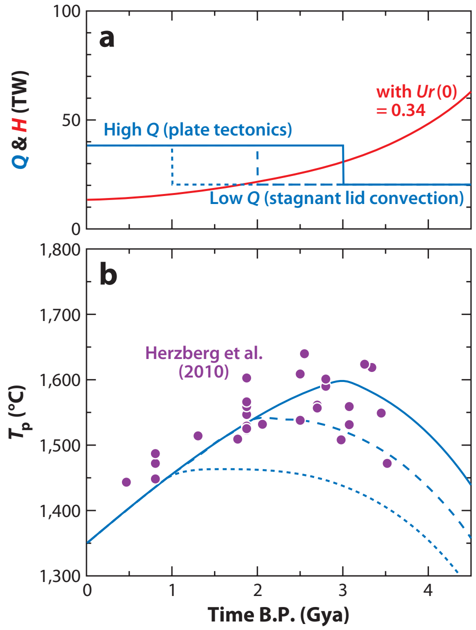


**Figure 7**

Similar to the thermal evolution modeling shown in **Figure 5**, but with a switch in heat-flow scaling from plate tectonics to stagnant lid convection at 1 Gya (*dotted*), 2 Gya (*dashed*), and 3 Gya (*solid*). The present-day Urey ratio is assumed to be 0.34 in this example. (*a*) Mantle heat flux $Q$ (*blue*) and mantle heat production $H$ (*red*). (*b*) Corresponding thermal histories. Solid circles denote petrological estimates on past potential temperature (Herzberg et al. 2010). Abbreviation: B.P., before present.

the resulting thermal history is still concave upward, which is opposite to what the petrological estimate indicates (**Figure 6*c***).

It thus appears that heat-flow scaling for mantle convection has to be much less sensitive to temperature than the classical scaling indicates, although the exact cause of such nonclassical behavior is still debated. Three possibilities have been proposed: (*a*) the grain-size-sensitive rheology of the lower mantle (Solomatov 1996, 2001), (*b*) dehydration stiffening upon mantle melting in the upper mantle (Korenaga 2003, 2006), and (*c*) the gradual hydration of the whole mantle (Korenaga 2008a). As experimental constraints on lower-mantle rheology are still scarce, assessing the plausibility of the first mechanism is difficult. I therefore summarize recent developments regarding the last two. Before doing so, I note for the reader that the continuous operation of plate tectonics throughout Earth's history is assumed when only one heat-flow scaling is employed (as done in **Figures 5** and **6** as well as in most previous studies on thermal evolution). However common, this is a bold assumption, so it may be worth exploring the outcome of initiating plate tectonics in the middle of Earth's history. One simple example is given in **Figure 7**. Switching to a relatively low heat flux expected for stagnant lid convection (**Figure 4**) shifts the timing of a crossover between heat flux and heat production and thus that of a thermal maximum. Whereas the onset of plate tectonics at 1 Gya results in too cold of a thermal history, the onset at 2–3 Gya seems to satisfy the petrological constraints on thermal history. Nonclassical scaling laws shown in **Figure 4*a*** are, however, just an example, and there are a range of possibilities for both plate tectonics and stagnant lid convection when the effects of mantle melting are considered (Korenaga 2009, 2010). Trying to narrow down the onset time of plate tectonics by hypothesizing a transition in the mode of convection would then become an arbitrary exercise, if not guided by some physical reasoning for such a transition.

When modeling the thermal evolution of Earth, therefore, it is important to monitor the plausibility of plate tectonics at the same time, and such an attempt was recently made during exploration of the aforementioned ideas of dehydration stiffening and mantle hydration (Korenaga 2011). How dehydration stiffening affects heat-flow scaling is the following (Korenaga 2003). When the mantle undergoes partial melting beneath mid-ocean ridges, a trace amount of water in the source mantle is nearly completely partitioned into the melt phase, and this dehydration results in a compositionally stiff, depleted mantle lithosphere (Hirth & Kohlstedt 1996) because the rheology of the mantle is sensitive to its water content (e.g., Karato et al. 1986; Mei & Kohlstedt 2000a,b). A hotter mantle in the past is likely to have experienced a greater degree of partial melting by starting to melt deeper beneath mid-ocean ridges (McKenzie & Bickle 1988, Langmuir et al. 1992), thus resulting in a thicker dehydrated lithosphere. A thicker lithosphere is more difficult to bend at a subduction zone [Equation (3)], thereby slowing down plate tectonics when the mantle was hotter than at the present. Although early attempts to quantify this hypothesis (e.g., Korenaga 2006) were based on an approximate energy balance known as the boundary layer theory, a new heat-flow scaling law based on fully dynamic convection models is now available (Korenaga 2010). Using the new scaling, Korenaga (2011) reevaluated the effect of dehydration stiffening on thermal evolution (**Figure 8**, closed-system evolution). Whereas modeling with the effect of dehydration stiffening can satisfy the petrological constraint on the thermal history (**Figure 8*a***), the effective viscosity contrast across oceanic lithosphere can approach the threshold for plate tectonics during the Proterozoic (**Figure 8*b***). That is, the thickening of dehydrated lithosphere by deeper melting in the past could have been too effective to slow down plate tectonics, jeopardizing the operation of plate tectonics itself. Thermal evolution with dehydration stiffening is thus at the verge of self-consistency.

In the mantle hydration hypothesis of Korenaga (2008a), the mantle is also assumed to be gradually hydrated by subduction; thus, the mantle in the past is not only hotter but also drier than at the present. A drier mantle lowers the viscosity contrast across oceanic lithosphere. A hotter mantle still creates a thicker dehydrated lithosphere, but the relative viscosity difference between lithosphere and asthenosphere is reduced because the latter is not as hydrated as at the present. This idea was also tested with the new heat-flow scaling (Korenaga 2011); the results are summarized in **Figure 8** (open-system evolution). Gradual mantle hydration does help guarantee the operation of plate tectonics throughout Earth's history (**Figure 8*b***).

A major difference between these two types of thermal evolution is whether the mantle is an open system regarding water transport, and this difference leads to considerably different predictions for the secular evolution of surface environment. In the original dehydration stiffening model, the net water content of the mantle does not change with time; even though part of the mantle dehydrates during mantle melting, the extracted water is assumed to return to the mantle eventually. This is why this model is referred to as a closed system in **Figure 8**. The volume of oceans stays constant in this case, and to maintain the constancy of continental freeboard (i.e., the mean sea level has always been close to the mean continental level) (Wise 1974), the zero-age seafloor depth must have been much shallower in the past (**Figure 8*c***) because more sluggish plate tectonics results in older and thus deeper seafloor on average (e.g., Parsons 1982). In contrast, the mantle hydration model requires a greater ocean volume in the past, and the zero-age seafloor depth does not have to deviate from the present-day value (**Figure 8*c***). At the moment, it is difficult to assess which scenario is more likely, but these results illuminate intrinsic connections among thermal evolution, plate tectonics, and surface environment. This issue is discussed below in more detail (see Section 4.1).

