A window for plate tectonics in terrestrial planet evolution?

Craig O'Neill a,*, Adrian Lenardic b, Matthew Weller b, Louis Moresi c,d, Steve Quenette c, Siqi Zhang a

Abstract

The tectonic regime of a planet depends critically on the contributions of basal and internal heating to the planetary mantle, and how these evolve through time. We use viscoplastic mantle convection simulations, with evolving core–mantle boundary temperatures, and radiogenic heat decay, to explore how these factors affect tectonic regime over the lifetime of a planet. The simulations demonstrate (i) hot, mantle conditions, coming out of a magma ocean phase of evolution, can produce a “hot” stagnant-lid regime, whilst a cooler post magma ocean mantle may begin in a plate tectonic regime; (ii) planets may evolve from an initial hot stagnant-lid condition, through an episodic regime lasting 1–3 Gyr, into a plate-tectonic regime, and finally into a cold, senescent stagnant lid regime after ~10 Gyr of evolution, as heat production and basal temperatures wane; and (iii) the thermal state of the post magma ocean mantle, which effectively sets the initial conditions for the sub-solidus mantle convection phase of planetary evolution, is one of the most sensitive parameters affecting planetary evolution – systems with exactly the same physical parameters may exhibit completely different tectonics depending on the initial state employed. Estimates of the early Earth’s temperatures suggest Earth may have begun in a hot stagnant lid mode, evolving into an episodic regime throughout most of the Archaean, before finally passing into a plate tectonic regime. The implication of these results is that, for many cases, plate tectonics may be a phase in planetary evolution between hot and cold stagnant states, rather than an end-member.

1. Introduction

Geologists have long debated the timing of the onset of plate tectonics on Earth (e.g. O’Neill et al. (2007b); and see Condie and Pearce (2008), and papers within). There has been a consensus that as Earth cools, tectonic activity will wane and eventually Earth will settle into a cold, stagnant-lid regime, similar to Mars today (e.g. Nimmo and Stevenson (2000), O’Neill et al. (2007a,b)). However, there is no such consensus on what form tectonics might have taken during the Earth’s deep geologic past (e.g. Davies (1993), Calvert et al. (1995), Condie and Kroner (2008), O’Neill et al. (2007b), Stern (2008), Moyen and van Hunen (2012), Moore and Webb (2013)).

This debate has extended into exosolar planets (O’Neill and Lenardic, 2007; Valencia et al., 2007; Korenaga, 2010; van Heck and Tackley, 2011; Foley et al., 2012; Noack and Breuer, 2014; Stein et al., 2012; Stamenkovic and Breuer, 2014; Karato, 2014), with arguments for and against the likelihood of active tectonics on larger superearths. One of the ambiguities in the debate is the extent to which the heating mode affects surface stresses and tectonic regime.

The ultimate tectonic state of a planet is a result of a balance between the coupling of the plates and the mantle beneath, and also the buoyancy forces driving convective motion. These two factors are critically sensitive to how a planet’s thermal state evolves through time; buoyancy forces are strongly coupled to the temperature drop across the convecting mantle, and induced lithospheric stresses to the internal viscosity, and thus mantle temperature.

Simple scaling theories, based on basally heated convection, demonstrate increased convective stresses with increasing Rayleigh number – which, in isolation, translates to higher lithospheric stresses for larger planets (Valencia et al., 2007). However, scaling relationships for mixed heating thermal convection can be quite different than for either purely basal, or internally, heated cases (Moore, 2008), and increases in internal heating rate for a given planet’s size has also been shown to be able to cause a transition from mobile lid convection, into an episodic regime, and eventually into stagnant lid convection (O’Neill et al., 2007a,b; O’Neill and Lenardic, 2007).
One of the issues bounding the debate over planetary tectonics is the importance of initial conditions in determining the evolutionary path through tectonic regimes. The pioneering work of Tozer (1972) suggested that the negative feedback effects of temperature-dependent viscosity buffer the system so that initial conditions quickly decay, and the system reaches equilibrium with its internal heat generation. More recently, it has been noted that, near the critical transition zone, non-linearities in the physical system lead to an inherent hysteresis, and the stable tectonic regime can depend strongly on starting state (Weller and Lenardic, 2012; Crowley and O’Connell, 2012; Weller et al., 2015). A simulation that starts hot may finish in a different tectonic regime to one that starts cold.

As a result, the time scale for the equilibration of planetary tectonics and thermal state may be long – so much so that a planet’s entire evolution may be governed by its response to initial conditions. The degree to which the starting state of a planet, as opposed to its evolving heat sources, governs its subsequent tectonic history is not clear, nor is the impact of hysteresis in rapidly evolving planetary systems (Weller et al., 2015). One way to address this problem is to simulate the tectonic evolution of planets with evolving heat production, and core temperatures, for a range of initial states. However, the problem is computationally challenging. Simulations with plastic yielding suffer slow convergence, and the lifetime of such a simulation could be $\gg 10$ Gyr. There are two possible philosophies here; either one attempts to simulate a small number of highly-resolved 3D runs, with an evolving parameterised core model, or one simulates a geometrically simpler system to allow a greater exploration of the parameter space – at the expense of a self-consistent core model. We have adopted the latter approach, and impose a core evolution, to allow a greater understanding of unknown parameters on the evolution of these systems. We also include exponential decay of heat production through time, and explore potential evolution scenarios for different initial thermal conditions.

2. Methods

We employ a widely available community code (Underworld: www.underworldproject.org; Moresi et al., 2007) to perform Cartesian simulations pertinent to the whole mantle with variable varying Rayleigh number, internal heating rates, and basal temperature conditions. The code solves the standard convection equations using the Boussinesq approximation to the equation of state. The momentum conservation equation is:

\[ \tau_{ij} - p_{j} = f_{i} \]  
(1)

where $\tau$ is the deviatoric stress tensor, $p$ the pressure, and $f$ a body force, representing gravity in the vertical direction (given by $Ra \cdot T$). The deviatoric stress, for Newtonian materials, is related to the strain rate $\varepsilon$ and hence velocity $v$ by:

\[ \tau_{ij} = 2\eta \dot{\varepsilon}_{ij} \equiv \eta(\dot{\varepsilon}_{ij} + \dot{v}_{ij}) \]  
(2)

This momentum equation is subject to the incompressibility constraint:

\[ \nabla \cdot v = 0 \]  
(3)

And the energy equation is given by:

\[ \partial T / \partial t + \nu \cdot \nabla T = \kappa \nabla^{2} T + Q(t) \]  
(4)

Here $T$ is the temperature, $t$ the time, $\kappa$ the thermal diffusivity, and $Q$ the rate of heat production. We non-dimensionalise the problem using the standard identities (Moresi and Solomatov, 1998):

\[ x_{i} = d x_{i}, \quad T = \Delta T' \quad \eta = \eta_{0} \eta' \]  
(5)

where $d$ is the layer depth ($2,890,000$ m, Fig. 1), $\eta_{0}$ is a reference viscosity ($1e21$ Pa.s), and $\Delta T$ is the reference temperature drop across the system ($2555$ K non-adiabatic drop), and primes denote non-dimensional values (the rest of the manuscript assumes non-dimensional values unless stated otherwise, and drops the primes for clarity). To obtain kinematic similarity, the system was scaled using a convective overturn time (e.g. Zhong and Gurnis, 1995) of 50 Myr (this defines the time and velocity scalings).

We do not consider phase transitions or depth-dependent properties in this current study. We have included a Frank-Kamenetskii style temperature dependent viscosity, of the form:

\[ \eta = A \cdot e^{-T_{1}/T} \]  
(6)

The viscosity varies from 1 at the bottom (where the reference basal temperature is 1), to $3 \times 10^{4}$ at the top where the surface temperature condition is set to 0 (i.e. $A$ is 3e4, and $T_{1} = 10.31$). Reese et al. (1999) and Noack and Breuer (2013) discuss the validity range of the Frank-Kamenetskii viscosity approximation we use, and how it corresponds to a corresponding Arrhenius law – for our range of interest the differences are irrelevant. We have also used a depth-dependent Byerlee style plastic yielding criteria, to enable the mobility of the surface plates. The yield criterion is of the form:

\[ \tau_{\text{yield}} = B_{\theta} + B_{z} \cdot z \]  
(7)

The default non-dimensional parameters are set to $B_{\theta} = 1 \times 10^{5}$ and $B_{z} = 1 \times 10^{7}$ (these equate to a cohesion of $\sim 12$ MPa, and a coefficient of friction of $\sim 0.1$, appropriate for a water-altered rheology (Escartin et al., 1997)). These models are very similar to the default models of Moresi and Solomatov (1998), who explore variation in yield parameters in detail, and this choice facilitates comparison and reproducibility. Further details of the plasticity formulation are found in that paper, and in the Appendix.

The basic model setup is outlined in Fig. 1. The models have periodic side boundary conditions, and free-slip top and bottom
boundary conditions. The top boundary is also held at a constant temperature. In order to study the evolution of tectonics in thermally evolving planets, we have implemented time-dependent heat production and CMB temperature functions, as shown in Fig. 2. Heat production $Q$ follows a simple exponential decay:

$$Q(t) = Q_0 e^{-at}$$

(8)

where $Q_0 = 2.25$ and $a = 3.34239$, are set to give equivalent present day heat production values, and a decay over time similar to that calculated based on decay of radioactive isotopes (Turcotte and Schubert, 1982). The core temperature evolution is a simple linear function, and is set to give Nimmo et al.’s (2004) maximum CMB temperatures at 4.5 Gya ($T_{base} = 1.2$ at $t = 0$), and decay to present values over an equivalent period ($T_{base} = 1$ at $t = 0.45$ – equivalent to ~4.5 Gyr). The evolution of core temperatures follows this function until $T_{base} = 0.5$, where it is held constant (this occurs approximately 15 Gyr after the beginning of the simulation, using a convective overturn scaling). We explore the response of the system to variations in these functions in a later section.