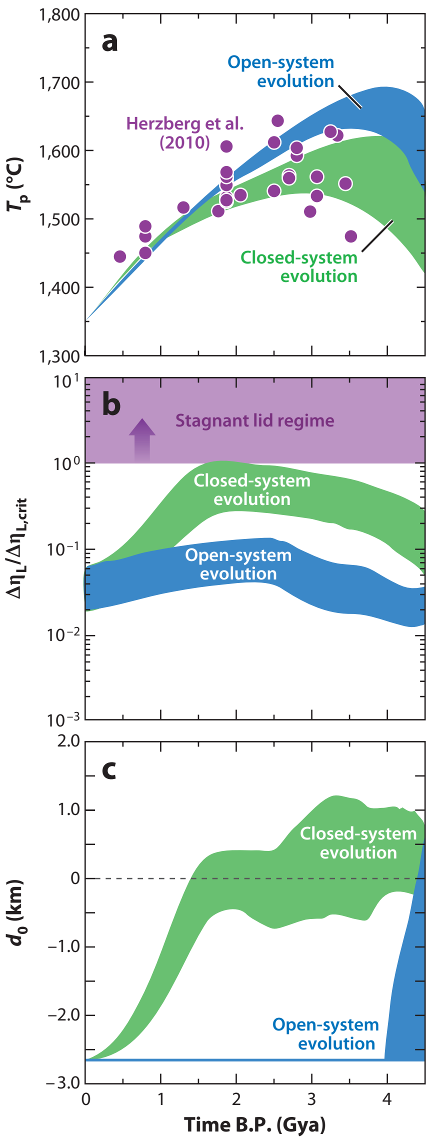


**Figure 8**

Summary of thermal evolution modeling with the new scaling of plate tectonics (Korenaga 2011). Green and blue shadings correspond to, respectively, the dehydration stiffening hypothesis (closed-system evolution) and the mantle hydration hypothesis (open-system evolution). Various uncertainties in model parameters are explored by Monte Carlo sampling, and these shadings cover from the first to third quartiles, representing 50% of total solutions. (*a*) Reconstructed thermal histories. Circles denote petrological estimates by Herzberg et al. (2010). (*b*) Effective viscosity contrast across oceanic lithosphere, normalized by the critical viscosity contrast for the operation of plate tectonics. (*c*) Secular evolution of zero-age seafloor depth based on the constancy of continental freeboard. For the validity of various assumptions behind the freeboard argument, see Korenaga (2008a, 2011).

## 2.4. Buoyancy of Oceanic Lithosphere

Oceanic lithosphere is composed of oceanic crust and depleted mantle lithosphere, both of which are the products of mantle melting and are chemically less dense than the asthenospheric mantle (**Figure 9*a***). Once oceanic lithosphere is subducted, such chemical buoyancy quickly diminishes at the depth of ~60 km because of basalt-eclogite transition (Ringwood & Green 1964), but for subduction to be initiated, it must be compensated for by density increase due to thermal contraction. For oceanic lithosphere to sink, therefore, it has to be cooled sufficiently, and because a hotter mantle produces a thicker lithosphere (**Figure 9*b***), the timescale for required cooling would have been longer in the past (**Figure 9*c***). On the basis of this notion, Davies (1992) questioned the plausibility of plate tectonics in the Precambrian because sufficient cooling cannot have been achieved by rapid plate tectonics in the past. This argument is, however, based on the classical heat-flow scaling (**Figure 4**). As summarized in the section above, the tempo of plate tectonics in the past should not differ greatly from that at the present, to satisfy the petrological estimates on thermal evolution and the geochemical budget of radioactive elements (**Figure 5**). Furthermore, geological data support more sluggish plate tectonics in the past (Bradley 2008).

Whether oceanic lithosphere can become negatively buoyant depends primarily on plate velocity. Mantle heat flux and plate velocity are in general related by the physics of thermal diffusion as (Turcotte & Schubert 1982)

$$Q \propto \Delta T \sqrt{v}. \tag{16}$$

Therefore, in the case of constant heat-flow scaling, the maximum age of oceanic lithosphere should be greater in the past than at the present as

$$\frac{t_{\max}(t)}{t_{\max}(0)} \sim \frac{v(0)}{v(t)} \sim \left(\frac{\Delta T\,(t)}{\Delta T\,(0)}\right)^2. \tag{17}$$

Under the assumption of constant heat flux, this scaling is valid irrespective of plate size evolution; if plates were smaller in the past, plate velocity would have had to decrease to maintain the constant heat flux. The buoyancy constraint on the likelihood of Precambrian plate tectonics, therefore, does not seem to be as stringent as previously thought (**Figure 9*c***).

There have been other arguments regarding the buoyancy of oceanic lithosphere. Davies (2006), for example, suggested that plate tectonics in the early Earth may have been possible if the mantle were considerably depleted in the past because such mantle would not have produced thick crust even at elevated temperatures. This idea, however, requires some (unknown) mechanism to refertilize such highly depleted mantle so that the melting of the present-day mantle could successfully produce ~6-km-thick normal oceanic crust. Perhaps more interesting is the suggestion, based on detailed mineral physics calculations, that even mature oceanic lithosphere at the present does not become negatively buoyant (Hynes 2005, Afonso et al. 2007). Because plate tectonics is nonetheless taking place, these authors further suggest that the net buoyancy of oceanic lithosphere may not be important for the initiation of subduction. From a dynamical point of view, this is a perplexing suggestion, so further consideration is warranted. It is also important to validate the predictions for the density of oceanic lithosphere at higher mantle temperatures. Crustal density shown in **Figure 9*a***, for example, is based on the parameterization of Korenaga et al. (2002), which assumes low-pressure crystallization. The internal construction of thick oceanic crust is a difficult problem, but future studies on anomalously thick oceanic crust in large igneous provinces may provide useful observational hints.