To address the issue of planetary evolution, the simulations minimally have to evolve from the initial starting conditions of a post-accretion planet, to the end of its tectonic lifetime – a period that could be as long 10–20 Gyr. At the same time, timestepping constraints, particularly during the hot early evolution, restrict the length of a solution timestep. This is a difficult task numerically, and in order to map out a solution space (as opposed to running only a few very large models), we have for this attempt primarily made use of 2D, 1 × 1 Cartesian simulations. The tectonic regime has previously been shown to be fairly insensitive to geometrical factors (e.g. Stein et al., 2004), and we demonstrate equivalence between our 2D, and 3D spherical simulations in later sections. The advantage of this approach is: (1) we can cover a large parameter suite, and run many models – condensed here – to demonstrate the robustness of results over a wide-range of parameters; (2) we can understand the physics of these systems very well, and delineate the main, first-order physical processes driving the often complex behaviour of these deceptively simple systems; (3) as these models run fast, we can run them for an extraordinarily long time (>15 Gyr), which is often not feasible for massively parallel 3D systems, but essential for understanding the long-term evolution of a planet with waning heat sources.

The model equations we solve are applicable to sub-solidus mantle convection and, as such, we are not tracking a planet’s evolution from accretion onward but from the post magma ocean phase of its evolution onward. This means that the initial conditions applied to our models represent the state of the Earth coming out a magma ocean phase of evolution. The nature of magma ocean evolution remains debated with several possible scenarios remaining viable (e.g. Elkins-Tanton et al., 2003; Debaille et al., 2009; Wood et al., 2006). Variations in initial conditions, imposed for our models, test how sensitive the system is to the state of the mantle after the bulk of the mantle has solidified and sub-solidus convection dominates its dynamic evolution. Variations in the total radiogenic content of the mantle can also affect the state of the post magma ocean mantle, and recent work suggests that differences in the accretion history of terrestrial planets can lead to variations in radiogenic heat content even for planets of the same bulk size and mass (Jellinek and Jackson, 2015).

### 3. Results

#### 3.1. Illustrative models: sensitivity to initial conditions

In this section we explore some illustrative statistically steady-state models to highlight the effects of system hysteresis, and associated dependence on initial state, even in relatively simple convecting regimes. Fig. 3a demonstrates this effect for two different simulations with exactly the same non-evolving convection parameters (as per Table 1, $T_{base} = 1$ and $Q = 1$). The first simulation begins with an initially linear temperature profile (from 0 at the surface to 1 at the base), with a small sinusoidal perturbation. The second simulation begins with input from a previous simulation, the latter being run with $Q = 2$, and which was in a stagnant lid regime. The difference is marked. The simulation starting from a cold, linear initial condition, rapidly evolves into a mobile-lid regime, from which it does not depart. The second simulation, which possesses exactly the same system parameters as the first, but starts from a hot initial condition inherited from a previous simulation, immediately enters into an episodic regime, from which it does not evolve. Though the episodic regime exhibits extreme peaks in heat flux, the long-term average is slightly less than for the mobile lid simulation. Similarly, Fig. 3b demonstrates the effect for simulations with $Q = 2$, and $T_{base} = 0.8$. Here, the simulation with the cold initial condition settles into a cyclic episodic overturn mode, whilst the simulation which initiated hot (from a previous output) rapidly evolves into a stagnant lid regime, and stays in that regime for the course of the simulation.

In Fig. 4, we demonstrate the impact of initial state in simulations of mantle convection in a 3D spherical shell. We evaluate the effects of decreasing and increasing internal heating rates on a planetary tectonic system by fixing the Ra, viscosity variation ($\delta \eta$), and core–mantle boundary and surface temperatures ($T_b, T_s$) at: $Ra = 1e5$, $\delta \eta = 1e4$, and $T_b = 1$, and $T_s = 0$. Viscosity is temperature, and depth dependent. The yield strength $\sigma_{yield} = 4.25e5$ was chosen such that the system exists in a relatively weak, near transitional stagnant-lid regime (transitions into stagnant-lid behaviour occurs at $\sigma_{yield} = 4.00e5$ for an increasing yield strength pathway [e.g. Weller and Lenardic (2012)]. The non-dimensional internal heating ($Q$) is varied from 60 to 5 in the decreasing pathway, and 0 to 60 in the increasing pathway. For each incremental change in internal heating, the preceding simulation’s thermal history is used as the initial conditions for the current simulation. Each change in $Q$ is allowed to run for sufficiently long time scales...
The cold initial condition simulation enters into an episodic overturn regime. \( Q = 0.8 \). The case with hot initial conditions evolves into a hot stagnant lid mode, but enters into an episodic overturn mode. (b) As per 3a, but with steady-state mobile-lid mode, whilst the simulation with a hot initial condition with \( Q \) \( T_0 = 1 \) enters a mobile-lid mode, and it remains in this mode as both \( Q \) \( T_0 \) change. The case with hot initial conditions evolves into a hot stagnant lid mode, but the cold initial condition simulation enters into an episodic overturn regime.