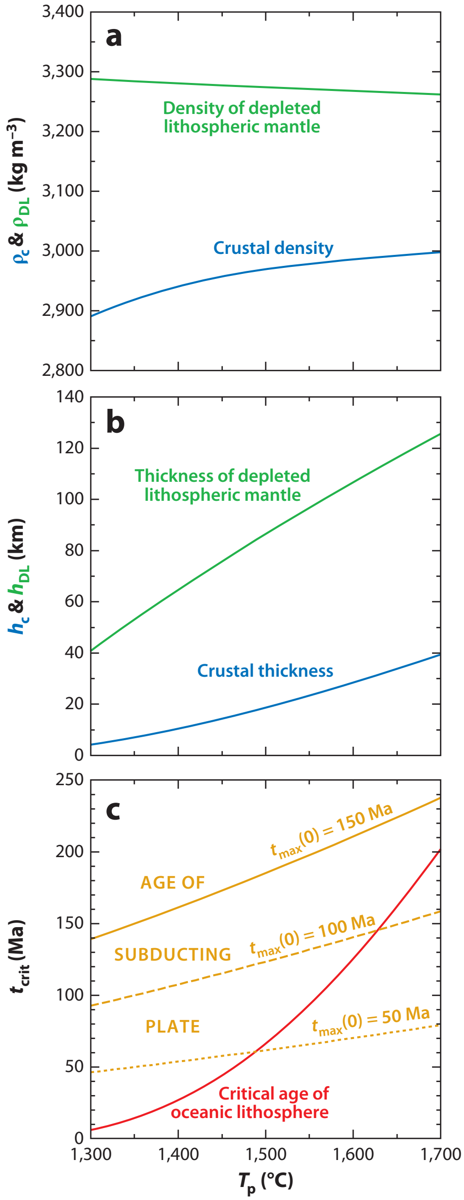


**Figure 9**

Chemical buoyancy of oceanic lithosphere as a function of mantle potential temperature. (*a*) Density of oceanic crust (*blue*) and depleted mantle (*green*). (*b*) Thickness of oceanic crust (*blue*) and depleted mantle (*green*). (*c*) Critical plate age to achieve neutral buoyancy (*red*). Orange curves denote the maximum plate age prediction for different present-day values: 50 Ma old (*dotted*), 100 Ma old (*dashed*), and 150 Ma old (*solid*). After Korenaga (2006).

## 2.5. Could Plate Tectonics Have Started in the Hadean?

It should be clear now that the likelihood of plate tectonics must be considered in the framework of thermal evolution. In terms of the regime diagram for mantle convection spanned by the internal Rayleigh number $Ra_i$ and the lithospheric viscosity contrast $\Delta\eta_L$ (**Figure 3**), a hotter mantle is expected to have a higher Rayleigh number and a lower viscosity contrast (**Figure 3**, arrow *a*), so the operation of plate tectonics appears to be guaranteed in the past. Such a simple evolution, however, corresponds to the thermal catastrophe solution (**Figure 6**) unless mantle heat production is considerably higher than what geochemistry indicates. To reproduce a more moderate thermal history consistent with available petrological constraints, the convective vigor of a hotter mantle has to be somehow reduced. The dehydration stiffening hypothesis provides a physical mechanism to achieve such a reduction by raising $\Delta\eta_L$ for a hotter mantle, although this increase in the viscosity contrast could also jeopardize the operation of plate tectonics (**Figure 3**, arrow *b*). The mantle hydration hypothesis alleviates the situation by making a hotter mantle in the past drier than that at the present, to keep the lithospheric viscosity contrast away from the threshold value for plate tectonics (**Figure 3**, arrow *c*). Both hypotheses take into account the effect of mantle melting (i.e., dehydration stiffening) on mantle heat flux; the only difference is that the mantle hydration hypothesis considers how the open-system behavior of the mantle could further influence its convective vigor. The likelihood of plate tectonics in the early Earth, therefore, may depend on the net outcome of ocean-mantle interaction over Earth's history.

At the same time, it is important to remember that these theoretical conjectures are based on several simplifying assumptions, as discussed more fully below (see Section 4.2). Earth is a complex system, and our understanding of its physical and chemical properties is still insufficient to confidently simulate most of what we want to know from first principles alone. A good example is readily found in the ongoing debate on why plate tectonics happens on Earth (see Section 2.1). Theoretical developments often benefit from observational insights, and the proper interpretation of observations requires a good understanding of a relevant theory. Scientific progress relies on continual feedback between theories and observations, and I thus turn to observational constraints on the history of plate tectonics.

# 3. OBSERVATIONAL CONSTRAINTS

## 3.1. Plate Tectonics Indicators

The most prominent feature of plate tectonics is the continuous creation of oceanic lithosphere at mid-ocean ridges and its destruction at subduction zones (**Figure 2*b***). As the primary product of plate tectonics is ceaselessly recycled into the mantle, inference on its possible operation in the distant past has to rely on other, more subsidiary products that could escape from recycling, such as ophiolites, arc assemblages, accretionary prisms, paired metamorphic belts, and orogens. Quite a few comprehensive reviews on these plate tectonics indicators have already been written by professional geologists (e.g., Condie & Pease 2008), so I summarize notable observations only briefly and focus more on the inherent limitations of these geological data.

Continental masses on Earth appear to have a tendency to occasionally assemble and form a supercontinent; the latest instance is the famous Pangea, which existed for the period of ~0.2–0.3 Gya, and the record of globally synchronous orogens suggests that four more supercontinents may have existed in Earth's history (Hoffman 1997) (**Figure 10**), although their specifics are still debated (e.g., Bleeker 2003, Bradley 2011). The formation of a supercontinent (or even a quasi-supercontinent) requires the closure of multiple ocean basins that existed among different continental masses, thereby indicating the subduction of oceanic lithosphere on a global scale.

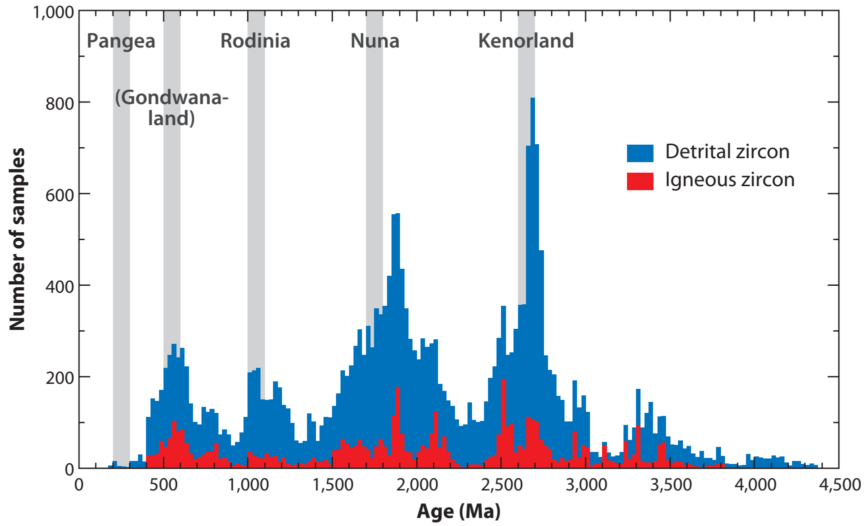


**Figure 10**

Distribution of U/Pb zircon ages in orogenic granitoids (*red*) and detrital zircon (*blue*) data sets compiled by Condie et al. (2009a). Also shown are the periods of supercontinent assemblies according to Hoffman (1997). The term Gondwanaland appears in parentheses because it may be better viewed as a building block of the more complete Pangea. Major peaks of zircon age distribution correlate well with the supercontinental cycle, reflecting high crustal production rates (e.g., Condie 1998) or high preservation rates during supercontinent assemblies (e.g., Hawkesworth et al. 2010).

As the oldest supercontinent Kenorland was assembled at ∼2.7 Gya (Hoffman 1997), it seems reasonable to assume that plate tectonics was already operational in the late Archean.

Geological indicators for plate tectonics at >2.7 Gya also exist. Eclogite xenoliths aged approximately 3 billion years (∼3 Ga) that were found in the Kaapvaal craton, for example, are suggested to have originated from subducted oceanic crust because their oxygen isotopic signature points to seawater alteration in the hydrothermal system (e.g., Shirey et al. 2001). A case for continental rifting at ∼3.2 Gya has been suggested from the Pilbara craton (Van Kranendonk et al. 2007), and the Itsaq gneiss complex in southwest Greenland may preserve a ∼3.6-Ga-old suture zone (Nutman et al. 2002). Also located in southwest Greenland, the ∼3.8-Ga-old Isua supracrustal belt has been interpreted to contain the oldest accretionary complex (Komiya et al. 1999) as well as the oldest evidence of oceanic crustal accretion by spreading (Furnes et al. 2007). Geological interpretations, however, generally become more controversial for older rocks. The accretionary complex interpretation of the Isua belt, for example, has been challenged by subsequent studies (Fedo et al. 2001, Rollinson 2002), and a similar interpretation suggested for the ∼3.4-Ga-old North Pole area of the Pilbara craton (e.g., Kitajima et al. 2001) is also disputed (e.g., Van Kranendonk et al. 2007).

Irrespective of whether or not such an interpretation of a particular geological province is correct, one difficulty common to these arguments—either for or against the operation of plate tectonics—is that they have to address a global-scale problem on the basis of highly regional evidence. As mentioned in Section 1, older geological samples are more geographically limited, and the situation is severe for the Archean. Observations with limited spatial extents may be explained equally well by other processes unrelated to plate tectonics, and conjectures abound for what could happen in the pre-plate-tectonics era (e.g., Davies 1992, Stern 2008). Paleomagnetism

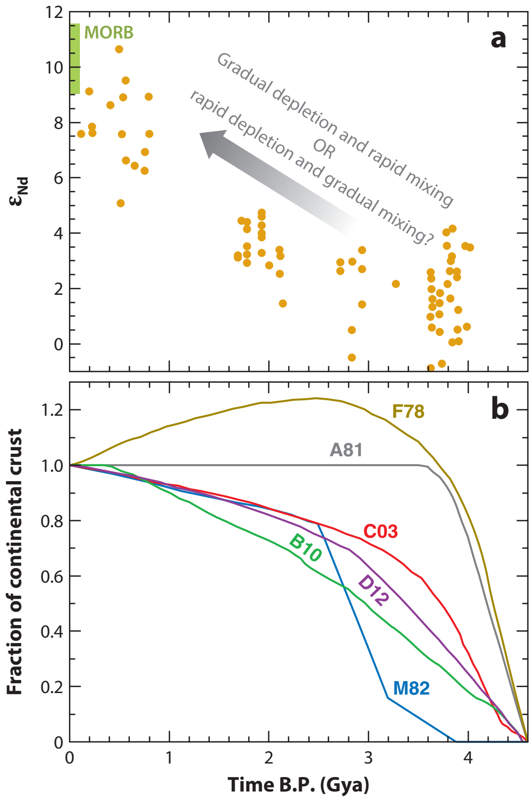


**Figure 11**

(*a*) $^{143}$Nd evolution of the depleted mantle as defined by juvenile granites and Phanerozoic ophiolites (Bennett 2003). $\varepsilon_{Nd}$ measures a deviation from the reference $^{143}$Nd evolution defined by the chondritic uniform reservoir (Jacobsen & Wasserburg 1980), which may be considered equivalent to that of the primitive, undifferentiated mantle. The green box denotes a range for present-day mid-ocean-ridge basalts (MORB). The secular trend of $\varepsilon_{Nd}$ may be explained by gradual depletion of the convecting mantle (and instant mixing in the mantle) or more rapid depletion of the mantle (and more gradual mixing). (*b*) Examples of continental growth models: F78 (Fyfe 1978), A81 (Armstrong 1981), M82 (McLennan & Taylor 1982), C03 (Campbell 2003), B10 (Belousova et al. 2010), and D12 (Dhuime et al. 2012). Abbreviation: B.P., before present.



can provide an independent check on the possibility of ancient plate tectonics by reconstructing the lateral motion of tectonic plates, but the current paucity of precisely dated paleomagnetic poles does not allow a conclusive statement beyond ~1.2 Gya (Evans & Pisarevsky 2008).

A more global perspective may be offered by the secular evolution of mantle geochemistry as recorded by igneous rocks (**Figure 11*a***). The evolution of the depleted mantle and the growth of continental crust are generally considered to be complementary to each other (e.g., Hofmann 1988), and whereas the genesis of continental crust may not necessarily demand the operation of modern-style plate tectonics, measurable changes in mantle chemistry should still serve as a proxy for processes on a substantial scale. Even though igneous samples in the distant past are geographically limited, their parental mantle is of a global nature. Continental growth models proposed in recent years are therefore commonly based on the isotope or trace-element evolution of the depleted mantle (e.g., Campbell 2003, Belousova et al. 2010, Dhuime et al. 2012), and they tend to show relatively gradual growth over Earth's history (**Figure 11*b***). A recent study by Dhuime et al. (2012) emphasizes a change in growth rate during the Archean (∼3 Gya) and suggests that it may reflect the onset of plate tectonics.

Continental growth in the early Earth is, however, still a controversial subject (e.g., Harrison 2009), and one issue that has commonly been overlooked is the efficiency of mantle mixing. When part of the mantle melts beneath mid-ocean ridges, for example, it differentiates into the crustal component and residual mantle component, the latter of which is also known as depleted mantle lithosphere (see Sections 2.3 and 2.4). This residual mantle is highly depleted and would not easily melt more even with additional heating because it has already lost most of its easily fusible components. If these differentiated components were to subduct together and remix completely, the original source mantle would be recovered, and if part of the crustal component were somehow to remain at the surface and contribute to continental growth, the result would be the overall chemical depletion of the convecting mantle, which is referred to as the evolution of the depleted mantle. For the extraction of the crustal component to be reflected in the average composition of the convecting mantle, however, the subducted components have to be sufficiently mixed with the rest of the mantle, and this mixing does not take place instantly. The mixing efficiency of mantle convection is often considered to be higher in the past because of more vigorous convection, but as shown in Section 2.3, this speculation is based on the traditional view on the thermal evolution of Earth (**Figure 6**), which fails to account for relevant geochemical and petrological data. In a more tenable scenario for thermal evolution, the vigor of mantle convection is predicted to have been more diminished in the past, so there could be a considerable lag time for continental growth to be reflected in the average mantle composition. Such a consideration would be important, for example, in the assessment of the validity of various model ages, which are calculated by assuming a homogeneous mantle. A long timescale for mantle mixing could undermine the traditional interpretation of geochemical data (e.g., Korenaga 2008b, section 2.2), but its significance does not seem to be widely appreciated yet.