**Fig. 3.** (a) Nusselt number versus time for two simulations with identical system parameters, but different starting conditions. System parameters as per Table 1, with \( Q = 1 \) and \( T_{\text{base}} = 1 \). The initial condition for the cold-start case is a linear temperature gradient between the top and the bottom, with a small sinusoidal perturbation. The initial condition for the hot-start case is a previous stagnant-lid simulation, which had a \( Q = 2 \). The simulation with a cold initial condition enters a steady-state mobile-lid mode, whilst the simulation with a hot initial condition enters into an episodic overturn mode. (b) As per 3a, but with \( Q = 2 \) and \( T(\text{base}) = 0.8 \). The case with hot initial conditions evolves into a hot stagnant lid mode, but the cold initial condition simulation enters into an episodic overturn regime.

**Table 1**

Default model parameters (non-dimensional).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( Ra )</td>
<td>( 1 \times 10^7 )</td>
</tr>
<tr>
<td>Viscosity</td>
<td>( \eta = \eta_0 \exp(-tT) )</td>
</tr>
<tr>
<td>( \eta_0 )</td>
<td>( 3 \times 10^4 )</td>
</tr>
<tr>
<td>( T_1 )</td>
<td>10.31</td>
</tr>
<tr>
<td>Yield parameters</td>
<td></td>
</tr>
<tr>
<td>( B_0 )</td>
<td>( 1 \times 10^2 )</td>
</tr>
<tr>
<td>( B_r )</td>
<td>( 1 \times 10^7 )</td>
</tr>
<tr>
<td>Thermal conditions</td>
<td></td>
</tr>
<tr>
<td>( Q )</td>
<td>0.5</td>
</tr>
<tr>
<td>( T_{\text{base}} )</td>
<td>1.0</td>
</tr>
<tr>
<td>( T_0 )</td>
<td>0.0</td>
</tr>
</tbody>
</table>

*This is based on the O’Farrell et al. (2012, hereafter OL12) argument on lowering heating rates in planar geometries to better approximate spherical system temperatures. For our \( Ra_{\text{base}} = 1e7 \) system, the ‘average’ \( Ra \) (Eq. (7) or OL12) is 4.47e5. Inputting this value, and an \( H \) value of 11 (normalised heating rate for Earth: heat_production/conductive_heat_flow for mantle heat production of 5e-12), into OL12 Eq. (14), with a mean system viscosity, gives a value for theta of 0.245. Inserting this into their planar equation (Eq. (15)) gives an \( H \) value of ~0.50.*

For high heat production (\( Q = 60 \)), the simulation is in a stagnant-lid mode. The system is run till steady state, then the thermal field is used as input into the next decremental simulation (\( Q = 59, 58, \) etc). The system remains in a stagnant lid mode until \( Q = 26 \), when it transits into mobile lid convection. It enters a sluggish model at \( Q = 15 \), before eventually entering a cold stagnant lid mode at \( Q = 0 \). If we reverse the evolutionary path at this point, using the “cold” stagnant thermal field (\( Q = 0 \)) as input into the next simulation (\( Q = 1 \), then output from \( Q = 1 \) is the input to \( Q = 2 \), etc), then the system quickly enters a mobile-lid regime, where it remains until \( Q = 59 \) where the system enters an episodic regime, before going stagnant at \( Q = 60 \). Thus for each of these evolutionary paths, different outcomes are possible for exactly the same system parameters, depending on the initial conditions.

Evidence suggests Earth is still losing its primordial heat, and the demonstrated sensitivity to initial conditions suggests that rather than statistically steady-state simulations, we should rather be attempting to model Earth evolution holistically, as its thermal state, and current tectonics, is strongly dependent on its tectono-thermal history. That is, the assumption that the mantle evolves through a series of quasi-equilibrium states (e.g. Davies (1980)), which is an underlying assumption behind parameterized convection and the use of statistically steady state calculations to determine scaling relations for mantle cooling, can break down for systems that display hysteresis effects (Moore and Lenardic, 2015). For systems that display episodic behaviour the differences between a statistically steady state treatment and models that treat true disequilibrium cooling can also become significant (Yuen et al., 1995). Collectively this suggest the need for true disequilibrium models and that is the purpose of the latter sections of this paper.

### 3.2. Evolutionary models

**Fig. 5** shows the evolution of a convecting system from an initially hot state, declining eventually into stagnant lid senescence, over the course of around ~15 Gyr. The times for the results will be presented in non-dimensional terms, as this is output from the code and allows direct comparison with other work. Based on system velocities and overturn times, a non-dimensional \( f = 1.0 \) is the equivalent of ~10 Gyr real time. The initial condition for this simulation was the output from a previous, statistically steady-state simulation with \( Q = 8 \), which was in a stagnant lid regime. The simulation here begins in a hot stagnant lid. It rapidly evolves, however, losing its primordial heat, and approaching congruency with its internal heat generation. The system fairly rapidly evolves in an episodic convection regime (Fig. 5b–d show an example of an overturn). By about \( t = 0.5 \) it has cooled sufficiently to enter a mobile-lid mode, and it remains in this mode as both \( Q \) and \( T_{\text{base}} \) wane, until eventually the buoyancy stresses are insufficient to move the lid, and system enters into its final senescent stagnant lid mode (Fig. 5f). The lid in this final mode exhibits some viscous deformation due to the low viscosity contrast across the system, similar to the sluggish-lid regime of Solomatov (1995).

**Fig. 6** shows the Nusselt number (effectively surface heat flux here), and root-mean square velocity versus time for the
Simulation shown in Fig. 5. As can be seen, the episodic overturn events dominate the heat flux in the early evolution of the simulation. The geological implications of these sorts of events have been discussed in O’Neill et al. (2007a,b), Condie et al. (2009), Condie and O’Neill (2011), and such events would have a significant impact on the geological record, and impart a strong time-dependence on it, as is perhaps observed in geologic data (O’Neill et al., 2007b). The heat lost during the subduction events is far above the background level, which demonstrates that this regime is an extremely efficient way of cooling the mantle on an early Earth. After about $t \approx 0.3$, the system has cooled sufficiently to enter a mobile-lid phase, during which heat production and CMB temperatures continue to decline, giving the regular decline in Nu observed in Fig. 6a. After $t \approx 1.75$, the system has irrevocably cooled and convection enters a “cold” stagnant-lid mode. Note the core-temperatures are held constant from $t = 1.6$ onwards, however, the association of the stagnant-lid transition time with this is fortuitous.

Interestingly, the system here both begins and ends in a stagnant–lid mode of convection; a hot-stagnant regime at the start, due to high internal temperatures and low internal viscosities, and a cold-stagnant mode at the end, due to the decline in buoyancy forces. Episodic convection here dominates the first few Gyr of tectonic evolution. And finally, plate tectonics is only a “phase” in the evolution of this model – the transition to plate tectonics occurs only once the system has sufficiently cooled; and plate tectonics wanes once the CMB temperatures have cooled too much. This raises the interesting question of whether plate tectonics is merely a phase in the evolution of terrestrial planets, and whether a window of opportunity for it exists.

3.3. Effect of initial conditions on evolutionary models

The evolutionary history of these models hinges critically on the initial model state. In this section we explore the sensitivity of evolutionary path to different thermal conditions, all other parameters being the same (i.e. Table 1, $B_0 = 1 \times 10^5$). To set up the initial conditions, all cases were run at $Ra = 1e7$, with the parameters as per Table 1, except the yield parameters which are set high enough to ensure stagnant lid convection ($B_0 = 1 \times 10^6$, $B_2 = 1 \times 10^8$). These ‘initial conditions setup’ simulations differ solely in their internal heat production $Q$, which results in different final equilibrated internal temperatures. Each of the final steady-state temperature fields are then used as an initial condition for an evolutionary convection simulation. The four simulations presented utilise exactly the same system parameters, and the same time-varying CMB temperatures and declining $Q$ (Fig. 2), they differ only in their starting conditions. Figs. 7 and 8 show two extremes of behaviour observed as a result of the different initial states.

In Fig. 7, the initial condition was an equilibrated temperature field for $Q = 2$. This is a comparatively cold start, and the system rapidly evolves into a mobile-lid regime. It remains in this regime over much of the course of its evolution, transitioning into a quasi-stagnant lid regime only when the CMB temperatures have declined to $\approx 0.5$ (e.g. at $t \approx 2.0$, Fig. 7d). In contrast, Fig. 8 illustrates the progression of a simulation with a starting temperature field equilibrated at $Q = 6$. In this case, the initially stagnant system rapidly flips into an episodic mode for most of its early history, transitioning into a regular mobile-lid mode only after $t \approx 0.9$ (e.g. Fig. 8e). The decline in CMB temperatures and $Q$ result in a transition into cold stagnant lid convection at around $t = 1.6$.