## 3.2. Secular Evolution of Metamorphism

Peak temperature and pressure conditions recorded in regional metamorphic belts have long been known to exhibit a peculiar secular trend, i.e., the near absence of high-pressure metamorphism in the Precambrian (e.g., de Roever 1956, Ernst 1972, Maruyama et al. 1996) (**Figure 12**). Both de Roever (1956) and Ernst (1972) interpreted this trend as a signature of Earth's secular cooling, and Maruyama & Liou (1998) went further by proposing that global geothermal gradients may have rapidly dropped at the Proterozoic-Phanerozoic boundary. According to Stern (2005), this secular trend may even indicate the onset of modern-style plate tectonics in the Neoproterozoic era, whereas Brown (2006) cautioned that plate tectonics on a hotter Earth may have left a different imprint in the older rock record. In the following, I discuss the robustness of these suggestions from a geodynamical perspective.

First of all, the secular cooling interpretation (e.g., Maruyama & Liou 1998) is problematic for two reasons. Although heat production was surely higher in the past, its relation with surface heat flux (i.e., geothermal gradients) is not simple (see Section 2.3). Heat flux is a function of mantle temperature, and its relation to mantle heat production is through the global heat balance [Equation (15)]. A proportional increase in heat flux with heat production can be achieved only under a special circumstance (**Figure *6b***), which fails to explain both the thermal history and thermal budget of Earth. Earth has been cooling for at least the past 3 Ga (Herzberg et al. 2010), but its surface heat flux is controlled by the vigor of mantle convection, which is not necessarily higher for a hotter mantle (**Figure 4**). More important, a secular decline in geothermal gradients is not what the metamorphic data actually indicate because Phanerozoic data also exhibit

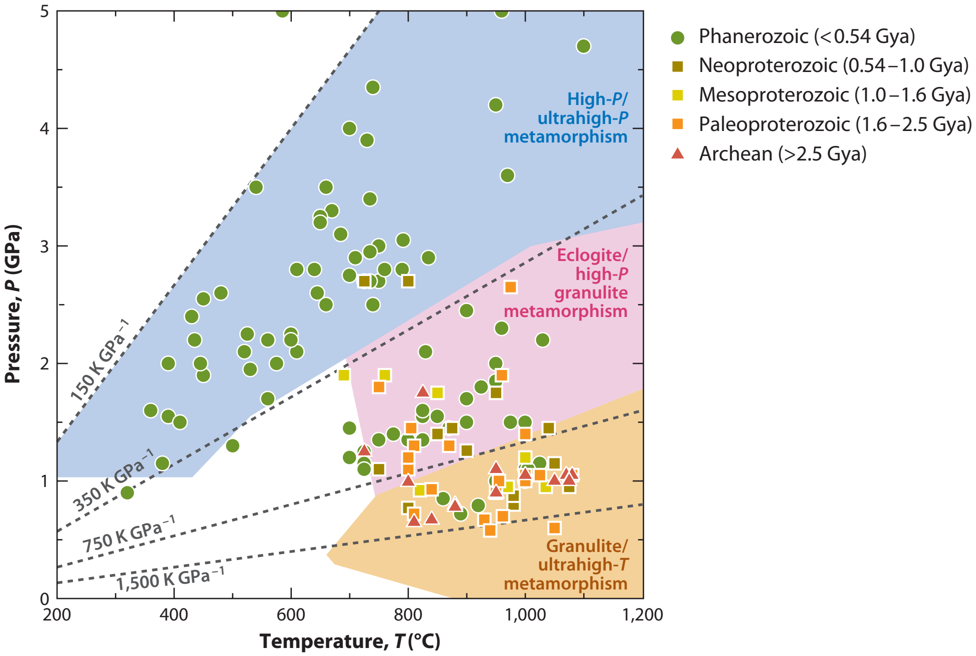


**Figure 12**

Peak pressure-temperature values for regional metamorphic belts in the world (Brown 2007). The ages of these metamorphic belts are indicated by different symbols (see legend). Background shadings correspond to three broad domains of metamorphism: high-pressure and ultrahigh-pressure metamorphism (*blue*), eclogite and high-pressure granulite metamorphism (*pink*), and granulite and ultrahigh-temperature metamorphism (*orange*). High-pressure metamorphism includes blueschist, amphibole-epidote eclogite facies, amphibole-lawsonite eclogite facies, lawsonite eclogite facies, and amphibole eclogite facies. These different domains correspond to different geothermal gradients (*dashed lines*).

low- and medium-pressure metamorphism (**Figure 12**). Such data are missing in the compilation made by Maruyama & Liou (1998), in which only Phanerozoic blueschists and eclogites (both belonging to high-pressure metamorphism) are compared with >1-Ga-old metamorphic rocks (see their figure 9). Miyashiro (1972) interpreted high-pressure metamorphism to result from the subduction zone environment and low-pressure metamorphism from the arc environment and backarc environment, and low-pressure metamorphism is widely distributed in Precambrian terranes as well. An indisputable secular trend in metamorphism is thus nothing more than the lack of high-pressure metamorphism prior to the Neoproterozoic.

The observation of high-pressure and ultrahigh-pressure metamorphism at the surface, however, has been a nagging issue in geodynamics. For such metamorphism to occur, crustal rocks have to be (*a*) brought down to great depths and metamorphosed under high pressure and temperature conditions, and then (*b*) exhumed rather quickly to retain peak metamorphic conditions. Van Hunen & van den Berg (2008) suggested that hot Archean slabs may have been too weak to drag down buoyant continental crust so that high-pressure metamorphism would not occur in the first place. Subducting slabs in the past are, however, likely to be strong not only because stiff dehydrated lithosphere was thicker (see Section 2.3) but also because subducting plates were older (see Section 2.4). The appearance of high-pressure metamorphism in the Neoproterozoic may instead imply some conditions for the exhumation stage, whose mechanism is, however, still widely debated (e.g., Davies & von Blanckenburg 1995, Pfiffner et al. 2000, Gerya et al. 2002, Boutelier et al. 2004, Warren et al. 2008). Husson et al. (2009) recently suggested that the transient behaviors of subduction such as slab rollback may be important for exhumation, and the dynamics of slab rollback is another target of active research (e.g., Funiciello et al. 2008, Stegman et al. 2010, Quinquis et al. 2011). Our understanding of subduction is still limited (e.g., Billen 2008). The subduction system is complex, involving a descending slab with oceanic crust and depleted mantle lithosphere, an overriding plate with continental (or oceanic) crust and lithospheric mantle, and subslab and wedge mantle components. The rheology of each component is a different function of, at least, temperature, pressure, stress, grain size, and water content, and the migration of water in the subduction environment is another hard problem. The rheology of individual components is not fully resolved, even in the case of the best-studied olivine aggregates (Korenaga & Karato 2008). As seen in the ongoing debate on the origin of plate tectonics on Earth (see Section 2.1), the detailed rheological evolution of oceanic lithosphere is still poorly understood, and even the subslab asthenosphere may not follow simple secular evolution because its water content could change with time (see Section 2.3). Much effort will be needed to predict the likely secular evolution of the subduction system with some confidence, and the proper interpretation of Phanerozoic high-pressure metamorphism may hinge on future developments in slab dynamics.



## 3.3. Possibility of Intermittent Plate Tectonics

Along with debates on when plate tectonics began on Earth and how it might have emerged from its predecessor, the possibility of intermittent plate tectonics, i.e., an intermediate situation between plate tectonics and stagnant lid convection, has also been suggested (e.g., O'Neill et al. 2007, Silver & Behn 2008, Condie et al. 2009b). Observational bases for these suggestions are of varying quality.

O'Neill et al. (2007), for example, presented paleomagnetic evidence for periods of rapid plate motions correlating with observed peaks in crustal age distribution. Such peaks in crustal age distribution are, however, usually ascribed to the periods of supercontinental assemblies (**Figure 10**), and it is impossible to unambiguously interpret paleomagnetic data from these periods in terms of plate motion because of the possibility of true polar wander (e.g., Evans 2003). Geochemical proxies used by Silver & Behn (2008) to estimate the history of subduction flux are more questionable. By assuming a correlation between subduction flux and mantle depletion, the authors derived one proxy by differentiating a polynomial fit to the continental growth curve of Collerson & Kamber (1999), which is based on the evolution of the Nb/Th ratio and is similar to that of McLennan & Taylor (1982) (M82 in **Figure 11*b***). This proxy is thus based on the subtlety of the growth curve, which is unsupported by the original Nb/Th data, and the growth curve itself is subject to large uncertainty, as discussed in Section 3.1. The other proxy, based on the $^4He/^3He$ ratios of ocean island basalts (OIB), is worse. The temporal evolution of this helium ratio was determined by correlating (rather arbitrarily) the probability distribution of OIB $^4He/^3He$ ratios with the age distribution of zircons (Parman 2007), and such an exercise is contentious at best. Also, the episodicity in the age distribution of zircons may reflect high preservation rates, rather than high crustal production rates, during supercontinental assemblies (e.g., Hawkesworth et al. 2010).

The possibility of a widespread magmatic shutdown between 2.45 and 2.2 Gya, as suggested by Condie et al. (2009b), is based on a rare coincidence of a variety of observations in this time window: no arc-type greenstones or tonalite-trondhjemite-granodiorite suites found, only one large igneous province reported, major unconformities on most cratons, and a gap in deposition of banded iron formation. The last two suggest a major drop in sea level, which may reflect unusually slow plate tectonics or even stagnant lid convection. The duration of this window is only 0.25 Ga, so its impact on thermal evolution would be minimal. Nevertheless, an observational hint for a substantial slowdown of plate tectonics would be important in discriminating among different models of thermal evolution (**Figure 8**). The closed-system and open-system evolution models discussed in Section 2.3 are both end-member scenarios with, respectively, zero and nearly constant transport of water into the mantle. The work of Condie et al. (2009b) may point to a more realistic, intermediate situation with time-varying water influx.

### 3.4. Preservation Bias and Its Causes

As discussed in Section 3.1, identifying noncontroversial indicators for plate tectonics is challenging beyond the late Archean because, in addition to addressing the difficulty of interpreting antique rocks, we must venture to generalize highly regional observations. An alternative approach based on continental growth and the evolution of mantle geochemistry may appear promising, but in light of the thermal evolution of Earth, its implicit assumption about the mixing efficiency of mantle convection is likely to be undermined, and more careful consideration of the spatial and temporal evolution of the depleted mantle is warranted. Section 3.2 explains that exploiting metamorphic rock records to infer the style of mantle convection demands us to predict the secular evolution of a subduction zone process that we do not fully understand even for its present-day operation. In both cases, substantial progress in geodynamics is needed to quantitatively test various speculations.

Preservation bias is another issue to be considered. We all hope that our favorite field area tells us something general and global, but its preservation up until today despite plate tectonic recycling may already demand a special condition. As different interpretations of zircon age distribution (e.g., Condie 1998, Hawkesworth et al. 2010) illustrate, considering the possibility of preservation bias leads us to a more complete understanding of relevant processes. The strength of lithospheric mantle beneath continental crust, for example, may explain the preferred preservation of a certain type of continental crust (Korenaga 2008a). With a relatively constant mantle heat flux through time (**Figure 5**), plume activities in the past are likely to have been more reduced than at the present (e.g., Sleep et al. 1988, Korenaga 2006), but contrary to this theoretical expectation, geological indicators for flood basalts are common in Archean terranes (e.g., Campbell et al. 1989, Ernst & Buchan 2003, Sandiford et al. 2004). Continental crust with flood basalts may have better survived if it were underlain by a residual mantle for flood basalt magmatism because such residual mantle formed by a large degree of melting would be dry and strong (Pollack 1986, Hirth et al. 2000, Katayama & Korenaga 2011). The early stratigraphy of the Pilbara craton in Australia is, for example, dominated by plume-related vertical tectonics (e.g., Van Kranendonk et al. 2007), and depleted lithosphere associated with such plume activities may have helped the survival of this craton through subsequent tectonic disturbances.

In relation to this, Herzberg & Rudnick (2012) recently suggested a possible connection between the thermal history of Earth and the formation ages of cratonic lithosphere. As the ambient mantle in the Archean was likely to have been >200 K hotter than at the present (Herzberg et al. 2010) (**Figure 5**), its melting under mid-ocean ridges would be similar to the melting of present-day plumes, producing thick crust and thick depleted mantle lithosphere (**Figure 9**), the latter of which could contribute to the formation of cratonic lithosphere. The broad maximum around 3 Gya in the thermal history coincides with the formation age for cratonic peridotites, which is estimated to be ~2.5–3.5 Gya with a peak at ~2.9 Gya (e.g., Griffin et al. 2002, Carlson et al. 2005, Pearson & Wittig 2012), and Herzberg & Rudnick (2012) proposed that the efficient production of cratonic lithosphere was limited to this time span because, in the concave-upward thermal history, it was the only time when Earth was hot enough to melt extensively. That is, cratonic lithosphere older than ~3.5 Ga does not exist today because it was not formed in the first place.