Fig. 4. Effect of initial conditions in a 3D spherical model, using CitcomS. The temperature-dependent viscosity contrast is $1 \times 10^4$, and is also depth dependent. Temperature of the CMB is 1, and the yield strength is 42,500. Internal heating ($Q$) is varied from 60 to 0 in the decreasing pathway, and 0 to 60 in the increasing pathway. Each run is divided into a viscosity plot (grey shells are regions of high viscosity “plates” and yellow bands are regions of yielding) and thermal profiles from the CMB to surface. All other parameters are held constant. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
Fig. 5. Time evolution of a simulation with parameters outlined in Table 1, with imposed decaying mantle heat production, and basal temperature, as per Fig. 2. The initial condition is from a previous stagnant-lid simulation, equilibrated at $Q = 8$ (equivalent to a hot early Hadean). The systems begins in a stagnant regime, before evolving into an episodic regime ($t = 0.263$). Eventually, with waning CMB temperatures, the system transits into a plate tectonic mode ($t = 0.513$, equivalent to ~5 Gyr), before winding down into a senescent stagnant-lid mode ($t = 1.525$, ~15 Gyr; Note arrow size rescales to system maximum each timestep).
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1990); or impacting (Canup, 2004)), then the possibility arises that
plotted against time for the four simulations at
Fig. 9 illustrates the time evolution of these models, with Nu plotted against time for the four simulations at $Q = 2, 4, 6$ and $8$ for the starting temperature field. A dramatic difference is seen between those simulations which start cold ($Q = 2$ and $4$), which enter into a plate-tectonic like mode almost immediately, and those simulations which start hot ($Q = 6$ and $8$), which go through an initial stagnant-$\to$ episodic transition, then stay in an episodic regime for most of their early history, only transiting into a mobile lid mode once it has lost most of its primordial heat. Thus a dra-
matic difference in subsequent tectonic history accompanies only modest changes in the initial thermal state. If Earth came out of a magma ocean phase “cold” – i.e. its thermal state was roughly in equilibrium with radioactive heat production at the time (roughly equivalent to the IC $Q = 4$ example) – Earth may have had plate tectonics over its entire history to date. If, however, the formation of the Earth was accompanied by significant addition of primordial heat (e.g. through either core formation (Stevenson, 1990); or impacting (Canup, 2004)), then the possibility arises that Earth was episodic throughout much of its history, transiting to a plate tectonic regime much later (e.g. O'Neill et al., 2007b). Similar conclusions hold for situations in which accretion leads to signifi-
cant stripping of radiogenic elements vs accretion histories that did not (Jellinek and Jackson, 2015). If the accretion history stripped a planet of radiogenics then it could be more likely to enter a plate tectonic phase (Jellinek and Jackson, 2015). It is also worth noting that it is not just the initial temperature which may set the evolution-
tory stage, Lenardic and Crowley (2012) note that, for equiva-
ent systems, starting in a plate-mode is more likely to maintain plate dynamics than starting with an immobile lithosphere.

3.4. Effect of mantle heat production decay in evolutionary models

Uncertainty exists in the absolute heat production of the mantle, its partitioning into the crust or other reservoirs, and how this has evolved through time. To encapsulate some of this ambiguity, we have simulated different evolving scenarios, for different values for heat production functions.

Here we explore a scenario with an initial cold starting condi-
tion equilibrated at $Q = 2$. With this starting condition, we explore a number of different evolving heat production functions. Expon-
tial heat decay is of the form described in Eq. (8). In each case, we vary the prefactor $Q_k$ for our heat production to be $2.0, 4.0, 6.0,$ and $8.0$. The decay constant $\lambda$ was kept at $3.34239$. We plot the evolution of the Nusselt number, for these “cold” start models, in Fig. 10.

The initial state again contributes the greatest to the variation between models. In the cold-start examples, the decay of heat produc-
tion through time leads the systems rapidly into a mobile-lid regime, and heat flow smoothly follows the decay in heat produc-
tion through time, offset in each case due to the variation in the prefactor in Eq. (8), though this difference diminishes through time. The decay in the average Nusselt number generally follows the decay in heat production, but the time-dependency and non-
linearities of the tectonic regime, particularly at early times, mod-
ulate that trend.

3.5. Effect of core–mantle boundary temperature changes on evolutionary models

Variations in the evolution of core–mantle boundary tempera-
tures affect these simulations primarily by permitting strong basally-driven buoyancy anomalies to dominate internal stresses, for greater or shorter periods of time. In Fig. 11 we have shown the effect of varying the time-dependence of core-temperature evolution, by varying the slope of the graph in Fig. 2. A slope of “1 m” equates to the slope of the CMB temperature curve graph in Fig. 8, whilst, for instance, “0.5 m” corresponds to half the slope, so the effective life of tectonic activity is longer. The beginning and ending CMB temperatures are held the same as per Fig. 2. In all simulations here $Q$ is set to $2.25$ initially, and decays with a con-
stant $\lambda$ kept at $3.34239$, as per previous simulations.

The result of changing the rate of CMB temperature decline is fairly intuitive. Strongly declining core temperatures, equivalent to large slopes (e.g. “2 m”) give rapid declines in tectonic activity; lower slopes, reflecting enhanced CMB temperatures for longer, result in prolonged tectonic activity.