This line of reasoning may be extended further to explain the relative paucity of Archean rocks with respect to Proterozoic rocks, and even the absence of Hadean rocks. Without strong lithospheric support, the continental crust may be easily recycled by plate tectonic processes. To explain the fate of abundant Hadean crust in his continental growth model (A81 in **Figure 11*b***), Armstrong (1981) proposed that the rate of crustal recycling was much higher than it is today, assuming rapid plate tectonics in the hotter past. Although his physical reasoning is at variance with the likely heat-flow scaling of mantle convection, the efficient recycling of Hadean crust may instead be explained by the unavailability of strong cratonic lithosphere. A similar effect can also be expected from the mantle hydration hypothesis (see Section 2.3); depleted lithosphere would not have been any stronger than asthenosphere before the latter became reasonably hydrated by subduction. These suggestions about the role of depleted lithosphere are, however, still speculative, and further studies are required to explore their plausibility. As discussed next (see Section 4.1), the dynamics of cratonic lithosphere also plays an important role when relating the thermal evolution of Earth to the surface environment.

# 4. DISCUSSION

## 4.1. Evolution of Surface Environment

The possibility of ocean-mantle interaction and its influence on thermal evolution (see Section 2.3) highlight two potentially time dependent variables, the water content of the mantle and the volume of oceans, both of which have tended to be assumed constant in previous studies. A hotter mantle in the past, for example, is usually expected to be less viscous, and this expectation is equivalent to assuming that the water content of the mantle does not change with time. Also, the volume of oceans is often considered to be constant (e.g., Schubert & Reymer 1985), probably for the sake of simplicity, in the absence of observational constraints suggesting otherwise. Indeed, whether the ocean volume could change with time, or whether Earth's mantle has been dehydrating or hydrating, is a subject of long-standing controversy involving geochemistry, mineral physics, and mantle dynamics (e.g., Fyfe 1978, Ito et al. 1983, Jarrard 2003, Rüpke et al. 2004, Smyth & Jacobsen 2006, Parai & Mukhopadhyay 2012).

As an entirely independent estimate from these studies, Korenaga (2011) derived the upper bound on the long-term water influx into the mantle to be ~2–3 $\times$ $10^{14}$ g year$^{-1}$ on the basis of the constancy of continental freeboard and the thermal evolution of Earth (corresponding to the case of open-system evolution shown in **Figure 8**). This rate is equivalent to a ~50% increase in the ocean volume at 3 Gya, which is accommodated by deeper ocean basins in more sluggish plate tectonics. In contrast, if there were no net water influx, the mean depth of mid-ocean ridges would have had to be at the sea surface for >2 Ga to satisfy the freeboard constraint (**Figure 8*c***). A realistic situation may be found between these end-member scenarios, but which end-member is more likely? At first sight, the constant depth of mid-ocean ridges assumed in open-system evolution may seem unrealistic because oceanic lithosphere is more chemically buoyant for a hotter mantle (see Section 2.4). What is important under the freeboard consideration is, however, the relative buoyancy of oceanic lithosphere with respect to continental lithosphere, and the buoyancy of continental lithosphere could have been higher as well because of greater thickness in the past. In fact, numerical studies on the stability of continental lithosphere generally exhibit its gradual thinning by convective erosion (e.g., Doin et al. 1997, Shapiro et al. 1999). The relation

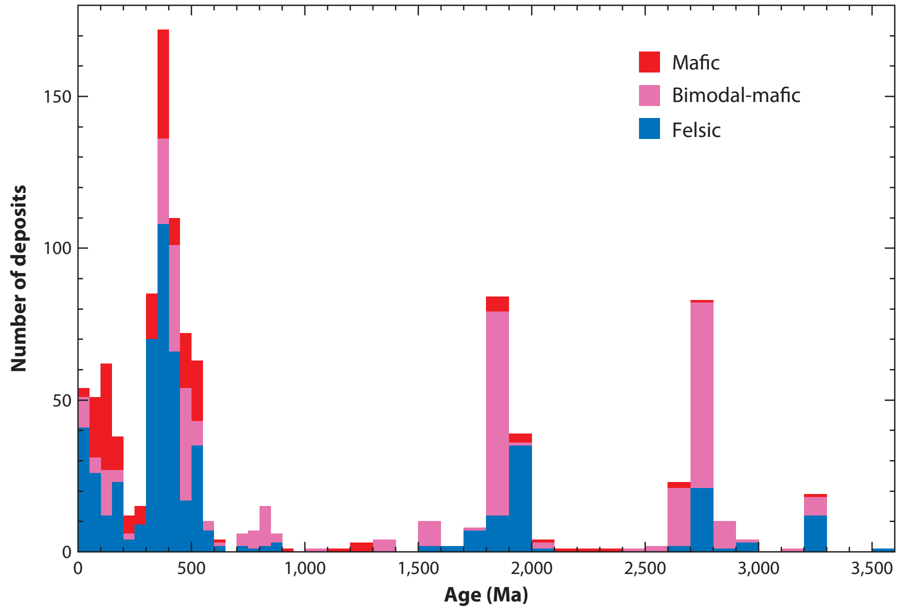


**Figure 13**

Age distribution of volcanogenic massive sulfide deposits in the world (Mosier et al. 2009). Colors denote classification of deposits based on host lithologies: mafic (*red*), bimodal-mafic (*pink*), and felsic (*blue*).

between the depth of mid-ocean ridges and the mean continental level, therefore, may not vary much with time.

Geological constraints on the history of ocean volume are limited. The existence of oceans in the Archean is well supported by the formation of sedimentary rocks and pillow lavas (e.g., Nutman et al. 1984, Kusky et al. 2001), and that in the Hadean is also suggested by the oxygen isotope signature of Hadean zircons (e.g., Mojzsis et al. 2001, Wilde et al. 2001). These observations, however, do not place useful constraints on the ocean volume or the ridge depth; the geological record of mineral deposits may have more potential for doing so. The formation of volcanogenic massive sulfide (VMS) deposits, which are commonly found at mid-ocean-ridge hydrothermal vents today (e.g., Hannington et al. 2005), requires the depth of seafloor to exceed ∼2 km (Ohmoto 1996). For hydrothermal fluids to transport sufficient amounts of metals and reduced sulfur to form a deposit, the fluids must be heated to >300°C, but such hot fluids would boil at pressures lower than ∼20 MPa (corresponding to the seawater depth of ∼2 km) and could not transport sufficient quantities of metals. Because of their economic value, massive sulfide deposits have been explored around the world (e.g., Franklin et al. 2005), and even Archean deposits are relatively common (**Figure 13**). VMS deposits are formed universally in extensional tectonic settings, including both oceanic and continental environments. Those formed at mid-ocean ridges are usually subducted and thus are rare in the geological record of VMS deposits, but the existence of mafic-type VMS deposits, which are thought to form in a primitive oceanic backarc (Mosier et al. 2009), suggests that mid-ocean ridges may have been sufficiently deep in the Proterozoic and Archean.

The constancy of continental freeboard is only an approximate notion (e.g., Eriksson et al. 2006), depending on the balance of two competing processes: (*a*) the erosion of continents above sea level and (*b*) the deposition of sediments on continents below sea level. When all continental masses were under water, possibly in the Archean (e.g., Arndt 1999), however, there was no obvious mechanism to bring them all up to sea level. The net subduction of water may therefore be a process essential to activating the constancy of the freeboard by exposing a dry landmass (Korenaga 2011). Because the emergence of a dry landmass in Earth's history facilitates the carbon cycle (Kasting & Catling 2003) and may explain the uprise of oxygen (Kump & Barley 2007), ocean-mantle interaction has far-reaching implications for the evolution of the atmosphere. Also, the rate of mantle degassing is directly related to the vigor of mantle convection, so the notion of more sluggish plate tectonics in the past becomes important in this context as well (e.g., Padhi et al. 2012).

From a theoretical point of view, the ideal initial water distribution for the operation of plate tectonics is surface water underlain by a dry mantle; the former weakens the otherwise stiff lithosphere, and the latter leads to high convective stress. The possibility of Hadean plate tectonics suggested by the geochemistry of ancient zircons (e.g., Hopkins et al. 2010) may support such an initial condition. Research on Hadean zircons is a rapidly developing field, however, and these interpretations are expected to fluctuate for a while. The initial state of subsolidus mantle convection may also be resolved by studying how the putative magma ocean solidified (e.g., Abe 1997, Solomatov 2007, Elkins-Tanton 2012).



## 4.2. Unresolved Geodynamical Issues

Throughout this review article, our deficiency in theoretical understanding is often suggested as a primary bottleneck for a relevant issue, e.g., the rheological evolution of oceanic lithosphere for the initiation of subduction (see Section 2.1), the efficiency of mantle mixing for the evolution of the depleted mantle (see Section 3.1), and subduction zone dynamics for high-pressure metamorphism (see Section 3.2). In addition, our understanding of heat-flow scaling for mantle convection (see Section 2.3) is still limited. The scaling derived by Korenaga (2010) for plate tectonics is, for example, based on two-dimensional numerical simulations that incorporate the pseudoplastic approximation of brittle failure, a single ductile deformation mechanism, no depth dependency in reference viscosity, and no phase transition. The significance of these simplifying assumptions remains to be assessed.

At a more fundamental level, we may also ask about the validity of the parameterized convection approach (see Section 2.3), which has long been used in the study of thermal evolution (e.g., Schubert et al. 1980, Christensen 1985, Richter 1985, Davies 1993, Honda 1995, Solomatov 2001, Korenaga 2003, Davies 2009). In this approach, heat-flow scaling developed for steady-state convection is applied to model the evolving Earth, assuming that the relation between mantle average temperature and surface heat flux at each time step is approximated reasonably well by the scaling. Whereas this assumption has been validated for simple convection systems (e.g., Daly 1980, Choblet & Sotin 2000), its applicability to more realistic, complex systems is not obvious. Also, when a coupled ocean-mantle system in parameterized convection is considered (e.g., McGovern & Schubert 1989, Korenaga 2011), the loss or gain of water must be assumed to be instantly reflected in the composition of the entire mantle. As discussed above (see Section 3.1), however, chemical homogenization by convective mixing is expected to entail a considerably long timescale.

An alternative approach may be to directly simulate thermal evolution with numerical modeling, but this is also a challenging path because such numerical simulation has to reproduce, at the end of simulation, the present-day Earth sufficiently closely, in terms of its major characteristics such as mantle temperature, surface heat flux, the number of plates, and the amount of continental crust. In contrast, these realistic complications are implicitly taken into account in parameterized convection by normalizing assumed heat-flow scaling by the present-day condition (**Figure 4**). Here, scaling laws obtained from studying the systematics of steady-state convection are used to quantify the sensitivity of the present-day Earth system with respect to changes in the mantle temperature, although the accuracy of such an estimate based on relatively simple convection systems is a concern, as noted above. Despite such difficulty, suppose that we are able to obtain a successful simulation result. What we also have to do in this case is to test its robustness by conducting a variety of sensitivity tests on assumed material properties because most of them, in particular transport properties such as viscosity, are poorly constrained under high temperature and pressure conditions by available experimental data. Sensitivity tests in parameterized convection are much less costly, not only because the number of model parameters is already minimized but also because relevant scaling relations are already given.

Parameterized convection may be too simplistic, but choosing direct simulation may also be too naive, as it would not necessarily guarantee a better, more robust understanding in the face of our limited knowledge of the physics and chemistry of Earth materials. In general, running multiple numerical models alone does not give us a deep understanding of what is being modeled. Such a process should be followed by some kind of theoretical abstraction of modeling results, an effort to extract physical scaling with broader applicability. For example, a direct simulation could be run, and its result could be compared with the prediction from corresponding parameterized convection. When these two approaches yield different results, we may benefit from asking, for example, whether there is any systematics for the discrepancy and whether such systematics can be parameterized using fluid mechanics. A next-generation approach for thermal evolution modeling may be enabled by extending the traditional approach with more elaborate parameterization, and the spirit of abstraction will always be useful if we wish to understand the complex Earth system (e.g., Berner 2004).

## 5. SUMMARY AND OUTLOOK

On the basis of theoretical considerations and available observational constraints, the possibility of initiating plate tectonics in the Hadean is difficult to reject, and regardless of the exact timing of onset, the subsequent evolution of plate tectonics is likely to be characterized by relatively constant surface heat flux and a gradual increase in plate velocity. This secular evolution of plate tectonics is consistent with the geochemical budget of internal heat production (McDonough & Sun 1995, Lyubetskaya & Korenaga 2007b), the secular evolution of passive margin life span (Bradley 2008), the petrological estimate of thermal history (Herzberg et al. 2010), and the atmospheric budget of radiogenic xenon (Padhi et al. 2012). The likelihood of plate tectonics in the early Earth is suggested to depend on the nature of ocean-mantle interaction, i.e., whether subduction results in the net hydration of the convecting mantle over Earth's history. With net water influx to the mantle, the convecting mantle would have been drier in the past with more voluminous oceans, and such a situation is ideal for driving plate tectonics.

Substantially improving the status quo requires that several important challenges be undertaken. First, needless to say, seeking further observational constraints in continental rocks and Hadean zircons will continue to be vital. In particular, geological constraints on the history of ocean volume, such as mafic-type VMS deposits, or on the state of mantle hydration in the past are in critical demand. Second, a better understanding of the rheological evolution of oceanic lithosphere is a key to the long-standing puzzle of why plate tectonics takes place on Earth. Innovative field programs in marine geophysics may be essential to make progress on this issue. Finally, concurrent developments in relevant theoretical issues, such as mantle mixing and subduction zone dynamics, will allow us to better interpret geological data. This review takes a theory-oriented approach, and given our limited understanding on various theoretical problems, its main conclusion may be revised considerably in the future by more sophisticated modeling of thermal evolution.

The virtue of the theory-oriented approach should, however, remain useful in placing diverse and fragmentary geological observations in a coherent framework and building a satisfactory theory for the Earth system evolution.

## DISCLOSURE STATEMENT

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