4. Discussions and conclusions

Plate tectonics is often assumed to be the default tectonic state of a planet Earth's size, or larger (e.g. Valencia et al., 2007). Our simulations show this need not be the case. As a planet evolves it may transit through a number of tectonic regimes, from a hot-
stagnant end member in its early evolution, to a cold-stagnant state during senescence. In-between, a planet may pass through episodic overturn or plate tectonic regimes. Previous work had suggested a stagnant to episodic, and finally plate tectonic evolu-
tion of the Earth, based on statistically steady-state simulations, and geological evidence (e.g. O'Neill et al. (2007b); O'Neill and Debaille, 2014). We have shown here this progression is plausible in evolutionary models also, depending on the initial conditions imposed on the model. It has also been previously suggested, from scaling theory and steady-state simulations, that a hysteresis may impact tectonic evolution (Weller and Lenardic, 2012; Weller et al., 2015) and we have demonstrated this behaviour is applicable in evolutionary models as well. Consequently, it is apparent that the specific tectonic evolution of a planet is not solely governed by thermal controls, but the path is also heavily influenced by the initial state of the planet.

Plate tectonics is favoured by (comparatively) cold starting condi-
tions, or systems dominated by basal heating, as these favour strong buoyancy forces, and effective coupling between the mantle and lithosphere. In contrast, systems starting “hot”, or with a higher ratio of internal to basal heating, tend to favour stagnant or episodic tectonics, due to low internal viscosities and thus poor coupling between the mantle and plates, and comparatively lower system stresses.

![Fig. 6. Nusselt number vs non-dimensional time for the simulation shown in Fig. 5](image-url)
The transition between an episodic and mobile-lid evolutionary path remains viable given our uncertainty in mantle thermal state at the time when sub-solidus convection came to dominate its dynamics. For a hot state, Earth could follow a stagnant-episodic evolutionary path for much of the Precambrian. However, if early mantle temperature closely followed the mantle heat production curve (e.g. Turcotte and Schubert, 1982), for the radioisotopes $^{235}$U, $^{238}$U, $^{40}$K and $^{232}$Th, then it could have possessed plate tectonics for its entire evolution. We did not explicitly model short-lived isotopes such as $^{26}$Al or $^{60}$Fe, as they were operative for such a short period of time (i.e. during accretion) we consider them a component of the primordial heat budget.

Unfortunately, for the Earth, our knowledge of its inception, magma ocean evolution, and post magma ocean conditions – our initial condition – is limited by the dearth of geological evidence in the Hadean. Recently, though, constraints have been implied, either directly through $P$-$T$ estimates from Hadean zircons (Hopkins et al., 2008), suggesting low-$T$ felsic melting reminiscent of subduction processes, or indirectly using short-lived isotope systematics (e.g. Debaille et al., 2013; O’Neill and Debaille, 2014), which suggest long-mixing times of mantle isotopic heterogeneities, characteristic of stagnant-lid convection. These constraints have the potential to shed light on the thermal state of this period, but they require a geodynamical framework to understand their implications – something which modelling can provide.

Additionally, models have been proposed (e.g. Stevenson (1990)), which estimate the thermal budget, and calculate initial temperatures, based on assumed dynamics of accretion. Based on this it has been suggested that the Earth begun in an entirely molten state, and as such a mantle solidus may be an appropriate

**Fig. 7.** Time evolution of a system with the default values listed in Table 1, but with an initial condition taken from a previous stagnant-lid simulation with $Q = 2$. The model transits rapidly into a steady plate-tectonic mode (a), which continues, with a gradually thickening thermal boundary layer (b, c), until the simulation eventually transits into a cold stagnant-lid mode (d).
Fig. 8. Time evolution of a system with the default values listed in Table 1, but with an initial condition taken from a previous stagnant-lid simulation with $Q = 6$. The model begins in a stagnant mode, before transiting into an episodic regime (a–d). At $t \sim 1.0$ the system evolves into a plate-tectonic regime (e), which eventually wanes into a cold stagnant lid regime (f).
initial condition for a planet. But not only is the mantle solidus at high-pressures poorly defined, it can vary significantly for different water/volatile contents, and variable compositions expected from magma ocean crystallisation sequences (e.g. Elkins-Tanton et al. (2003) and Debaille et al. (2009)). Work on Martian convection suggests an initial overturn of the unstably stratified magma ocean, which would perturb initial conditions significantly (e.g. Debaille et al. (2009) and Zhang and O'Neill (2015)). Furthermore, work on high-temperature partitioning of metal–silicate mixes suggests a magma ocean ~850 km deep on the Earth (e.g. Wood et al. 2007).
implying a hot adiabat may be a more appropriate starting condition for the lower mantle.

Poorly prescribed initial conditions can generate large start-up effects – something we explicitly wanted to avoid, as the initial conditions in these models exert an important control. Given the uncertainty in these initial conditions, we have opted for a ‘mini mum-artefacts’ approach – allowing a system to come to equilibrium at the initial conditions, and running the system forward from there. There is obviously room for improvement in the refinement of initial conditions in these models, as it may be that in many ways the Hadean era reflects the adjustment of Earth to its initial conditions.

It is worthwhile explicitly outlining the limitations in the simulations presented. Our main purpose here was to explore the effect of varying contribution of basal and internal heating to plate tectonic models with strongly temperature dependent viscosity, particularly in the context of plausible evolutionary scenarios for core temperature and mantle heat production. What these models do not incorporate is depth-dependent structure; such as mantle phase changes or layered viscosity structures, such as an asthenosphere – which also affect the wavelength of convection and thus stresses (Hoinik et al., 2012). These add dimensions of complication and expand the parameter space, but will be an important avenue for further work. The asthenosphere in particular has been demonstrated to be fundamentally important in plate generation and lithospheric stresses (Richards et al., 2001; O’Neill et al., 2008; Hoinik and Lenardic, 2009; King, 2015), but apparently does not exist on Venus (O’Neill et al., 2013). As a result we have not included an asthenosphere explicitly to allow a generalisation of our results to Venus-type planets, but the effect of this on planetary evolution should be further explored. And, of course, there are uncertainties in the temperature-dependence of mantle viscosity, and the brittle parameterization used here (including factors such as strain-rate and damage dependent rheologies), that need to be considered further. Lastly, it has been demonstrated that perturbations in either surface conditions and/or rheology during a planet’s evolutionary path can significantly affect the outcome (Lenardic et al., 2008). This may come about due to extreme climate evolution (e.g. Venus), impacts, or volatile loss (Sandu et al., 2011), and require a sophisticated approach to parameterizing the coupling between the solid and surface systems.

The latter effects, noted above, are strongly coupled to a planet’s volcanic history. This has been explored by Moore and Webb (2013), and Nakagawa and Tackley (2012, 2015). Moore and Webb (2013) suggested a ‘heat-pipe’ regime – essentially a hot stagnant-lid regime with voluminous (non-tectonic) volcanic resurfacing, may have been applicable to the Hadean. They demonstrated melt extraction can be an efficient mechanism in transporting heat. However, their current formulation ignored depletion of the mantle, leading to an overestimation of efficiency. As internal temperatures waned, Moore and Webb (2013) found the heat pipe regime naturally evolved into a tectonically active regime, similar to the progression shown in our models. Similarly, Nakagawa and Tackley (2012, 2015) found magmatic heat transport to be a significant effect, and generated thick (~100 km) basaltic piles during stagnant interludes. They did not consider intra-crustal melting of the basalt crust, however, which would lead to buoyant, continental-type material, nor did they consider the rheological effects of mantle depletion. The primary difference between this work and melt-transport formulations is that, essentially, heat-pipe volcanism modulates the dynamics of a stagnant lid. Much of the heat loss due to volcanism in stagnant interludes is, in our models, simply stored internally, then lost during the episodic overturn events.

Ultimately, this style of model should be run in a spherical geometry incorporating the above complexities, and also coupled with a parameterized core evolution model (e.g. Butler and Peltier, 2002; Butler et al., 2005), where the basal mantle temperatures evolves with core temperatures, which themselves depend on the core–mantle boundary heat flux through time. For
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We have demonstrated in this work that variations in the magnitude, and ratio, of basal to internal heating, can fundamentally affect the tectonic regime of a planet. We have also demonstrated that the initial conditions of a simulation can play a critical role, particularly near regime boundaries, due to inherent hysteresis in such systems (Crowley and O’Connell, 2012; Wellier and Lenardic, 2012). The evolutionary models we present here demonstrate the plausibility of performing such simulations for time-varying mantle heat production, and CMB temperatures. Critical to the evolutionary path of such models, however, is the initial state of the mantle. For initial conditions near equilibrium with internal heat generation rates, the simulations possessed plate tectonics for their entire evolution, slowly waning into a sesquient stagnant-lid phase after 10–15 Gyr of evolution. However, if the formation of the Earth imparted significant thermal energy, as some models have suggested (e.g. Stevenson (1990)), then the simulations suggest Earth may have in fact begun in a hot stagnant-lid state, similar to Io. These evolutionary simulations then transit into an episodic regime for 1–3 Gyr. As they lose much of this primordial heat, and heat generation decays, the simulations transit into a plate-tectonic regime, before eventually decaying into a cold stagnant-lid regime after >10 Gyr of evolution. Thus a planet may begin, and end, its evolution in a stagnant state, and episodic plate tectonics may be the norm for planets beginning their geologic evolution from an initially hot thermal state. Consequently, plate tectonics may in fact be a tectonic phase evolving planets pass through, and there appears to exist a discrete window of opportunity for plate tectonic activity in the evolutionary arc of a planet.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.pepi.2016.04.002.

References


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